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PALAEO-ICE SHEET DYNAMICS AND DEPOSITIONAL SETTINGS OF THE LATE DEVENSIAN ICE SHEET IN SOUTH-WEST SCOTLAND

by

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Abstract

Recently recognised temporal changes in the configuration of former mid-latitude ice sheets during their long history has introduced doubt over the validity of traditional steady-state equilibrium models delineating Late Devensian glacial and deglacial events in south-west Scotland. This has prompted a reappraisal of the palaeo-ice sheet dynamics and mode of deglaciation in this region. In a wider context, the role played by the Southern Uplands ice sheet in Irish Sea glaciation is considered. Primary emphasis has been placed on addressing the sedimentological and glaciological effects of isostatic submergence, and the character of retreat of the ice sheet onto a part of the Scottish mainland which has remained substantially unconsidered.

A strategy for applying Landsat imagery to reconstruct flow geometries has been adopted which has allowed the palaeoglaciological structure of the south-west sector of the Scottish ice sheet to be reconstructed by reference to subglacial bedform alignment and cross-cut and overprinted bedform patterns. Ten major transient ice-flow stages have been identified during the last (Late Devensian) glaciation which demonstrate the style of ice stream development and reflect the time-dependent, dynamic evolution of the ice sheet. These findings provide a significant contrast to previous reconstructions. Zones of extensive pre-lineation deformation are thought to reflect former areas of sustained ice streaming which should be incorporated into future reconstructions of ice sheet elevation and internal flow kinetics, currently based on rigid-bed analogues.

During deglaciation, the major valleys of south-west Scotland maintained topographically constrained ice, which underwent a single major readvance during late-stage retreat (Southern Upland readvance). Within the Loch Ryan Basin this readvance was synchronous with reactivated ice flow of Highland origin (Highland readvance) which advanced beyond a previously inferred limit. A subsequent palaeosurge event (Loch Ryan palaeosurge), following the Highland readvance, is strongly suggested by macrofabric data within a far-travelled glacitectonite and the landform assemblage within the Loch Ryan Basin.

Available field evidence casts doubt on whether the ice margin retreat onto the south-west Scotland mainland was under marine-influence. A minimum deglacial age of 6 075 ± 55 years B.P. implies a relative sea-level height of at least 6 m above the present level during the Mid Holocene. This implies that the recession style of Eyles and McCabe (1989a), invoking an ice-contact glacimarine setting, cannot be verified.

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Acknowledgements

Physical geography is a discipline which fascinated me at school principally because I was able to relate many aspects of the taught syllabus to my childhood experiences, gained through hill-walking and other outdoor pursuits. As my breadth of knowledge of geomorphological landforms and processes developed, through being taught by Miss Stenhouse (Queen Anne High School, Dunfermline), the outdoor environment gradually started to make more sense to me and provided an added interest when climbing up Scotland's mountains. In retrospect, the initial appeal of the subject might not have existed if it had not been for my family's keen interest in natural history. I can still recall my earliest days when my maternal grandparents took me regularly around 'the wood' introducing me to all that the landscape had to offer, while my paternal grandparents complimented this by providing many excursions to coastal settings near Southport, which were equally educational. To round off my early introduction to the physical landscape, I was fortunate enough to go on annual summer holidays with my parents to some of Great Britain's finest spots. It made me realise that there are so many magnificent and varying landscapes in my home country- why go abroad dad! The enthusiasm of my family in providing me with a rich experience of science and the natural world is undoubtedly responsible for fuelling my eventual passion to produce a creditable piece of scientific work. My parents unfailing support and encouragement, not to mention their words of wisdom, have allowed me to finally fulfil my ambition. It is to them that I dedicate this doctoral thesis in acknowledgement of my resounding gratitude for all that they have done for me.

During my undergraduate studies at the University of St Andrews (1992-1996) I furthered my education in geography, strengthening the physical side of the discipline with various subjects in geology, and decided upon an honours degree in geoscience. I was fortunate enough to be taught by Prof. Colin Ballantyne, Dr. Charles Warren and Dr. John Walden. They presented the glacial and periglacial geomorphology option course in such an animated manner that it was impossible not to become fascinated by this immensely interesting subject. Consequently, the focus of my honours dissertation project was based upon some of the themes that they had taught, and allowed me to experience for the first time the enormous satisfaction which I would get from carrying out the research process. My first published research paper, arising from this undergraduate investigation, would not have materialised if it were not for Prof. Colin Ballantyne, my dissertation supervisor. Prof. Ballantyne persisted in making sure that I conduct my research in a logical and professional manner, and I look back now with much appreciation as this training has undoubtedly proved its worthiness in allowing me to produce this Ph.D. thesis.

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As well as those who were directly involved with my Ph.D. research, I would also like to extend my thanks to the other academic, technical and secretarial staff in the Department of Geography and Topographic Science who provided their support. Thanks also go to past and present postgraduate students who have helped me in one way or another, in particular my various office colleagues: Sally Gemmell, Mhairi Harvey, Lydia Pierce, David Shepherd and Darrel Swift. I am also most grateful to Peter Chung, Darrel Swift and Sarah Webber for assisting in the field.

During the early stages of my research, I spent a number of weeks at the Department of Geography, University of Sheffield under the supervision of Dr. Chris Clark. Not only did Chris make available the use of Landsat imagery, but took the time to introduce to me a new technique in geomorphological research which would subsequently form a fundamental part of this research. I am therefore eternally grateful to Chris for his patience and commitment. Also, I would like to express my thanks to Dr. Charlotte Bryant of the Natural Environmental Research Council Radiocarbon Laboratory, East Kilbride for preparing the mollusc samples and organising for them to be dated, and to Scottish Natural Heritage (Ayr and Newton Stewart offices) for allowing me access to their aerial photograph collection.

Throughout my time in Glasgow I have established a number of close friends, particularly Richard Young, Keith Malley, Sandra Ryan, Kenny McDonald, Ulrich Oslender, Derek McAllister and Pauline Ryan who I have known for almost the entire period that I have been conducting this research. It is to them that I owe much appreciation for their support and friendship which has seen me through to completion of this thesis. Likewise, I have been privileged to live at various times with some great flatmates who deserve a special mention, namely Richard Young, Judith Cudden, David George, Graeme Cook and Darrel Swift.

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Declaration

This thesis embodies the results of original research carried out by the author between October 1996 and April 2001. References to existing works are made as appropriate. Any remaining errors or omissions are the responsibility of the author.

Keith E. Salt

6th April, 2001.

CHAPTER ONE

Introduction

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Introduction

1.1: Prelude

In northern Europe, the recurrence of glacigenic sediments in the geological record stands as testimony to extreme climatic conditions throughout much of the last two million years, during which has evolved the complex landscape that we witness today. Major deterioration in climate is also represented in strata from the Early Proterozoic, Early and Late Palaeozoic and Cenozoic, with less profound cold periods associated with other geological intervals (Frakes, 1979; Frakes *et al.*, 1992). However despite this, it is the widespread extent of glacigenic evidence (morphological and sedimentological) relating to the Quaternary which has prompted the greatest interest in glacial geomorphological research.

In the British Isles, though, there is still at present a limited consensus on the timing and pattern of environmental change which resulted in the onset of glaciation (Ballantyne and Harris, 1994). Only the most extensive Anglian (480 ka-428 ka B.P. (Rose, 1989; Ehlers *et al.*, 1991)) and the most recent Devensian (122 ka-10 ka B.P. (Rose, 1989; Ehlers *et al.*, 1991)) glaciations are suitably established in terms of type sections or extent, with knowledge of the latter rather more detailed than for its predecessors (Ballantyne and Harris, 1994; Figure 1.1). The Devensian stage can be formally split into three substages, the youngest being the Late Devensian substage which independently has been subdivided on the basis of climatostratigraphic criteria into three chronozones: the Dimlington Stadial (*c*. 26-13 ka B.P.), the Windermere Interstadial (*c*. 13-11 ka B.P.) and the Loch Lomond Stadial (*c*. 11-10 ka B.P.) (Rose, 1985, 1989). It is during the cold period defining the Dimlington Stadial that the growth of the last British ice sheet was initiated, followed by its subsequent decay during abrupt thermal amelioration (Bard *et al.*, 1987). At its maximum, the extent of the ice sheet was such that it covered approximately two-thirds of the present land area of the British Isles (Figure 1.1), with major ice dispersal centres in Scotland located in the mountain massifs of the Northern and West Grampian Highlands and the Southern Uplands of south-west Scotland (e.g. Jolly, 1868; Geikie, 1901).

Quaternary research in south-west Scotland has never attracted the volume or detail of work that has been entertained in other regions of Scotland. Indeed, on a number of occasions the author has heard reference made to south-west Scotland as one of the 'black-holes' in Scottish glacial geology. This is despite the strikingly distinctive and characteristic nature of the landscape which is owed largely to the widespread processes connected with Late Devensian

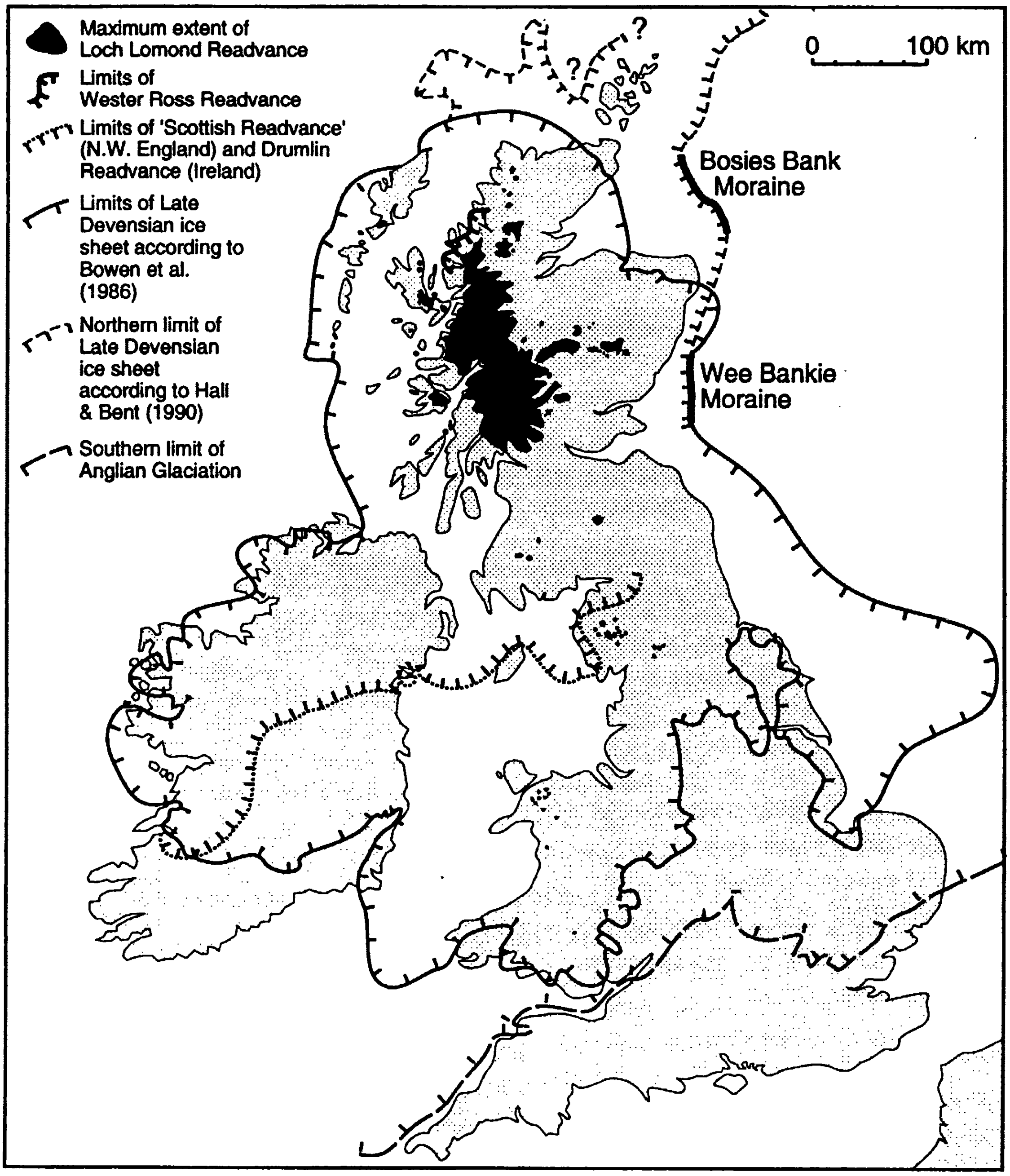


Figure 1.1 Extent of glacial limits in the British Isles. (From Ballantyne and Harris, 1994).

glaciation. Reflection upon the evident fresh stamp of glaciation, in particular on the Galloway region, has even prompted Eckford (1957) to humorously suggest that the area be designated as a Glacial Park. Despite the scarcity of work conducted in south-west Scotland there does exist a long history of Quaternary research, detailed in Section 1.4, which has in the last century attracted eminent authorities such as James and Archibald Geikie (A. Geikie, 1901; J. Geikie, 1905), and J. Kaye Charlesworth (1926a, b). Certainly, a number of localised detailed studies prompted for a time a reawakening of interest in glacial events in the region (e.g. Jardine, 1975, 1980; Cutler, 1979, Kerr, 1982a), however there has been no regional synthesis* which would place south-west Scotland in the context of Quaternary events in Northern Britain.

In agreement with the sentiments of Kerr (1982b), the area has in the past been largely considered as an 'appendage' to the research of other regions such as Cumbria in north-west England or Ulster in north-east Ireland. This is unfortunate as the area provides a key position in relation to Late Devensian glaciation in the Scottish Highlands, north-west England, north-east Ireland, north Wales and the Isle of Man. Examination by Kerr (1982a) of relevant work from these areas bordering the Irish Sea Basin reveals a considerable difference in interpretation of glacial events in the Mid to Late Devensian. Thus it is increasingly apparent that the central location of south-west Scotland, and the establishment of a glacial chronology, is of fundamental importance in the context of correlating these somewhat disparate adjacent areas.

A further pertinent research interest in this area relates to the reconciliation of the controversial status of readvances, that purportedly interrupted overall retreat of the Late Devensian ice sheet. In particular, persistent reference occurs in the literature to a 'Scottish Readvance' (e.g. Trotter, 1929; Huddart, 1970, 1971a, b, 1991, 1994; Figure 1.1), whose presence in the chronology of glacial events in south-west Scotland appears based on evidence which is perhaps suggestive, but under no circumstances unequivocal. Certainly, far more reliable evidence does exist beyond this region, such as in Cumbria (Trotter, 1924, 1929), suggesting a main glaciation followed by one or more readvances, and perhaps more so on the Isle of Man (Charlesworth, 1939) and in north-east Ireland (Charlesworth, 1939). Yet with regard to this model suspicions are understandably raised in what may be perceived as an ambitious desire to correlate events on the eastern seaboard of the Irish Sea with the reported 'Drumlin Readvance' in north-east Ireland (Synge, 1968, 1969; Figure 1.1). This scheme is perhaps naturally justified if emphasis is placed upon the distinctive, and somewhat comparative topography.

* This is with the exception of a review by Sissons (1974) which is largely concerned with northern and eastern Scotland.

Nevertheless, response to this interpretation is the continuous reiteration that south-west Scotland, at present, has no equivalent in its chronology despite similarities in mode and form of glacial deposition, and therefore does not accommodate a readvance.

The glaciological debate which exists in south-west Scotland is therefore mainly focused on ascertaining the nature of events during the period of Late Devensian glaciation in the north-east sector of the Irish Sea Basin, and thus the primary emphasis of this research is centred on addressing and unravelling the long-standing issues outlined above. Paralleled with this lies an opportunity to investigate the sedimentological and glaciological effects of isostatic submergence, and the retreat of the ice sheet onto a part of the Scottish mainland which remains substantially unconsidered in the context of Irish Sea glaciation.

The remainder of this chapter is divided into four sections. Section 1.2 further justifies and develops the aims of this study in greater detail. This is followed by an overview of the regional physiography of south-west Scotland (Section 1.3). Section 1.4 reviews the history of Quaternary studies in the region and establishes the present status of the glacial history. Finally, Section 1.5 provides a summary of the main points raised in this chapter.

1.2: Rationale and Research Objectives

1.2.1: Rationale

Historically, the popularity of glacial geomorphology can largely be attributed to the development of landform and process studies in those regions of the Northern hemisphere which have been dramatically affected by the results of glacier activity. In areas where glacial deposits are widespread, geologists have tended to become preoccupied with glacial stratigraphy and glacial chronology, whereas elsewhere the main focus of research has been glacial geomorphology, with particular emphasis on landscape evolution. Nevertheless, since the general acceptance by scientists of the 'Glacial Theory', which evolved in the last decades of the eighteenth century and early in the nineteenth century, an overriding theme has been universally accepted that sediments and landform assemblages of glacial origin are significantly important palaeoenvironmental indicators.

Detailed investigations into spatial patterns of glacial sediments and landforms on ancient landscapes, and their significance at local, regional and even continental scales, have a long-standing history (e.g. Geikie, 1877, 1894; Upham, 1895; Tyrell, 1898a, b; Flint, 1943). Today, the working goal in Late Devensian palaeoenvironmental reconstruction remains unaltered which is to establish the precise nature of glacial deposits and the sequence of inferred environmental events at a given locality, and then to classify the result and correlate the local evidence with other areas (Bowen, 1978). Despite the remarkable complexity of many of the early studies, in certain regions of the British Isles the accuracy of these initial accounts are questionable, given that in recent years several major deficiencies in the applied traditional approaches have been identified (e.g. Ireland (McCabe, 1987)).

During the second half of the nineteenth and early twentieth century individuals such as Jolly (1868), Geikie (1901), Gregory (1926) and Charlesworth (1926a, b) provided substantive explanations for the widespread occurrence of glacially-related deposits and the genesis of landforms in south-west Scotland. During that period, the outcome of their individual work was seen to be of special importance in the history of Quaternary research in Scotland, and was proclaimed as the "fruit of diligent field research" (Kerr, 1982b, p. 3). However given the current level of scepticism over the validity of specific traditional interpretations, there is concern that the existing descriptions of glacigenic sequences, and contrasting accounts of Late Devensian glacial events in south-west Scotland, are encumbered with genetic inferences making up-to-date stratigraphic correlation and synoptic studies of this region difficult in the wider context of the Irish Sea Basin.

Alongside this, it is increasingly apparent that former palaeoenvironmental interpretations of complex glacigenic sequences, especially on the seaboard margins of the Irish Sea, have failed to take account of influences that a marine ice sheet might have had on the sedimentary record. In the past, most sequences were interpreted in terms of glacio-terrestrial models, however, Eyles and McCabe (1989a) have recently presented evidence to suggest rapid Late Devensian ice sheet retreat and marine flooding in the Irish Sea basin.

Overall, a number of reasons, the most prominent ones of which have been highlighted, have prompted leading workers to consider reviewing key areas. This is particularly necessary in regions such as south-west Scotland where there exist obvious contrasts in the interpretation of the nature of glacial and deglacial events (Kerr, 1982b), and in instances a lack of clarification of episodes such as ice sheet retreat onto the mainland. Existing doubts over the inferred glacial stratigraphy and history of Late Devensian events provided by early workers in these regions appear to relate directly to at least the following factors: (1) limited and inadequate field procedures; (2) over-emphasis on the then existing exposures; (3) genetic terminology; and (4) the choice of conceptual model used to describe and assess the glacigenic deposit (Bowen *et al.*, 1986; McCabe, 1987).

In recent years, many regional studies have identified and placed greater emphasis upon the links between sediment-landform association and genetic processes than conventional lithostratigraphic practices (e.g. Boulton and Paul, 1976; Boulton and Eyles, 1979; Eyles, 1983a, b, c; Eyles and Eyles, 1983; Evans, 1988). For the purpose of classification, this may be described as an holistic approach, wherein the landforms and subsurface sediments that characterise a landscape are genetically related to the process involved in their development. Thus, it is seen to provide a far stronger and more reliable framework for comparison of field evidence and development of stratigraphic concepts during reconstruction of former glaciated environments. Two new and powerful approaches to the study of glacial depositional systems have emerged from this methodology: (a) the study of glacial landsystems and process-form models; and (b) sequence stratigraphy, both of which are described in detail in Section 3.3.1.

Aside from the importance of re-evaluating the Quaternary history of this key area in Scotland, it is believed that the opportunity to place a handle on the palaeoglaciology of the Southern Uplands ice sheet through its glacial cycle, in terms of the major elements which make up the integrated, large-scale structure of such an ice sheet (e.g. ice divides, ice streams), has major significance to glaciology beyond this regional-scale investigation. This is particularly with respect not only to constraining the overall dynamics and structure of this sector of the British ice sheet which impacts on previous British ice sheet models (e.g. Boulton *et al.*, 1977) and glacial rebound models (e.g. Lambeck, 1993), but also addresses certain fundamentally important glaciological issues such as whether ice mass fluctuations and instabilities, if detected, are controlled by climatic variability, or whether they are intrinsically related to components which determine the internal dynamics of ice flow (e.g. a subglacial deformable bed).

1.2.2: Objectives

Overall, there are at least four considerations which make this research imperative:

(1) a methodological framework has been developed allowing the re-evaluation of contrasting accounts of Late Devensian glacial events in south-west Scotland;

(2) evidence exists to suggest that possible marine-influenced ice sheet retreat occurred in the Irish Sea Basin which has been unaccounted for in south-west Scotland;

(3) the status of readvances during the late stages of deglaciation in south-west Scotland is yet to be resolved;

(4) the role played by south-west Scotland in the context of Irish Sea glaciation and Quaternary events in Northern Britain is undetermined despite the key positioning of this region.

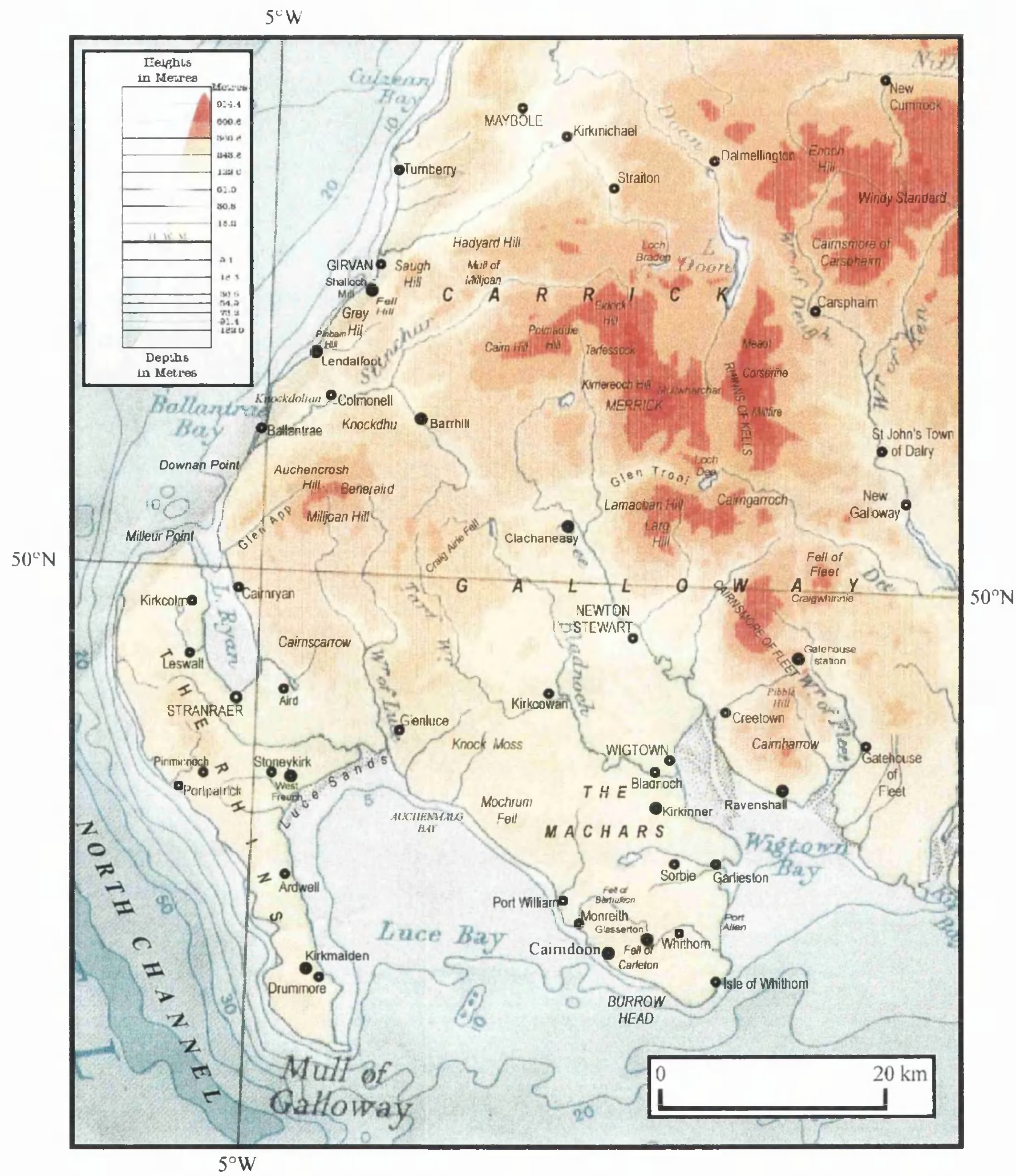


Figure 1.2 Topography and main physical features of south-west Scotland. Submarine contours given in fathoms. (Adapted from the 1: 1 000 000 O.S. map of XVII Century England, 1930).

It is acknowledged that the potential to encounter similar equivocal evidence as earlier workers did may present problems with interpretation, however, the importance of addressing the outstanding issues in this region provides the basis for pursuing this investigation.

1.3: Regional Physiography of South-West Scotland

1.3.1: Physical setting of south-west Scotland

South-west Scotland is defined here, for the purpose of description, as the region covered by Sheets 1 and 3 (Kirkmaiden and Stranraer), Sheet 2 (Whithorn), Sheets 4W and 4E (Kirkcowan and Wigtown), Sheet 7 (Girvan) and Sheets 8W and 8E (Carrick and Loch Doon) of the 1:50,000 British Geological Survey map series (Figures 1.2, 1.3). The north and east limits can be demarcated by a line drawn from Girvan, on the west coast, to the summit of Cairnsmore of Carsphairn in the east (55°14'N), and directly south through Dundrennan to the Solway Firth coastline (4°13'W). The western and southern limits are marked by the coastline of the North Channel and Solway Firth respectively. Throughout the text, reference may also be made to the northern half of this region as the Carrick area and the southern half as the Galloway area.

The following sections contain a general overview of the landscape characteristics in each of the five districts identified above.

1.3.1.1: Rhins of Galloway (Kirkmaiden and Stranraer)

This district marks the termination of the Southern Uplands mountain range against the North channel and comprises the distinctive Rhins of Galloway (rinn = a point or promontory in Gaelic), a north-south peninsula known locally as 'The Rhins' which is joined to the main region of south-west Scotland by the low relief Stranraer isthmus (Stone, 1995; Figure 1.2 (also reproduced as a separate map in the back wallet)). The Rhins therefore form the most south-westerly corner of the Scottish mainland with the Mull of Galloway marking the most southerly point. From here, a moderately high ridge of land underlain by Lower Palaeozoic strata trends north-north-west for 45 km to Milleur Point at the north tip, thus separating the Stranraer Permian basin of Luce Bay and Loch Ryan from that of the North Channel (Kerr, 1982a; Stone, 1995). In turn, the 10 km wide isthmus separates Luce Bay in the south from Loch Ryan in the north.

On the western coast of the Rhins, steep bedrock cliffs are prevalent (up to 70 m high) whereas elsewhere more open bays exist where the coastline intersects low-lying valleys cutting across the Rhins. It is suggested that the valleys represent preferential erosion by ice of structural weaknesses in the bedrock (Jardine, 1966). The actual 'spine' of the peninsula is, however, formed by higher ground with hills to the south of Kirkmaiden reaching 164 m above OD, and the central part, between Stranraer in the east and Portpatrick in the west, rising to 182 m above OD. The east coast of the Rhins, on the western coast of Loch Ryan and Luce Bay, is low-lying and topographically subdued where it merges with the Stranraer isthmus. The relief in this area is never greater than 40 m above OD, and generally lies below 15 m, particularly in the south-east part of the isthmus where an undulating topography gives way to extensive beaches and intertidal mudflats at the head of Luce Bay.

In contrast, the topography to the north-east of the isthmus changes abruptly coinciding with the Loch Ryan Fault; the eastern boundary structure of the Stranraer basin (Stone, 1995). A marked north-west to south-east trending fault scarp has been produced by the more resistant Lower Palaeozoic strata, beyond which the land surface rises to a peneplain at approximately 230 m above OD (Stone, 1995). This surface is incised by a number of streams draining either south-west into Loch Ryan or south-south-east into the Water of Luce which enters the Solway Firth at the head of Luce Bay.

1.3.1.2: Whithorn

The Whithorn district (Figure 1.2) lies towards the southern end of the Wigtown peninsula (Section 1.3.1.3) which itself extends southwards into the Solway Firth, terminating in the high bedrock cliffs of Burrow Head. The peninsula acts as a physical barrier separating Luce Bay to the west from Wigtown Bay to the east. The topography of the area is generally subdued however an undulating relief predominates inland as a result of a strong topographic lineament reflecting both the regional strike of the underlying strata and accentuation of the extensive Lower Palaeozoic sedimentary bedrock by inferred glacial erosion (Barnes, 1989). The highest elevations in the Whithorn district are the Fell of Carleton (146 m) and the Fell of Barhullion (136 m) occurring adjacent to the south-west coastline near Monreith, while elsewhere the land surface remains below 90 m above OD. Drainage in the area is restricted to a limited number of streams with the more prominent ones entering either into Luce Bay at Monreith or Wigtown Bay at the Isle of Whithorn.

1.3.1.3: Wigtownshire (Kirkcowan and Wigtown)

Nearly two-thirds of the area covered by the Wigtownshire district (Figure 1.2) consist of hilly ground and moorlands with irregular and independent ridges. The highest ground in the district lies in the north-east corner, to the east of the fault-controlled Cree valley (Floyd, 1999), whose river (River Cree) flows between the base of Cairnsmore of Fleet and the Rhinns of Kells on the one side, and the lower and flatter moors of Wigtownshire on the other. The River Cree also marks the eastern boundary of Wigtownshire from neighbouring Kirkcudbrightshire. In the centre of the district rises the granite mass of Cairnsmore of Fleet, which reaches a height of 771 m above OD (Floyd, 1999). Surrounding it are further prominent mountains, amongst which are the Fell of Fleet (470 m) and Craigwhinnie (417 m) to the east, and Pibble Hill, Cambret Hill (351 m) and Cairnharrow (456 m) to the south.

To the west of the River Cree the hills are far apart from one another and do not reach an equivalent height, amongst the highest being Craig Airie Fell (320 m), White Fell (287 m) and Craigmoddie Fell (248 m) in the north-west; Culvennan Fell (213 m), Barskeoch (176 m) and Craigeach Fell (131 m) in the centre; and Mochrum Fell (197 m) and Bennan Hill (152 m) in the south-west. There is no stretch of flat ground of any great extent in Wigtownshire, with the exception of a small area of just a few square miles around the mouth of the River Cree. The area of Wigtownshire south of Kirkcowan, which is known locally as 'The Machars', consists of a drumlinised topography, with streamlined mounds running predominantly over most of the area in a north-east to south-west orientation (Section 1.3.3).

The coastline along the shores of Wigtown bay display periodic steep cliffs extending up to about 30 m above OD in places whereas the remainder of the coastline consists of banks of drift, rising from flat shingle beaches. The main drainage patterns in the area are those of the Rivers Cree, Tarf and Bladenoch which all enter the Solway Firth at the head of Wigtown Bay. Besides these, several large streams such as Palnure Burn and Moneypool Burn drain other independent areas.

1.3.1.4: South-west Ayrshire (Girvan)

The general physical character of this area (Figure 1.2) presents little variety compared to that of Wigtownshire. The area is hilly throughout, with the only comparatively level ground being found alongside the main streams. Despite this regional view, if the line of the River Stinchar is taken, it is apparent that more elevated ground prevails to the south, whereas the landscape to the north of the river is more diversified. The land to the south forms a northward extension of the peneplain surface described in Section 1.3.1.1. and near the northern margin the surface begins to ascend gently to a ridge of hills trending north-east to south-west; the highest point being Beneraird (439 m) which overlooks the long linear valley of Glen App whose river (Water of App) flows south-west into the North Channel. North of the River Stinchar, the ground does not attain a height greater than 300 m (Grey Hill, 297 m). The dominant valleys, or scarp features, of the Stinchar and Glen App coincide with the incision of the north-east trending Caledonoid Southern Upland fault and provide an obvious 'grain' to the landscape

(Stone,1995; Floyd, 1999). The peculiar straightness of these two valleys, in particular the latter, serve to differentiate them from other longitudinal valleys such as that of the Water of Girvan, which is not fault-controlled (Stone, 1995).

The coastline from the northern edge of the district, southwards to where the River Stinchar enters the North Channel, presents a narrow flat margin of ground between the current coastline and a steep rise which marks the former coastline, where it ascends inland. In a number of places, this platform is interrupted by the projection of residual headlands entering into the sea. South of the Stinchar, a line of cliffs descend abruptly into the sea all the way south into Loch Ryan.

1.3.1.5: Carrick and Loch Doon

This area has a varied topography which ranges from 8 m above OD in the north-west corner, where the coastline is just 1 km away, to the dominant peak of Merrick which, at 843 m, is the highest peak in Scotland south of the Highlands (Figure 1.2). The western half of the Carrick-Loch Doon district has a fairly subdued topography which rises gently southwards and eastwards before gradually falling towards the valley of the River Cree which flows south into the Wigtownshire district. In the north-west, there is a continuation of the scarp valleys described in Section 1.3.1.4 (Floyd, 1999). By contrast, the eastern part of the district is characterised by mountainous topography, including the rugged peaks of Craignaw (645 m), Mulwharchar (692 m) and Craigmawhannal (357 m). These peaks are in turn flanked by the twin parallel ridges of the Merrick and Kells ranges.

The district contains a number of lochs of various sizes, the vast majority of which are natural bodies of standing water, although at Loch Doon (largest in the Southern Uplands) the natural water-level has been raised by dams constructed in the 1930s (Floyd, 1999). A number of artificial reservoirs have also been built including Clatteringshaws Loch, and Loch Braden to the west of Loch Doon has been created by raising the water-level of several smaller lochs by dams (Floyd, 1999). Other notable natural water bodies in the area include lochs Trool, Riecawr, Macaterick, Enoch, Neldricken, Valley and Dee. This area also features as the watershed between rivers flowing northwards into the Firth of Clyde, westwards into the North Channel, and southwards into the Solway Firth (Floyd, 1999).

1.3.2: Pre-Quaternary geological history of south-west Scotland

Geologically, the Carrick and Galloway districts are part of an extensive outcrop of Lower Palaeozoic strata which forms the Southern Uplands of Scotland, an area bounded to the north

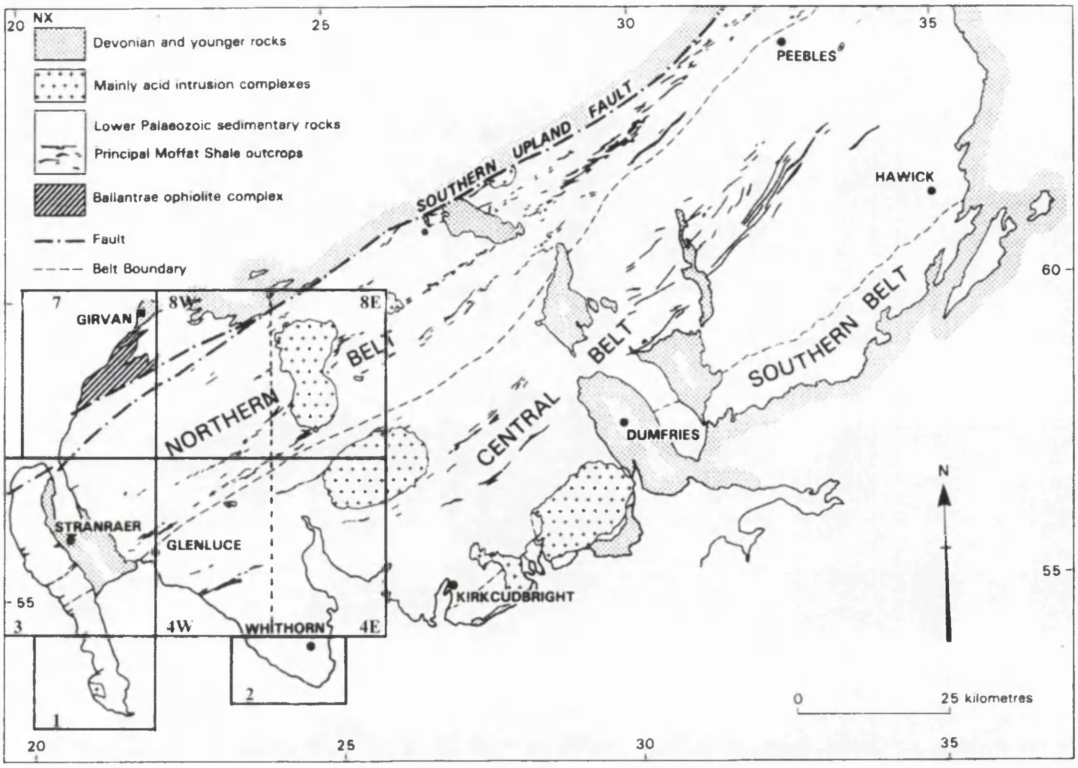


Figure 1.3 Sheets 1 and 3 (Kirkmaiden and Stranraer), Sheet 2 (Whithorn), Sheets 4W and 4E (Kirkcowan and Wigtown), Sheet 7 (Girvan) and Sheets 8W and 8E (Carrick and Loch Doon) of the B.G.S. 1:50 000 series in their regional geological context within the Southern Uplands, south-west Scotland. (Modified from Barnes, 1989).

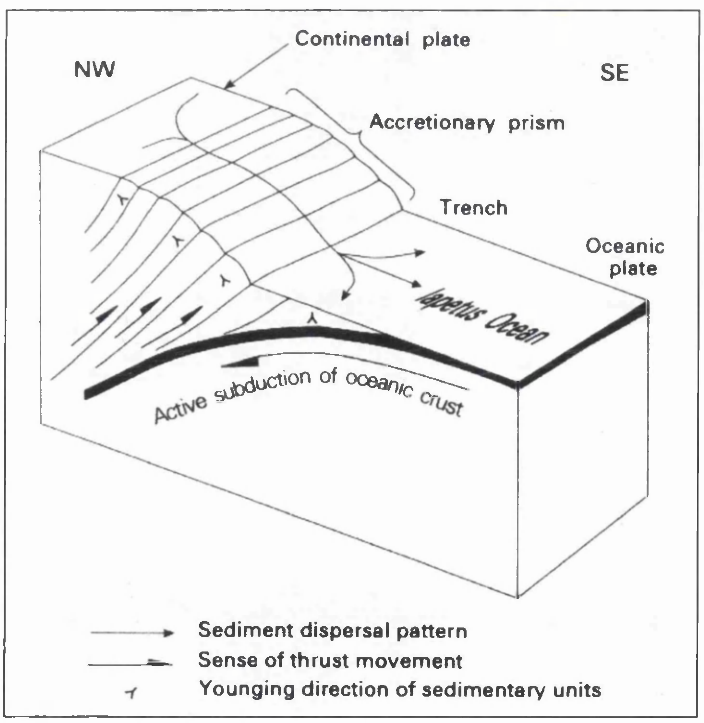


Figure 1.4 The accretionary prism model developed by Leggett *et al* (1979) for the formation of the Southern Uplands thrust belt, south-west Scotland. (From Stone, 1995).

by the north-east to south-west trending Southern Upland Fault and to the south by unconformably overlying Devonian and Carboniferous strata (Greig, 1971; Barnes, 1989; Figure 1.3). Early work conducted by Peach and Horne (1899) on the Southern Uplands concluded that the basic stratigraphy of the south-west Scotland region was formed from three parallel outcrops, one of Ordovician and two of Silurian (Llandovery and Wenlock) strata. The respective 'Northern', 'Central' and 'Southern' belts, as they were distinguished, are each made up predominantly of unfossiliferous greywacke, though with traces of interbedded fossiliferous siltstone beds in the Southern Belt, and fossiliferous black mudstone and chert sequences ('Moffat Shale'), creating thin, discontinuous, strike-parallel outcrops in the Northern and Central Belts (Barnes, 1989; Figure 1.3).

These outcrops attest to processes active at the margin of a precursor to the Atlantic, the Iapetus Ocean, some 400 to 500 million years ago (Stone, 1996). During the Cambrian Period (*c.* 540-510 Ma), the Iapetus Ocean is thought to have probably expanded to its maximum size, with further evolution and eventual destruction by collision between the bordering continental plates (Laurentia and Avalonia) emplacing the outcrop of Lower Palaeozoic (*c.* 510-410 Ma) rocks of the Southern Uplands (Caledonian Orogeny) (Stone, 1996). Overall, there is a south-east younging trend in the area, with the oldest strata in the north-west of Carrick and progressively younger rocks appearing southwards into Galloway (Toghill, 1970). The oldest outcrops are an ophiolitic assemblage, in broad terms representing relics of the oceanic crust and mantle, which cover about 75 km^2 on the Ayrshire coast south of Girvan (Stone, 1996). As you progress further south, the Lower Palaeozoic rocks are divided by major strike-parallel faults into a series of blocks; a feature highlighted in the topographic expression of the area (Section 1.3.1.4). Each block is represented by only a limited stratigraphic succession, dominated by turbidite deposits of greywacke, siltstone and shale (Barnes, 1989; Stone, 1996).

The structural and stratigraphic pattern recognised in south-west Scotland is that of an imbricate thrust system (Webb, 1983) with the overall geological relationship within the Southern Uplands being explained by an 'accretionary prism' model (Leggett *et al.*, 1979; Figure 1.4). It is thought that the Southern Uplands complex developed as successive thin layers of sediment were scraped off from the surface of the descending oceanic plate and stacked at the continental margin in a series of slices. Rotation of the thrust stack to the vertical occurred as a result of suggested continental collision as the Iapetus Ocean closed in the Late Silurian or Early Devonian (Leggett *et al.*, 1979).

During the final stages of deformation, at the end of the Caledonian Orogeny, a widespread regional dyke swarm was intruded into many parts of the Carrick and Galloway regions (Stone, 1996). Dykes of Late Caledonian age are common features of the coastal sections throughout the Whithorn and Rhins of Galloway districts. Intrusions range up to several metres in thickness, although the majority are between about 10 cm and 1 m (Barnes, 1989; Stone, 1995). The climax of igneous activity was reached as deformation ended with the intrusion of major granitic plutons which dominate south-west Scotland (Stone, 1996). Three plutons of early Devonian age, comprising the main plutons of the south-west Scotland granitic suite, are present in the Carrick-Loch Doon and Wigtownshire areas (Figure 1.3). These are Cairnsmore of Carsphairn, Loch Doon and Cairnsmore of Fleet which have all been radiometrically dated to within a few million years of 400 Ma (Halliday *et al.*, 1980).

The Loch Doon pluton mass (408 ± 2 Ma; Halliday *et al.*, 1980) is one of the major plutons in the Southern Uplands with a north-south length of 19 km, an east-west breadth varying between 5 and 10 km, and an overall area of 125 km^2. The mass is centred on the peak of Mullwharchar, and stretches from Loch Doon in the north to Loch Dee in the south. A large number of minor intrusions occur as part of a dyke swarm associated with the emplacement of the Loch Doon pluton. To the east and west the pluton is flanked by the ridges of the Rhinns of Kells and the Merrick range respectively. The Cairnsmore of Fleet pluton (392 ± 2 Ma; Halliday *et al.*, 1980), occupying an area of 198 km^2, lies between Newton Stewart and Loch Ken and has its western portion dominated by the Cairnsmore of Fleet massif. The third pluton, Cairnsmore of Carsphairn (410 ± 4 Ma; Thirlwall, 1988), is by far the smallest of the plutons covering only 10 km^2 and is centred on the hill of the same name. The intrusion of these granite plutons was accompanied by contact metamorphism of the surrounding parent material (Stone, 1996).

Following the Caledonian Orogeny and intrusion of the granitic plutons in Carrick and Galloway, the Southern Upland region was subjected to uplift and erosion (Stone, 1996). Major sedimentary basins developed as a direct consequence of crustal extension during Late Devonian times (Figure 1.5). It is suggested that their formation was possibly activated through movement of north-west trending normal faults during Early Carboniferous times which extend throughout the Southern Uplands (Stone, 1996). The Loch Ryan basin, defined by these faults, preserves a fragmentary record of fluviatile, estuarine and marine sedimentation throughout the period from the Late Devonian to Late Carboniferous times (*c.* 360-290 Ma) (Jackson *et al.*, 1995; Stone, 1995, 1996). Along the north Solway coast, which effectively marks the northern margin of the Solway Basin, Early Carboniferous rocks are exposed (Figure 1.5). Away from

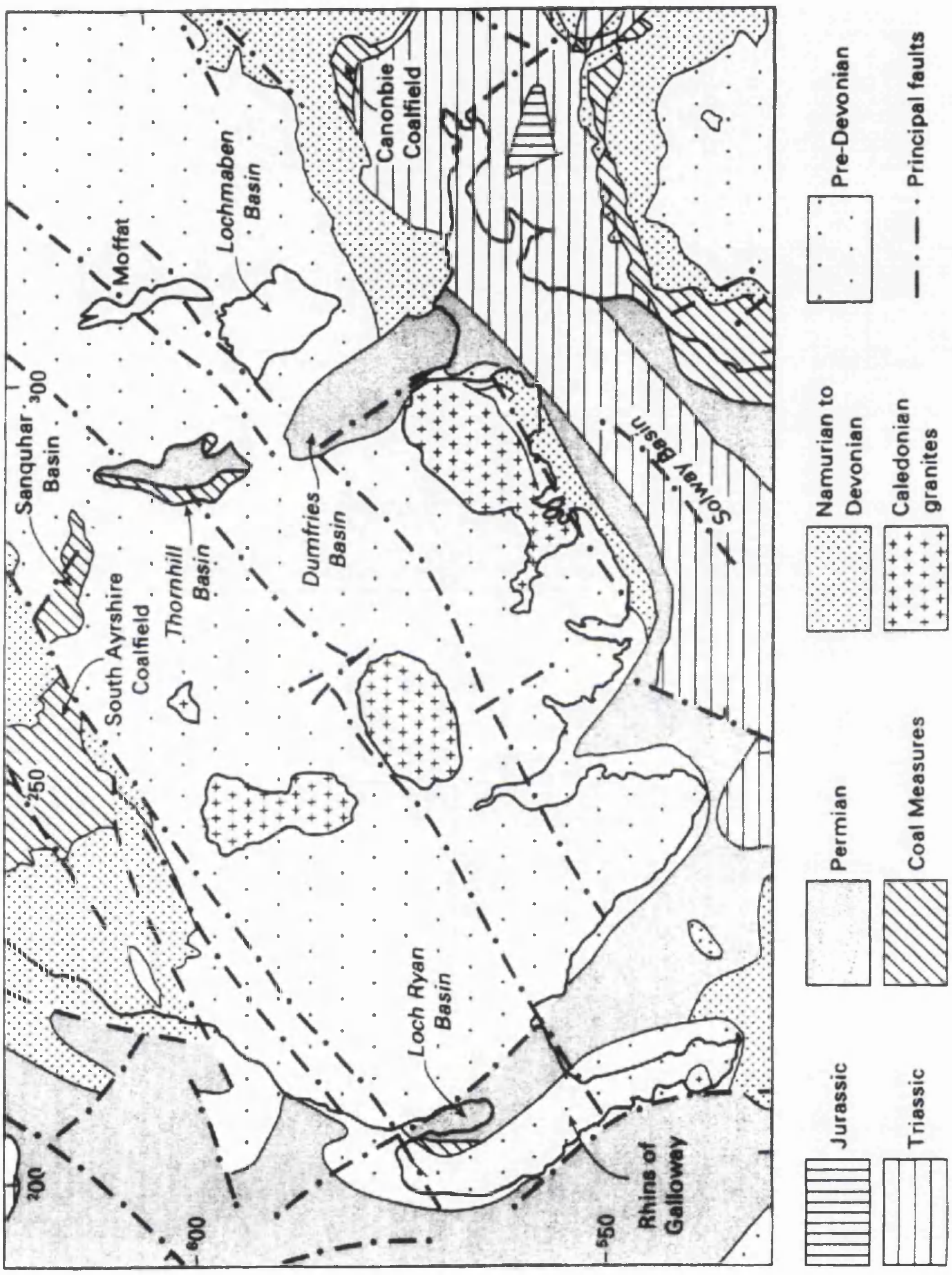


Figure 1.5 Principal faults and sedimentary basins in south-west Scotland. (From Stone, 1996).

this margin similar aged strata reveal evidence of a changing sea-level at this time (Stone, 1996).

There are no strata preserved to record the geological history of south-west Scotland between the Triassic and the Quaternary (Stone, 1996). This is with the exception of occasional Tertiary dykes, thought to be about 55-60 million years old, which belong to swarms associated with igneous centres in Arran and Mull (Harrison *et al.*, 1987; Hitchen and Ritchie, 1993).

1.3.3: Pleistocene and Holocene deposits of south-west Scotland

The information in this section is intended to provide a basic factual archive relating to the type and range of Quaternary deposit in the Carrick and Galloway regions of south-west Scotland. It is not the intention here to discuss any proposed chronology of events relating to the deposition of the sediments as this subject is addressed in Section 1.4. Drift deposits mapped in the districts of Whithorn, Rhins of Galloway and Carrick-Loch Doon are summarised from descriptions published in the accompanying revised British Geological Survey memoirs (Barnes, 1989[1] ; Stone, 1995[2] ; Floyd, 1999[3]). For the districts of south-west Ayrshire and Wigtownshire, information has been extracted from the initial surveys reported in the Memoirs of the Geological Survey of Scotland by Young (1869) and Irvine (1878) respectively. Classification and interpretation of Quaternary deposits in these districts by early glacial geologists (e.g. Gregory, 1926; Charlesworth, 1926a, b) are, for the most part, excluded in this section so as to avoid erroneous representations which may have arisen from preconceived notions.

It is necessary to reinforce the point introduced in Section 1.2 that the descriptions of the drift geology in these districts are largely based upon the initial surveys conducted over a century ago, before the vast development of knowledge of processes and sediments associated with ice movement and ice melting had taken place. Description of deposits taken at the time should be accurate although interpretation of landforms and inferred processes of formation, where made, may be subject to reinterpretation. Nevertheless, it is necessary at this initial stage to establish a

[1] Whithorn was first surveyed by Craik in 1871-72. The first edition of the map, published in 1872, was accompanied by a memoir published in 1873. The map was revised by Peach and Horne and a second edition was published in 1923. A partial resurvey of the district was made by Barnes in 1983-84. Revised solid and drift maps were published in 1987 and a revised memoir in 1989.

[2] The Rhins of Galloway was first surveyed by Irvine in 1872 (Sheet 1), and Geikie and Irvine in 1873 (Sheet 3) with maps and accompanying memoir produced at this time. The area was resurveyed by Peach and Horne between 1888 and 1898 and a second map was published in 1923. Revised drift and solid maps were published in 1982 and 1993 respectively, and a revised memoir in 1989 and 1995.

[3] This is the first Geological Survey publication specifically describing the Carrick-Loch Doon district.

benchmark for information relating to Quaternary deposits in the area which can be later supported or challenged where appropriate.

1.3.3.1: Rhins of Galloway (Kirkmaiden and Stranraer)

Widely distributed throughout the Rhins of Galloway has been found a till of varying thickness (up to 2-3 m) which tends to drape over irregularities in the underlying Lower Palaeozoic strata, but is far more pronounced where it has been moulded by the movement of ice into flow-parallel drumlin forms, thereby producing a distinctive landscape. There has been no evidence reported which suggests that glacial deposits older than the last glaciation are present in the area (Kerr, 1982a). The texture of the till has been found to be variable ranging from a stiff clay containing relatively few clasts to a sandy clay concentrated with subangular pebbles and cobbles. Similarly, the tills vary in colour and in composition, relating to the nature of the underlying bedrock, indicating that a large part of their material has not been carried very far prior to deposition (Kerr, 1982a).

Where the underlying strata are of Ordovician or Silurian age the matrix of the till is usually brownish grey and the clasts are mainly greywacke and hard shales; in areas where the clast content is of Permian age (Old Red Sandstone), the till matrix is red and sandy and encloses clasts of red sandstone (Kerr, 1982a). The latter type is thought to be directly associated with deposition by ice which previously eroded Permian red sandstone of the Stranraer basin to the east and north-east (Section 1.4). In several places the till has been divided into two superincumbent layers on the basis of differences in colour, consistency, particle size, erratic content, till fabric, presence or absence of carbonates and presence or absence of shells (Kerr, 1982a). To indicate their stratigraphic relationship, and to assist correlation with Irish tills, the tills have been provisionally named 'upper' and 'lower'. The junction between the tills is reportedly marked by a layer of sand which in places has been noted to penetrate into the lower till in wedges (Greig, 1971).

Documented characteristics of the upper till suggest that it is of a generally red or light brown colour, friable in texture and ranges from sandy loam to sandy clay loam with a much lower silt content than the lower till. The erratic material is much less than for the lower till with the majority of greywacke clasts being of local provenance (Loch Doon and Loch Ronald are suggested sources; Greig, 1971). The most consistent feature of the upper till is the record of a north-east to south-west fabric, which supports a traditional theory (Charlesworth, 1926a) of Late Devensian ice moving south-west across the Rhins of Galloway (Section 1.4).

The lower till is reported to be more patchy in its distribution than the upper till and its colour is darker; usually dark brown, purple or blue-grey (Kerr, 1982a). Overall, the lower till has been described as massive and compact ('stiff') with a low stone content and a matrix which is clay loam in texture. At two sites, marine shells have been recovered (Kerr, 1982a). The erratic content of the till is mostly sandstone, and the proportion of erratic to non-erratic material is high. Fabric analysis has revealed clear evidence of a north to south or north-west to south-east direction in contrast with the orientation in the upper till. The diagnostic character of the lower till appears to be its strong reaction to the addition of hydrochloric acid, in contrast to the upper till which fails to react to acid. In general terms the lower till has been classified as a lodgement till, pertaining to deposition and compression beneath an advancing ice sheet. By contrast, the more prevalent upper till is thought to have been associated with the decay of stagnant, *in situ* ice.

East of the Rhins peninsula, one of the most extensive accumulations of Quaternary deposits in the district forms the spread of the Stranraer isthmus. Between the outcrops of solid strata of the Rhins of Galloway and the steep rock scarp extending from the east of Luce Bay to Cairnryan stretches an expanse of superficial deposits about 30 square miles in extent (Jardine, 1966). Borehole records at West Freugh [NX 109543], five miles south-east of Stranraer, reveal that up to 43 m of sand and gravel have been deposited on a rockhead of red sandstone which can be placed well below current sea-level (Lawrie and Craig, 1945). Jardine (1966) introduces an interesting concept that nowhere in the neighbourhood of Stranraer is solid rock exposed. As the height of the land in the vicinity of West Freugh is about 15 m above OD, it has been suggested (unless the site overlies a deep hollow in the bedrock surface) that were the Quaternary sediment not present, the Rhins of Galloway would now be separated from the remainder of Wigtownshire by a deep channel of unknown width.

Explanations for the mode of formation of this channel appear to be mutually based on the erosion of this trough by either the exclusive work of one physical agent or the collective work of several. Two theories exist to explain the excavation of soft Upper Palaeozoic rock underlying the Stranraer isthmus: (1) glacial ice travelling southwards from Ayrshire during one or more glaciations of the Pleistocene (Jardine, 1966); or (2) the combined work of a Ryan-Luce river and marine waters attacking the west of Wigtownshire both from the north and south during one of the Pleistocene episodes of low sea-level (Jardine, 1966).

The deposits forming the isthmus are thought to mark a hiatus in retreat or possibly a slight readvance (Sections 2.5.4.3; 2.5.4.4; 4.2.4) of a fairly static ice front occupying Loch Ryan.

Stratified sand and gravel mounds, often with disturbed layering, are particularly common in the central and northern parts of the isthmus. These are interpreted as kame mounds, deposited beneath and between stagnant ice masses; when the ice finally melted, the stratified sand and gravel bodies were left as positive features although the removal of their support caused slumping along their margins. Around and beyond these mounds, in what would be a distal position from the glacier margin, an outwash plain is believed to have been built up by a system of braided meltwater streams. It is suggested that deposition may have been submarine as well as fluviatile following the discovery by Geikie and Irvine (1873) of shelly marine clay overlying sand and gravel, representing possible prograding delta foresets. Similar processes are suggested to have occurred on a smaller scale within ice-dammed meltwater lakes in close proximity to the glacier snout. Small lochans and vegetated hollows scattered throughout this area may have originated as kettle holes, formed by subsidence of the outwash sands once buried blocks of ice had finally melted.

Elsewhere on the Rhins peninsula, scattered irregular deposits of sand and gravel are thought to represent the eroded remains of previously more extensive outwash fans and terraces deposited during a deglacial phase. During deglaciation, sea-level fluctuated in response to eustatic-isostatic imbalance. The remains of shingle terraces containing a marine fauna at heights up to approximately 20 m above OD infer beach formation in response to relatively high Lateglacial sea-levels (Jardine, 1971). On more sheltered stretches of coastline, in particular the shores of Loch Ryan and the west coast of Luce Bay, many good examples of raised beach have been reported, ranging from about 5 m to 8 m above OD. On the west coast of the Rhins peninsula the only extensive raised beaches recorded are preserved in some sheltered bays and are of similar height above sea-level. At several localities the raised beaches are noted to continue seaward taking the form of a wave-cut platform at about 5 m above OD. On many stretches of the western Rhins coast similar rock platforms have developed but without the raised beach association although their height above sea-level is very variable within a range from 2m to 10 m. At the head of Luce Bay aeolian sand covers a wide area encroaching northwards on to raised beach deposits.

The present drainage system, as it developed, partially reworked drift deposits which occupied pre-existing valleys. Incision into accumulations of sand and gravel has left protruding terraces above modern alluvial terraces, notably in the valley of the Water of Luce. Elsewhere, patches of alluvial sand and gravel border many of the current streams, and on occasions are reported to overlie glacial drift and/or raised beach deposits. On the high ground in the north-east of the Rhins of Galloway district extensive peat mosses have built up to depths of several metres.

1.3.3.2: Whithorn

The till characteristics found in the Whithorn district have been described, on a limited basis, in the same manner as the deposits outlined in Section 1.3.3.1. They are believed to have a widespread coverage throughout the peninsula, following the undulations and irregularities of the underlying strata, particularly near to the southern point of Burrow Head (Barnes, 1989). Again, the drift is topographically most significant in the distinctive whale-backed drumlins which are prevalent across most of this region, although they diminish in concentration towards the south. The thickest deposits of till are reported along the west coast where they are draped over the coastline and locally include interbedded sand and gravel (Barnes, 1989). In places, the till has again been divided into two varieties, whose characteristics and mode of deposition appear to correspond with the upper and lower tills described above.

Glacifluvial sand and gravel deposits of the area have been categorised into three types (Barnes, 1989): (1) mounds with kame morphology (e.g. Arbrack [NX 461375]); (2) caps to low hills (e.g. in Monreith Park [NX 357425]); and (3) flat-topped outwash fans of which relics occur on the coast west of Cairndoon [NX 373392], at Burrow Head [NX 451342] and at Port Allen [NX 477409]. The kames are commonly unsorted and without stratification, however the other deposits are noted to be typically well-bedded. These beds comprise clean fine sand, are commonly laminated, and occur between units of variably sorted gravel in which the grain size ranges up to 40 cm. Clasts are reported as dominantly well-rounded, consisting of greywacke and granite, with some, more angular, shaly material (Barnes, 1989). In the Burrow Head deposit, the beds are recorded as dipping 10° to the south and some gravel horizons incorporate lenticular clasts with well developed grain imbrication (Barnes, 1989). These characteristics infer that the depositing currents in this area flowed southwards. On occasions in the west, where thick deposits of till overlie the coastline, they are commonly fringed by a well-developed raised beach terrace with a shingle surface at about 5 m above OD. Cut into bedrock, on the opposite east coast of the peninsula, a raised, wave-cut platform with thin shingle deposits occurs at the same elevation, backed by a discontinuous low feature marking the former coastline (Barnes, 1989).

Additional Quaternary deposits reported in the Whithorn area include spreads of peat grading into peaty clay which occur in low-lying areas above the till and filling hollows in bedrock. It is suggested that the peat deposits originated in a lacustrine environment (Barnes, 1989). Larger deposits, such as Whithorn Moss [NX 409422] and Glasserton [NX 424390], have been reported with depths of perhaps several metres thick. Also, thin alluvial deposits occur in the

Ket and the Drummulin Burn, the only significant drainage channels in the area, and comprise coarse sand and gravel with some dispersed larger clasts in a silty sand matrix (Barnes, 1989).

1.3.3.3: Wigtownshire (Kirkcowan and Wigtown)

In the Wigtownshire district, Irvine (1872) remarked that evidence of ice movement over the land was present in the form of discernible rounded hills and smoothed bedrock hummocks. This evidence is especially prominent among the lower hills of the area, where good examples of roches moutonnées have been found. In the majority of cases, it is believed that the rock surfaces have been exposed to subaerial weathering for a long period of time resulting in the obliteration of striae and other markings. Only where a fresh exposure of these rocks has been made have grooves and fine striae been observed on their surface (Irvine, 1878). The localities where striated rock surfaces have been observed suggest that the general trend of former ice flow was from the high ground to the north of the area, in a south and south-west direction into the Irish Sea (Section 1.4).

Deposits of till appear to present the usual character of a stiff clay, unstratified, full of smoothed and striated clasts of various sizes, and a general colour between grey and bluish grey (Irvine, 1878). The fabric of the till appears to generally correspond with the direction of ice flow suggested by the striae. Furthermore, numerous granite boulders from the Cairnsmore mass have also been found in the till on the lower ground supporting this direction (Irvine, 1878). Over most of the Wigtownshire area the till again occurs in the form of drumlins, with perhaps the best examples occurring between Glenluce [NX 205575] and Kirkcowan [NX 328608], and also around Wigtown [NX 415545], Kirkinner [NX 414513] and Sorbie [NX 433473] (Irvine, 1878). It has been noted that these landforms generally run in parallel but independent ridges varying in height from ten to forty metres, though on occasions they are clustered together in groups of four to six connected mounds (Irvine, 1878). In the north-west of the district the drumlins are reported to be generally surrounded by large flat peat mosses, while in other places they rise from bare rocky ground sparsely covered with vegetation. Only in a few cases have drumlins standing out of smooth covers of till been recorded (Irvine, 1878).

The presence of a sand and gravel series is noted to be poorly represented in this district. A few small kames do occur over the moorlands of Wigtownshire, and gravel deposits have been seen along the valleys of the Cree and Bladenoch, the shores of Luce Bay, and at the head of Wigtown Bay (Irvine, 1878). It is suggested that the best development of kames lies about five miles north-west of Kirkcowan. Considerable deposits of gravel are also reported to cover portions of the valley of the Cree north of Newton Stewart, and are well seen in places on either side of the river (Irvine, 1878).

In the high ground of Cairnsmore of Fleet traces of loose, stony and sandy moraine drift have been recorded which are believed to represent the position of small glaciers which were nourished in these areas towards the end of the last glacial period (Irvine, 1878). A group of well-marked moraine mounds are also present on the moor north of the dismantled railway between former Creetown Station [NX 475599] and Culcronchie Viaduct [NX 507613]. Many of the more prominent moraines in this area, however, are thought to occur on Cullendoch Moor [NX 558655], near the sources of the Big Water of Fleet, and along the line of the former railway between the Fleet viaduct [NX 558643] and Gatehouse station [NX 546625] (Irvine, 1878). The drift, where exposed, appears to be generally very loose, sandy, and gravelly with rounded or subangular clasts of mostly granite and occasional angular fragments of schist.

Along the shore of Luce Bay, from Auchenmalg Bay [NX 233516] to Port William [NX 337438], a narrow platform of gravel extends inland from the present beach for 50 to 100 m; and is thought to mark the most recent of the raised beaches in this district (Irvine, 1878). It sits at approximately 10.7 m above OD (35 feet), and terminates along its inner edge at the base of a vegetated escarpment which for the most part consists of till. Along the east side of Wigtown Bay considerable fragments of the same terrace have been reported, while along the north side of the same bay from Creetown [NX 476585] to Ravenshall [NX 522523] it forms a strip of flat ground between the present high-water mark and the former sea-cliff (Irvine, 1878).

Alluvial deposits are apparently limited in extent in the Wigtownshire area. They predominantly lie in the main river valleys and along some of the principal tributaries. A considerable extent of marine alluvium has been mapped at the head of Wigton Bay, on either sides of the mouths of the Cree and Bladenoch (Irvine, 1878). It consists of large mudflats which are exposed at low-tide. The whole of Fleet Bay is also occupied by a more sandy marine alluvium, likewise exposed during recession of the tide. Fragments of older river terraces have also been seen flanking the lower and more recent alluvium of several of the major rivers (Irvine, 1878). At the junction of the Tarf and Bladenoch [NX 347604], south-east of Kirkcowan, portions of two such higher terraces reportedly rise above the existing river flats (Irvine, 1878).

Hilly ground in Wigtownshire is for the most part covered with peat, with the largest peat mosses occurring on the low ground between Glenluce and Kirkcowan. On the higher ground, the peat forms only a thin covering. Other extensive accumulations of peat have been reported in some major hollows between drumlins, such as near Mochrum [NX 346464] (Irvine, 1878).

1.3.3.4: South-west Ayrshire (Girvan)

In the region of south-west Ayrshire, there is concurrent evidence with the other districts that two tills were deposited during the last glaciation (Young, 1869). The lowest till has been reported, for the most part, in stream cuttings where these incisions are deep enough. Its usual character is that of a stiff, light-brown, and sometimes greenish-grey clay, containing numerous well striated and polished clasts (Young, 1869). The upper till is generally of a "dirty-brownish colour" (Young, 1869, p. 14) and notably has a larger, more angular, and less polished clast content than corresponding stones in the lower deposit. It has also frequently been found to be stratified, containing significant quantities of sand and gravel (Young, 1869). Many large boulders are also reported to occur in this drift, which appear to pass into a coarse shingle and boulder gravel. This latter kind of drift has been seen scattered over the hilly ground south of Stinchar valley, where it is noted to form mounds in places. Between this deposit and the upper till occasionally occur beds of 'earthy gravel' and sand marking an interface (Young, 1869).

In the hollow of the ridge between the Fell and Byre Hills, above Ardmillan House [NX 173942], ice-eroded hummocks of diorite are noted which in places still retain their striae (Young *et al.*, 1869). Where the coastline is low, the striae on the rocks reflect the ice flow direction from the interior (approximately east-west). Examples of this have been provided from Shalloch Mill [NX 179956], where at the time of initial investigation, there had been removed a covering of till revealing a fresh, well-scratched, and smoothed bedrock surface. However, where the ground rises more steeply from the coastline, the trend of the striae is reported to correspond on the whole, with that of the coastline (north-south) (Young, 1869). Examples of this relation have been given from the flanks of Pinbain Hill [NX 147924], at Downan Point [NX 068806], and a little south of the shore end of Stinchar fault.

Other deposits which have been recorded in the area included kames of sand and gravel north of Knockdolian Hill [NX 113848], and sporadic mounds of coarse gravel and angular debris along the northern flank of the Auchencrosh Hills [NX 107788] (Young *et al.*, 1869). It has been suggested that these drift forms appear to be connected with the upper till, however no explanation of this relationship is given. Erratic blocks are reported in a number of locations within south-west Ayrshire, and often have been seen to attain a great size on many occasions (Young, 1869). Several very large boulders of greywacke are noted at Pyell Craig, on the north bank of the Stinchar between Ballantrae and Colmonell. Granite boulders, which are thought to originate from the Loch Doon area, are also present in the area, particularly in the valley of the Girvan, and on the beach north of the mouth of that river (Young, 1869).

One of the most marked features of the coastal scenery of this area is a raised beach at approximately 7.5 m above OD (25 feet) (Young, 1869). It runs from Girvan to Ballantrae along the coastline as a narrow platform, with the inner margin representing the former coastline, from which the hills further inland rise steeply. The terrace is reported to be predominantly cut out of the till and continues southwards from Girvan to Lendalfoot and Ballantrae, although in places it is excavated into bedrock (Young, 1869). Occasionally, the raised beach has been heightened due to apparent accumulations of wind-blown sand. Traces of old coastlines at higher levels than the 7.5 m raised beach have been recorded at Girvan and Ballantrae with two such terraces at the latter location standing at approximately 15 m (50 feet) and 23 m (75 feet) above OD (Young, 1869). Alluvial sediments are confined to the current course of streams in the area, namely the Girvan and Stinchar.

The whole district south of Glen App reveals little of the underlying bedrock or drift as the area is dominated by swathes of peat. This growth appears to have attained its greatest thickness on the higher grounds and gentler slopes, whereas on steeply inclined ground the peat fails to reach a thickness much greater than one metre (Young, 1869).

1.3.3.5: Carrick and Loch Doon

In accordance with what has been reported elsewhere in the Galloway region, over much of the ground in this district lies an irregular mantle of till consisting of a stiff stony clay, with numerous pebbles and boulders which are commonly rounded and striated (Floyd, 1999). The till deposit is thought to be particularly thick in valleys and lower ground in the north-west and patchy, if not absent, from the hill tops. In the south and east of the area, rock is apparently visible at the surface over large areas and there are only small isolated patches of till infilling some of the valleys (Floyd, 1999). The till is thought to be of local origin since its character and erratic content normally reflect the nature of the underlying bedrock. Over areas of Lower Palaeozoic rocks, the till is reported to be brownish grey and contains mostly boulders of greywacke and granite, the latter derived from the Loch Doon mass (Floyd, 1999). Where underlain by Old Red Sandstone strata, the till is found to be red, sandy and full of boulders of red sandstone. Only in particular locations has the till been shaped into drumlins, notably the upper reaches of the valley of the Duisk River, south-east of Barrhill [NX 266792], where they are recorded as having a north-north-west trend (Floyd, 1999). A contrasting upper till, which is more sandy and friable than the lower, is thought to represent an ablation till (Floyd, 1999).

This till has been found to be sometimes crudely stratified, containing more angular boulders than the underlying lodgement till and may be locally separated from the latter by a thin unit of silty sand and gravel (Floyd, 1999).

Throughout the Carrick-Loch Doon district mounds, ridges and spreads of unsorted and poorly consolidated debris are common. These deposits are considered to represent moraines formed during advance and retreat of the Late Devensian ice sheet, leaving behind a heterogeneous mixture of clay, sand and gravel with occasional very large erratics, whose lithology is of local provenance (Floyd, 1999). In places, it appears difficult to distinguish the morainic material from the upper ablation till, especially in areas where both are said to rest on the lower lodgement till. By contrast, only small scattered patches of glacial outwash deposits are reported to occur in the area (Floyd, 1999). Typical examples have been noted around the headwaters of the Water of Assel [NX 244947], on the south bank of Penwhapple [NX 262968], in the valley of the Duisk River near Barrhill [NX 221831], and in the Cree valley south of Clachaneasy [NX 354750]. They consist of isolated mounds and flat spreads of well-bedded sand and gravel (Floyd, 1999).

Following ice retreat in this area, it is suggested that many of the watercourses were dammed by isolated masses of melting ice or morainic material, allowing temporary lakes to form (Floyd, 1999). Deposits of clay, silt and sand appear to have been laid down in these lakes, which were either silted up or were abandoned as the postglacial drainage became established, leaving behind flat spreads of lacustrine alluvium (Floyd, 1999). In places, it has been noted that present-day streams have incised themselves into these deposits causing them to stand as higher terraces above more recent alluvial terraces. Examples are given along the valley of the Cree near Brigton [NX 360745], at Loch Slochy at the head of the Balloch Lane [NX 425923], and alongside the Carsphairn Lane at Carsphairn [NX 560932] (Floyd, 1999).

Throughout the district the growth of peat mosses has taken place with reported accumulation in areas up to several metres thick. In the more mountainous eastern part of the district, however, the peat is thought to be mostly restricted to the valleys where it often forms on top of the morainic deposits. The more subdued topography in the south-west has allowed the accumulation of a huge blanket of peat, more than 100 km^2 in extent (Floyd, 1999).

Table 1.1 History of prominent Quaternary research conducted in Carrick and Galloway regions, south-west Scotland.

Discipline	*Chronology of research*	*Nature of investigation*
Glacial modelling	Jolly, W. (1868)	Glacier action in Galloway.
	Geikie, A. (1901)	The scenery of Scotland
	Gregory, J.W (1925)	Moraines, till and glacial sequences in south-west Scotland
	Charlesworth, J.K. (1926a)	Glacial geology of the Southern Uplands of Scotland, west of Annandale and Upper Clydesdale.
	Eckford, R.J.A. (1957)	The Merrick region: Galloway's glacial park
	Cutler, H.D. (1979)	Glaciation and drumlins of Galloway, south-west Scotland
	Kerr, W.B. (1982a)	Pleistocene ice movements in the Rhins of Galloway
Lateglacial readvance(s) (supports§ rejects† reviews‡)	§Charlesworth, J.K. (1926b)	Readvance marginal kame moraines of the south of Scotland and some later stages of retreat
	§Sissons, J.B. (1967)	Evolution of Scotland's scenery
	†Sissons, J.B. (1974)	The Quaternary in Scotland - A review
	†Cutler, H.D. (1979)	Glaciation and drumlins of Galloway, south-west Scotland
	†Sissons, J.B. (1981)	The last Scottish ice-sheets: Facts, speculation and discussion
	‡Kerr, W.B. (1982c)	Ice advances in Galloway?
Loch Lomond readvance	Charlesworth, J.K. (1926a)	Glacial geology of the Southern Uplands of Scotland, west of Annandale and Upper Clydesdale.
	Moar, N.T. (1969)	Late Weichselian and Flandrian pollen diagrams from south-west Scotland
	Cutler, H.D. (1979)	Glaciation and drumlins of Galloway, south-west Scotland
	Cornish, R. (1981)	Glaciers of the Loch Lomond Stadial in the western Southern Uplands of Scotland
Drumlin analysis	Geikie, J. (1905)	Structural and field geology
	Gregory, J.W. (1926)	Scottish drumlins
	Hollingworth, S.E. (1931)	Glaciation of western Edenside and the drumlins of Edenside and the Solway basin
	Smith, H.M. (1971)	South-west Scotland: drumlin analysis
	Cutler, H.D. (1979)	Glaciation and drumlins of Galloway, south-west Scotland
Raised beaches	Donner, J.J. (1963)	The Late and Postglacial raised beaches in Scotland
	Jardine W.G. (1964)	Postglacial sea-levels in south-west Scotland
	Jardine, W.G. (1967)	Sediments of the Flandrian transgression in south-west Scotland
	Jardine, W.G. (1975)	Chronology of Holocene marine transgression and regression in south-west Scotland
	Jardine, W.G. & Morrison, A. (1975)	The archaeological significance of Holocene coastal deposits in south-west Scotland
	Jardine, W.G. (1977)	The Quaternary marine record in south-west Scotland
Periglaciation	Galloway, R.W. (1969)	Ice wedges and involutions in Scotland
	Watson, E. (1977)	The periglacial environment of Great Britain during the Devensian
Climate reconstruction	Moar, N.T. (1969)	Late Weichselian and Flandrian pollen diagrams from south-west Scotland
	Bishop, W.W. & Coope, G.R. (1977)	Evidence for Lateglacial and Early Flandrian environments in south-west Scotland

1.4: Review of Quaternary Studies and Present Status of the Late Devensian Glacial History of South-West Scotland

The following account expands upon an earlier review by Kerr (1982b) of work already published in this field. It is beyond the scope of this section to describe in detail each individual investigation which has been undertaken in the Carrick and Galloway districts, and so the reader is directed to the source text if further information is required (Table 1.1). Section 1.4.2 provides an understanding of the current position of glacial reconstruction in the region of south-west Scotland.

1.4.1: Review of Quaternary research in south-west Scotland

The first descriptions of physiographic features in south-west Scotland occurred in the second half of the nineteenth century when interest in natural history and the landscape was abundant. Investigations tended to be concentrated on the more prominent landscape features such as waterfalls although there was always an underlying curiosity as to what action shaped the landscape. This can be picked up in the literary style to which descriptions of certain landscapes were presented suggesting that early workers had an undoubted feel for the landscape. An example of this genre is the paper by Jolly (1868, p. 155) on "The evidence of glacial action in Galloway" where the following description of the corrie backing Loch Dungeon [NX 516840] is provided:

> "Right before us rose a wall of greywacke, precipitous, serrated, gashed and furrowed to a wonderful degree and descending sheer into the dark waters of a lake. Standing by its shores and looking right across to the wild thunder splintered pinnacles and torn precipitous peaks of Millfire and Mildown one feels that he looks on one of the sublimer phases of nature and is over-powered with a wave of savage, solitary, grandeur and of the almighty forces that have torn and changed the face of the globe".

Jolly later addressed the Edinburgh Geological Society on the 5th March 1868 on the topic of glacial erosion in the uplands of Galloway and on the 19th March 1868 on the glacial deposits of the lowlands.

In the early part of last century James and Archibald Geikie were renowned for their contributions to what then was referred to as 'the problems of the ice age'. If doubts existed as to the extent or severity of glacial activity in south-west Scotland, compared to that reported in the Highlands, these were soon dispersed by Archibald Geikie in his publication entitled 'The Scenery of Scotland' (1901). In the opening paragraph to the text on 'The ancient glaciers of the Southern Uplands' Geikie (1901, p. 341) remarks:

> "That they have been intensely glaciated, however, will be recognised by any one who seeks for proofs of ice-work. In spite of the thick mantle of boulder-clay that covers so much of the valleys and lower hill-slopes, and in spite also of the tendency of so many of the rocks to decay and to be concealed beneath a coating of turf or peat, abundant polished and striated rocks may be found in every district from the headlands of Wigtownshire to those of St. Abb's Head. In some parts of Galloway, indeed, the roches moutonnées are hardly less perfect and conspicuous than in most of the Highlands".

From the direction of striae and distribution of scattered erratics, Geikie (1901) concluded that it was clearly evident that the Southern Uplands formed a centre of dispersion for the southern part of a Scottish ice sheet. Evidence for this is given in summary form on a map of the glaciation of Scotland (Figure 1.6) which accompanies the text, Geikie (1901, p. 341-42):

> "A vast mass of ice flowed northwards into the plain of Ayrshire, where, joining the stream that was descending from the Highlands, it bent round to the west and went southwards down the Firth of Clyde. Still thicker and more extensive was the great ice-field that crept off the southern side of Galloway into the Solway Firth and the Irish Sea, both of which, at the height of the Ice Age, were filled with ice".

This map provides the foundation for all subsequent models of glaciation in the area to the extent where the official handbook for the Southern Uplands (Greig, 1971) contains an almost identical diagram (Figure 1.7). Meanwhile, James Geikie focused his attention on the mode of deposition of glacigenic sediment in the Galloway region. Geikie (1905) recognised that the drift covering this area was perhaps not just a simple deposit of one glacier but instead a complex deposit comprised of many separate layers of boulder clay (till). This theory was developed to explain the formation of drumlins in Galloway as resulting from the erosion of a previously deposited sheet of till, the product of one glaciation, by a later advance of ice (Geikie, 1905).

On the 12th March 1925 Professor J.W. Gregory of the University of Glasgow presented a paper entitled, "The moraines, boulder clay, and glacial sequence of south-western Scotland", to the Royal Society of Edinburgh. One of the main themes of the investigation was to explain the restricted distribution of moraines in south-west Scotland, and to understand the significance of this occurrence despite excellent preservation of moraines in some localities. Gregory (1925) attempted, on the whole, to explain the limited dispersion of moraines in terms of non-depositional events; the partial deposition of morainic material into adjacent seas; and destruction of moraines during glacial episodes.

The following year Gregory (1926) published a second piece of research which considers the origin of Scottish drumlins, with particular reference to those in the vicinity of Glasgow.

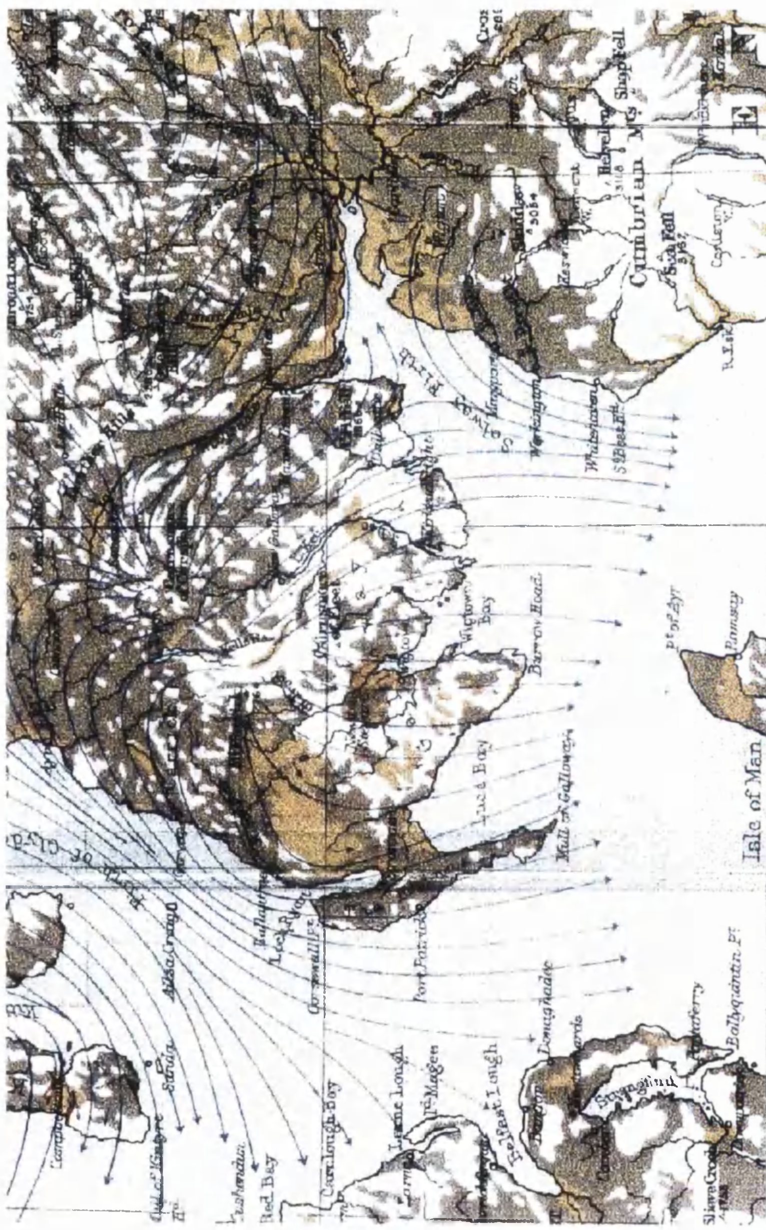


Figure 1.6 Archibald Geikie's original map of glaciation in south-west Scotland, north-west England and north-east Ireland. (From Geikie, 1901).

Gregory dismissed the involvement of ice in drumlin formation and replaced glacial action with the work of the prevailing wind, Gregory (1926, p 436):

> "The clue to the origin of the Scottish drumlins is given by the trees which grow on them. The trees show the direction of the prevalent wind, and they are bent parallel to the drumlins. The normal drumlin direction is, in fact, that of the prevailing wind, and most Scottish drumlins may be explained as the product of sub-aerial denudation of sheets of boulder clay. The rain beats most frequently and violently against the western end of the drumlin, which is therefore the steeper; and the drumlins taper leeward with streamline curves due to wind and rain".

However, Gregory (1926) acknowledges that there are many exceptions in Scotland to the parallelism of drumlins and the prevailing wind. The influence of the underlying bedrock is used to explain these anomalies, with a typical example taken from the drumlin field on the Machars and Whithorn peninsula in Wigtownshire, Galloway district. In this area Gregory (1926) observed that the southern drumlins trend approximately north to south, at right angles to the prevailing wind, and noted that here the till is thin and denudation has revealed the Lower Palaeozoic bedrock. This led Gregory (1926) to invoke the following scenario that in the first instance glaciers had flowed down the Cree valley carving the Lower Palaeozoic bedrock into an irregular surface with hummocks trending north and south. The till was subsequently deposited as a continuous sheet, and during its 'consolidation' would have shrunk predominantly where it was thickest; so even if the surface had been originally level, meltwater streams would have formed over the depressions in the bedrock, incised into the till, and left mounds of till on the rock hummocks (drumlins). In support of this, Gregory (1926) argues that the drift to the north of this area is much thicker and so the underlying bedrock is rarely exposed, hence the drumlin field curves into the east-north-east direction, taking on a long-axis orientation parallel to the prevailing wind and at right angles to former ice flow direction.

Also in 1926, J. Kaye Charlesworth contributed two significant papers to the Transactions of the Royal Society of Edinburgh, read to the society by another eminent geologist of the period, Dr. J. Horne. Together, these papers were the product of dedicated fieldwork and until recently provided the only substantive investigation in this region. Charlesworth's first paper (1926a) entitled "The glacial geology of the Southern Uplands of Scotland, west of Annandale and Upper Clydesdale" describes in great detail the proposed general character of Late Devensian glaciation. A regional-scale representation is provided in Section 2.4.2 however in short the main conclusion of Charlesworth's investigation was the endorsement of Geikie's (1901) model of a central Galloway ice dispersal centre. Many of the ideas on Late Devensian ice movements stemmed from the careful mapping of various erratics to the north, west and south of the mountains in the Loch Doon area, together with mapping of the distribution of ice-moulded landforms and striated bedrock surfaces.

In the Machars of Wigtownshire, Charlesworth (1926a, p. 20) identified a change over a short distance in the direction of the long axes of drumlins which were separated by what was described as a moraine (Section 2.5.4.4). The view that these two sets of drumlins were produced simultaneously was dismissed and Charlesworth (1926a, p.20) concludes:

> "It seems necessary, therefore, to conclude that the drumlins within and without the moraine are the products of two distinct glaciations separated by a retreat of uncertain magnitude. The drumlins are regarded as sub-glacial or englacial features".

This statement required an 'oscillation' in the ice sheet which was only detected where for any reason the direction of ice flow changed. Landforms associated with this proposed readvance have been described in a subsequent paper by Charlesworth (1926b) which make up what is termed the 'Lammermuir-Stranraer Moraine' (Section 2.5.4.3). The argument for this moraine limit was based on the recognition of sand and gravel deposits with characteristic kamiform topography. In the west for example, Charlesworth associated the 'kame and kettle' topography (consisting of kames, kettle holes, kame terraces, eskers and meltwater channels) at Leswalt in the Rhins of Galloway [NX 025640] with the sands and gravels at the head of Loch Ryan. Similar topography and deposits at Glenluce [NX 192575] and Kirkcowan [NX 328608] extended the proposed readvance line further east. The basis for this interpretation was therefore field evidence which later workers disputed (e.g. Sissons, 1974, 1981; Price, 1975, 1977; Cutler, 1979; Section 2.5.4.4).

After 1926 there occurs a sedentary period with respect to publications until the appearance of the 1st edition of the geology guide to the 'South of Scotland' in 1935. This volume was edited by J. Pringle and appeared in revised form in 1948 with few alterations made to the original manuscript. In the chapter on 'Glacial and Superficial Deposits' the model of glaciation presented by Geikie (1901) and Charlesworth (1926a) is replicated with little comment (Figure 1.7). These publications were prepared by the then Geological Survey and were based upon the published 1:63 360 geology maps (combined solid and drift) and accompanying memoirs. Some of the memoirs date from the late nineteenth century (e.g. Whithorn from 1873) indicating the long history of geological mapping in south-west Scotland. In 1981 the 1:50 000 scale sheets were released and the solid and drift geology separated for the first time and published as separate maps. These new maps provide invaluable information for any study of Quaternary events in the United Kingdom. The most recent edition of the 'South of Scotland'

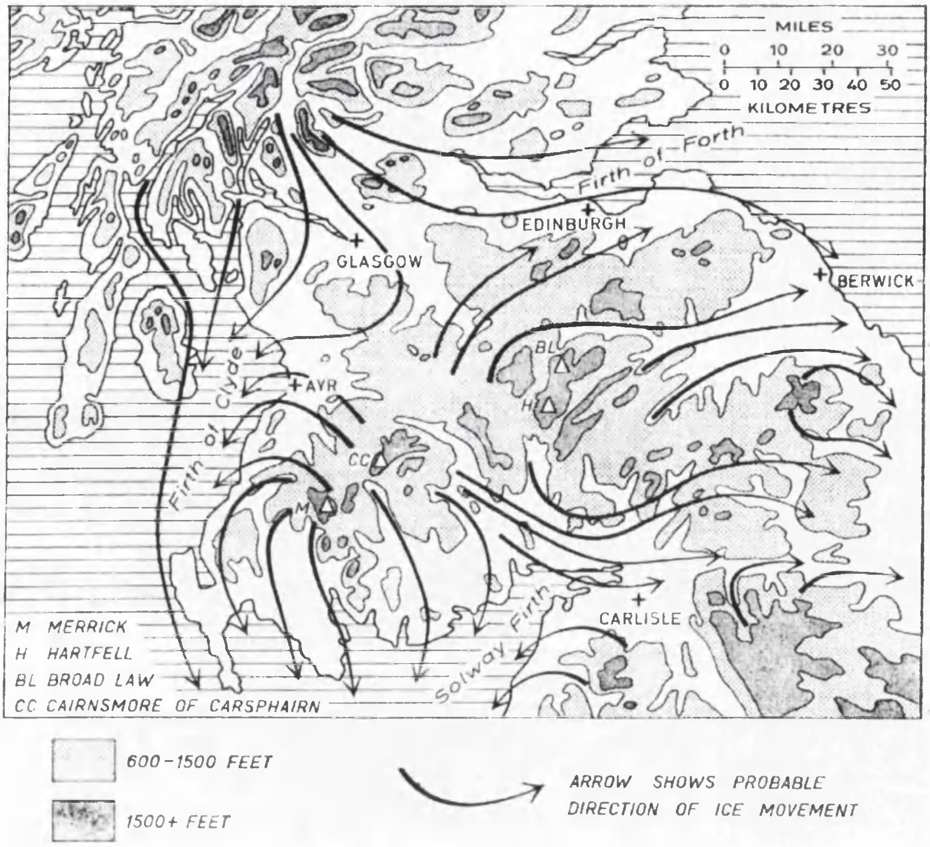


Figure 1.7 Suggested directions of ice movement over southern Scotland by the Institute of Geological Sciences (From Greig, 1971).

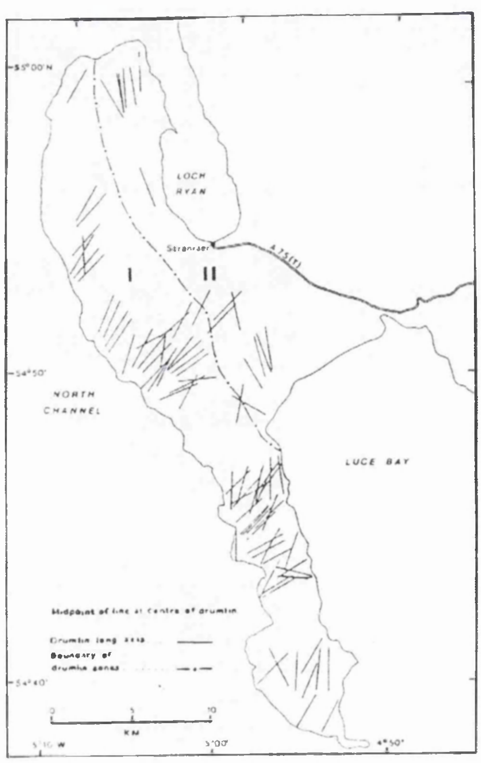


Figure 1.8 Drumlin orientations and zones. (From Kerr, 1982a).

was published in 1971 and edited by D.C. Greig. The chapter on the 'Pleistocene and Post-Glacial' is based largely on the pioneering work of Charlesworth (1926a, b) and maintains the concept of a readvance.

During the period 1926 to 1979 no further published investigations, to the knowledge of the author, were produced which concentrate upon Late Devensian ice sheet reconstruction in south-west Scotland. However, a recent revival of interest over this subject has taken place in the form of authoritative studies by Cutler (1979) and Kerr (1982a). Both investigations pursue the origin and nature of ice movements, and glaciation chronology, over the Machars of Wigtownshire and the Rhins of Galloway respectively, basing their interpretation upon evidence gathered from till fabrics, erratics, striae and drumlin orientations. Of significance throughout the Rhins is a swing in the trend of the drumlins, identical to that noted earlier on the Machars, from the dominant north-east to south-west trend which is particular to the eastern area of the peninsula. The trend is so persistent that Kerr (1982a) has divided the drumlins of the Rhins of Galloway into two zones (Figure 1.8). The majority (Zone I) conform to the dominant ice movement direction as determined by fabric analysis and are aligned north-east to south-west. The dominant direction in Zone II is at right angles to this with drumlin axes aligned north-west to south-east.

Following deglaciation of the main Devensian ice sheet, the mountains which have the greatest elevations nourished small glaciers in what is customarily called the 'Loch Lomond Readvance' (Section 2.5.4.5). Among many of the north-facing corries have been left remarkably fresh looking morainic deposits which Jolly (1868) noted. Cornish (1981) has described these deposits and Moar (1969) produced pollen diagrams from lake deposits, which placed the readvance in the period 10 200-10 000 yrs B.P. Charlesworth (1926a) referred to this readvance as the 'Corrie Stage' and Cutler (1978) the 'Corrie-Merrick Stage'. During the Loch Lomond stadial the lowlands, while free from ice, would have experienced periglacial conditions. Watson (1977) refers to ice wedges at Newton Stewart and a review of the regional evidence can be found in Galloway (1969). Similarly, a review of Late Devensian environmental conditions in south-west Scotland is provided in Bishop and Coope (1977). Moar (1969) discusses the pollen diagrams which have been produced from the region which suggest that most of Galloway would have been ice free by 12 500 yrs B.P.

Investigations into Late Quaternary shorelines and coastal deposits are extremely scarce in south-west Scotland perhaps because of the perplexing interplay between eustatic changes in sea-level related to continuous change in the volumes of the world's glaciers, and isostatic uplift of the land due to the removal of the ice load (Kerr, 1982b). The complexity of unravelling the field evidence is described by Jardine (1977, p. 100):

> "Evidence of former high levels of the sea in relation to the land takes various forms. Remnants of old sea cliffs now some distance inland or high above present sea level, raised shore platforms cut in solid rock or in glacial till, river terraces graded to heights above sea level, raised estuarine beach or coastal sediments of saline or brackish water environments all bear witness to former sea levels relatively higher than at present".

The first work to be published in this field was by Donner (1963), who described the 'Late and Postglacial beaches of Scotland'. Evidence was recorded for marine transgression at seven sites along the Solway Firth coastline and discussed in terms of a 'Lateglacial 100 foot beach' and a 'postglacial 20 foot beach'. This work has since been advanced by Jardine (1971, 1975, 1977).

1.4.2: Present status of the glacial history of south-west Scotland

It is strikingly apparent from the review above that British geomorphologists over the last century have made little effort to avail themselves of the opportunity to apply modern multidisciplinary approaches in an area which offers a remarkable regional diversity of landforms and Quaternary deposits. As Table 1.1 illustrates there is an unexplainable scarcity of research conducted in south-west Scotland throughout the history of Quaternary studies. Many investigations that have been published, have devoted themselves to particular aspects of glacial phenomena (e.g. Jardine, 1964; Galloway, 1969) rather than advancing the earlier work of Geikie (1901), Gregory (1925) and Charlesworth (1926a, b). These studies may be viewed by the individual as important for the advancement of their particular interest in Quaternary research, but they fail to provide any significant clarity to the regional-scale picture. Furthermore, the aims and approaches of individual workers involved with such research has been equally varied, leading more often than not to contrasting results and opinions.

Despite the long history of glacial research in south-west Scotland, an appreciation of the scope of glaciation in shaping the evolution of the landscape has only started to widen over the last two decades, explaining the fifty year gap of research in the region. Parallel to this explanation can be reiterated the fact that sedimentological analysis of glacial deposits appears to have only recently engaged researchers in Galloway (e.g. Cutler, 1979; Cornish, 1981; Kerr, 1982a; Section 1.2.1). Therefore the current status is that there has been a restricted growth in the sophistication and level of Quaternary studies in south-west Scotland compared to that of other parts of the British Isles. Consequently, the current glaciological community still relies heavily on the efforts of the pioneering studies which may potentially have failed to effectuate the maximum resolution of available field evidence. The need for this cautionary note is warranted given the role played by south-west Scotland in the broader context of glaciation in the Irish Sea Basin (Kidson and Tooley, 1977). A full treatment of this subject is presented in Chapter 2.

1.5: Synopsis

This chapter has outlined the following points:

- During the maximum extent of the Dimlington Stadial, major ice dispersal centres in Scotland were located in the mountain massifs of the Northern and West Grampian Highlands and the Southern Uplands of south-west Scotland.
- The central location of south-west Scotland in relation to the Scottish Highlands, north-west England, north-east Ireland, north Wales and the Isle of Man is of fundamental importance in the context of correlating the Late Devensian glaciation of these adjacent areas.
- There has been no regional synthesis which would place south-west Scotland in the context of Quaternary events in Northern Britain. The area has been considered as an 'appendage' to the research of other peripheral regions.
- The accuracy of many early studies throughout the British Isles is questionable, given that in recent years several major deficiencies in applied traditional approaches have been identified. Modern regional studies have identified and placed greater emphasis upon links between sediment-landform association and genetic processes than conventional lithostratigraphic practices.
- The status of readvances in south-west Scotland is based on evidence which is suggestive, but not equivocal.
- Palaeoenvironmental interpretation of glacigenic sequences on the north-east seaboard margin of the Irish Sea have failed to take account of influences that a marine ice sheet might have had on the sedimentary record.
- Quaternary research in south-west Scotland has never attracted the volume of work as in other regions of Scotland, despite its regional diversity of landforms and Quaternary deposits. Reference has been made to this region as one of the 'black-holes' in Scottish glacial geology.
- The current glaciological community still relies heavily upon the efforts of the pioneering studies of Geikie (1901), Gregory (1925) and Charlesworth (1926a, b). This emphasises the static state of Quaternary studies in Galloway and Carrick throughout the twentieth century.

A general overview of landscape characteristics, pre-Quaternary geological history and Quaternary deposits of the Carrick and Galloway districts has also been accommodated.

CHAPTER TWO

Glaciation within the Irish Sea Basin

CHAPTER TWO

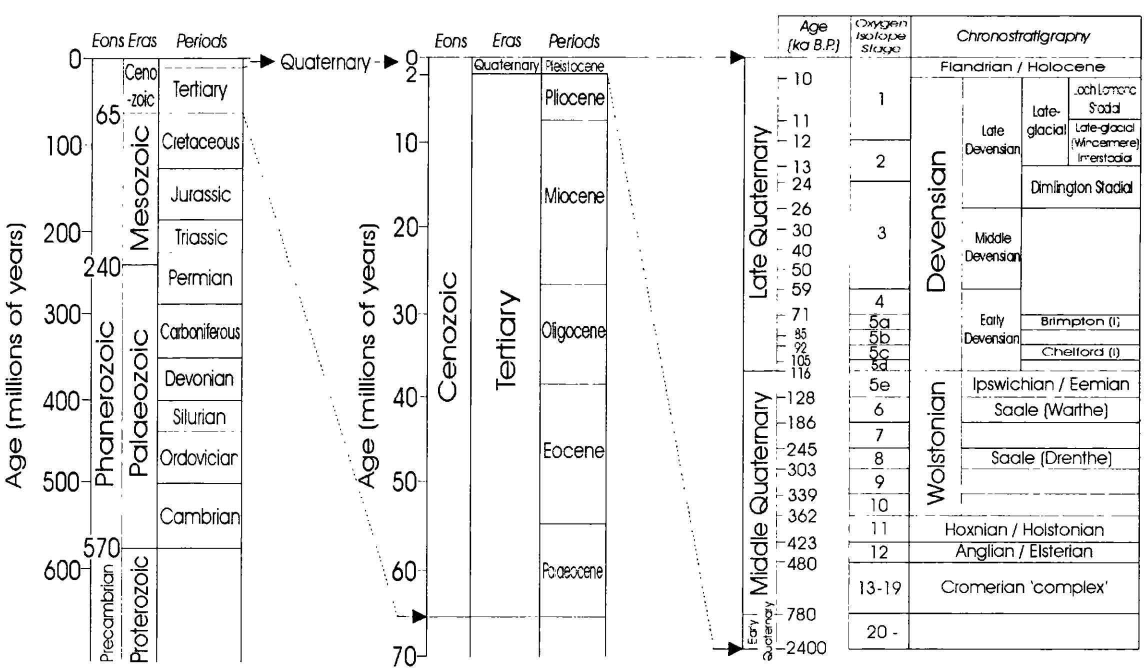
Glaciation within the Irish Sea Basin

2.1: Introduction

One of the main directions of this research is to establish a regional synthesis which will place the Galloway and Carrick districts of south-west Scotland in the context of Quaternary events in the Irish Sea Basin. It is therefore necessary to address this theme from a wider perspective. This chapter explores the structural and stratigraphic development of the Irish Sea Basin and the imprint left by episodic glacial and interglacial events.

The first half of the chapter commences with a description of the physical characteristics of the Irish Sea Basin (Section 2.1). This is followed by a chronological account of pre-Devensian and

Table 2.1 The Quaternary and its divisions in relation to the geological time-scale. (Modified from Lowe and Walker, 1984 and Gordon, 1997).



Devensian events (Sections 2.2, 2.3) which leads to an assessment of full glacial ice dynamics as currently inferred for south-west Scotland and the Irish Sea Basin (Section 2.4). The second half of the chapter concentrates on deglacial events by addressing the following issues (Sections 2.5, 2.6): (1) proposed models of deglaciation (terrestrial and marine); (2) ice sheet readvances and oscillations; and (3) classification of depositional systems throughout the Irish Sea Basin. Finally, an overview is provided of sea-level history in this region (Section 2.7). Table 2.1 summarises the geological time-scale referred to in this chapter.

2.1.1: Physical setting of the Irish Sea Basin

The Irish Sea Basin (4000 km^2) lies at the centre of the British Isles, and is surrounded by Ireland, southern Scotland, north-west England and North Wales (Figure 2.1). The Irish Sea has restricted outlets to the Atlantic Ocean through the graben-like North Channel in the north-west separating south-west Scotland from Northern Ireland (<20 km), and through the St. George's Channel to the south constricted by the Pembroke Peninsula of south-west Wales and the Llêyn Peninsula of North Wales (<60 km). Centrally the basin broadens to over 230 km at its widest point immediately south of the Isle of Man.

A strict demarcation of the Irish Sea Basin would confine it to the area stretching between England and Ireland, however for convenience of description it is more practicable to regard all the sedimentary basins (North Channel Basin, Stranraer Basin, Solway Firth Basin, East Irish Sea Basin, Peel Basin, West Irish Sea Basin, Central Irish Sea Basin, Cheshire Basin, Caernarfon Bay Basin, Kish Bank Basin, St George's Channel Basin; Naylor and Shannon, 1982; Jackson *et al.*, 1995) lying between the whole of Great Britain and Ireland as the 'Irish Sea Basin' (Figure 2.2). That part lying north of a line from Holyhead, North Wales to Dublin, Southern Ireland may be described as the northern Irish Sea Basin with the area to the south as the southern Basin (Kidson, 1977). Increasingly, the term "Celtic Sea" has come to be applied to the area between the south coasts of Ireland and Wales and the Atlantic coast of Cornwall, south of the southern approaches to St. George's Channel. The offshore area considered in this section is bounded to the north and south by latitudes 55° 20′ N and 52°N (*c.* 360 km), within which it extends to the coasts of south-west Scotland, Northern Ireland, England, north-west Wales and the Republic of Ireland, and surrounds the Isle of Man.

Bathymetrically, the Irish Sea Basin is largely composed of platforms (<60 m water depth) bordering the Scottish, English, Irish and Welsh coasts (e.g. Solway Firth, Liverpool Bay, Cardigan Bay, Caernarfon Bay), within which there are very localised enclosed deeps giving water depths down to 137 m (Eyles and McCabe, 1989a; Jackson *et al.*, 1995) (Figure 2.1). An extensive deep-water zone is formed by a sinuous, north-to-south trough, 30 to 70 km wide, that extends from off Islay in the north, through the North Channel, the Manx Depression and the St. George's Channel, to the Celtic Deep (Jackson *et al.*, 1995). In this trough water depths are generally 60 to 120 m but, as on the platforms, deeper waters lie in separate, elongate, enclosed deeps. The most notable deeps are located within the physiographically complex Beaufort's Dyke in the North Channel, which is locally 318 m in depth (Jackson *et al.*, 1995). For means of classification Jackson *et al.* (1995) have divided bathymetric features on the Irish Sea floor into five characteristic zones:

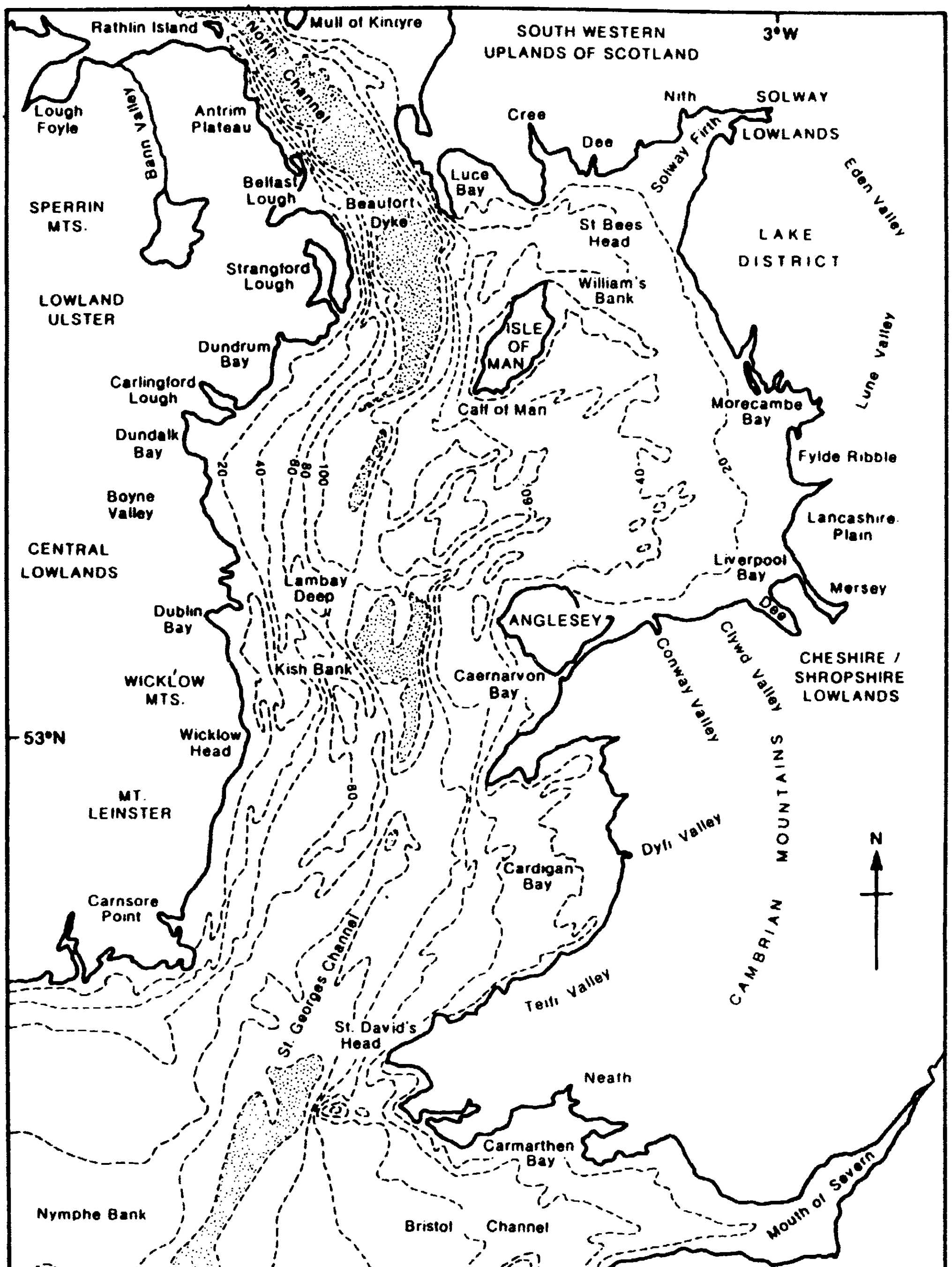


Figure 2.1 Physiographic setting and bathymetry of the Irish Sea Basin and surrounding waters. (From Eyles and McCabe, 1989).

(1) *Estuaries and coastal embayments* that have water depths ranging between high water mark and about 10m.

(2) *Inner-shelf platforms* that vary in width from 5 km to greater than 100 km and generally have gentle shelf gradients of 1:100 to 1:2000, with steeper slopes only near open coasts and where sea-floor rock outcrop creates a rugged topography. Water depths on the inner shelves normally range from 10 to 60 m.

(3) *The Western Trough*, which is the broad trough of subdued slopes of gradient 1:50 or less, running sinuously from the North Channel to St. George's Channel through the western Irish Sea. It forms a zone up to 80 km wide varying in depth from 60 to over 140 m, and forms the central part of the Celtic Trough.

(4) *Enclosed deeps* that form areas less than 5 km wide, generally up to 30 km long, and form 10 to 165 m deeper than the surrounding sea floor. They all display similar dimensions, have smooth sides and floors, are rather irregularly shaped in plan, and have side gradients less than 1:10 (Wingfield, 1989).

(5) *Rocky prominences* that are predominantly areally restricted zones of rugged topography caused by rocky outcrops at headlands, islets, and shoals; thus this category is generally associated with coastal margins

2.1.2: Pre-Quaternary geological history of the Irish Sea Basin

2.1.2.1: Physiographic setting

Large sections of the Irish Sea coastline are situated close to, and run subparallel to, main basin-margin faults or other elongate structure lines (Jackson *et al.*, 1995). Nevertheless, no particular time period can be assigned to the evolutionary structural sequence which has developed this association, although each successive structural phase is initially dependent on former events (Moseley, 1972). The solid geology outcrop pattern of the Irish Sea Basin can be broadly divided along the Anglesey - Isle of Man Ridge from the Mull of Galloway (SW Scotland), through the Isle of Man, to Anglesey (NW Wales) (Figure 2.2). To the east and west of this structural high a number of Carboniferous - Triassic and younger sedimentary basins are present. The majority of these main basins and associated structural depressions owe their siting to regional downwarping and subsequent block faulting and graben development probably during the latter stages of the Hercynian Orogeny (Dobson, 1977a).

The Solway Firth Basin, which lies immediately south of the Kirkcudbrightshire coast in the north-east Irish Sea, has been shown by gravity data to be underlain by a narrow north-east trending basin (Jackson *et al.*, 1995) (Figures 1.5, 2.2). This trough, strongly influenced by the trend of the basement rocks, extends from the northern tip of the Isle of Man towards the onshore Permo-Triassic basin of the Carlisle - Vale of Eden area. It is contiguous with the north-west trending Stranraer or Loch Ryan Basin (Figure 1.5) which extends from the Southern Uplands Fault line and is bounded on the east by the Loch Ryan Fault (Kelling and Welsh, 1970). On the coast at Loch Ryan, Permian rocks have been recognised but offshore correlations are difficult (Dobson, 1977a). Permo-Triassic rocks exposed on the Isle of Man represent an extension of the Solway Firth Basin.

The East Irish Sea Basin is a much more expansive half-graben feature by comparison with the other basins of this region and is less obviously influenced by the underlying structural framework (Jackson *et al.*, 1995) (Figure 2.2). The northern margin is formed by the Lower Palaeozoic rocks of the Southern Uplands of Scotland (Section 1.3.2) while to the east lie the Carboniferous uplands of the western Pennines and the Lower Palaeozoic uplift of the Lake District. The southern margin is formed by the Lower Palaeozoic uplift of North Wales and further west lies the Precambrian complex of Anglesey (Jackson *et al.*, 1995). Many of the intrabasinal faults have a north-west to south-east alignment subparallel to the English coastline. Geophysical evidence points to the existence of a buried Palaeozoic ridge dominated by a series of granites (Dobson, 1977a) connecting the Isle of Man (Ramsey) to the Whitehaven area at the north-west of the Lake District massif. The East Irish Sea basin is contiguous with the onshore Triassic rocks of the Liverpool - Wirral area and thence is bounded to the Cheshire Basin by a relatively shallow sill.

In the western portion of the Irish Sea Basin (Central Irish Sea and West Irish Sea Basins), PreCambrian to Carboniferous strata outcrop predominantly, although the nature and extent of the Permo-Triassic is limited (Dobson, 1977a) (Figure 2.2). The smaller Triassic North Channel and Permo-Triassic Peel and Caernarfon Bay basins occur respectively to the north and south of these older lithologies (Figure 2.2). The North Channel Basin is a complex asymmetric fault-bounded trough with a dominant north-east to south-west structural regime controlled by the major faults which frame the Midland Valley of Scotland, to the north of the Southern Uplands massif (Dobson, 1977a). Immediately east of Dublin the north-east to south-west aligned half graben Kish Bank Basin contains a sequence (> 4 km depth) of Permo-Triassic to Liassic (Lower Jurassic) age overlain by a Carboniferous succession which in turn rests unconformably on a Lower Palaeozoic basement (Dobson, 1977a; Jackson *et al.*, 1995). Major faults along the northern and south-west margins of the basin ensure a marked asymmetry for the sedimentary fill (Dobson, 1977a).

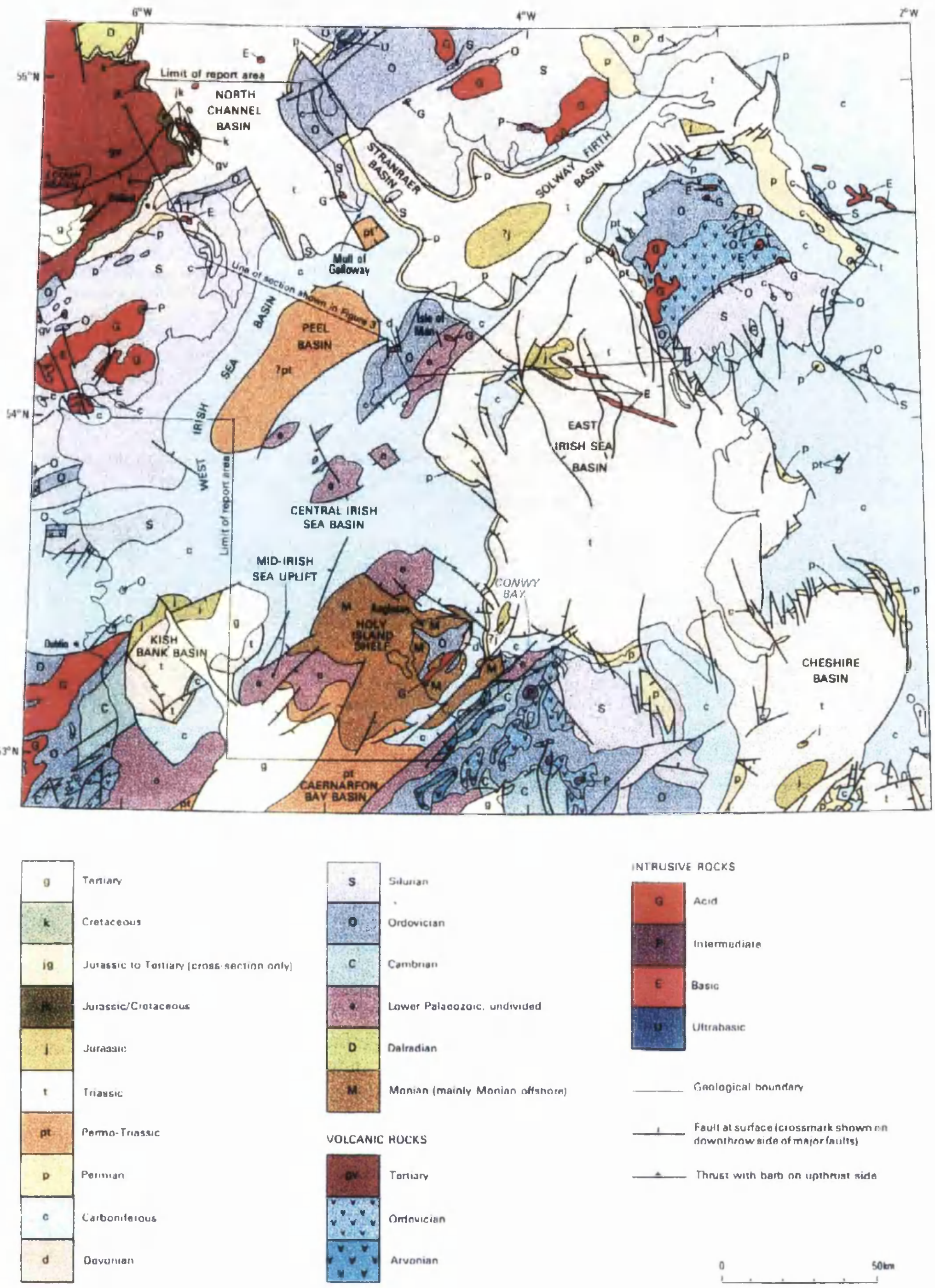


Figure 2.2 Generalised Pre-Quaternary geology of the constituent basins of the Irish Sea. Based on the 1:1 000 000 Geology of the United Kingdom, Ireland and the adjacent Continental Shelf. Information referring to the limit of report area and line of section do not relate to this investigation. (From Jackson *et al.*, 1995).

2.1.2.2: Structural and stratigraphic development

The following account attempts to trace the main sequence of events within the Irish Sea Basin, as outlined by Jackson *et al.* (1995), since the PreCambrian period. The basic regional framework responsible for the orientation and form of the basins/blocks within the Irish Sea Basin is suggested to have had a profound influence on glacial and post-glacial events, both offshore and on the seaboard margins (Dobson, 1977a). For reasons of simplification, necessary generalisations have been applied where appropriate together with periods where the stratigraphic record is incomplete as a result of limited geophysical data at depth.

PreCambrian and Lower Palaeozoic rocks occur around much of the Irish Sea in Ireland, south-west Scotland (Section 1.3.2), the Lake District, North Wales, and on the Isle of Man. The offshore limit of these rocks has been delineated mainly from seismic records whereby their structureless internal character distinguishes them from younger bedded strata (Wright *et* al., 1971). The Lower Palaeozoic succession which fringes the periphery of the Irish Sea Basin is largely composed of marine mudstones and turbiditic sandstones, with intercalated volcanic and intrusive rocks (Jackson *et al.*, 1995). The strata were deposited near the margins of the PreCambrian to Early Palaeozoic Iapetus Ocean, which extended from Scandinavia through the British Isles to Newfoundland (Soper and Hutton, 1984). As detailed in Section 1.3.2, sedimentary rocks of Ordovician to Silurian age, deposited on the northern margin of that ocean, now form most of the Southern Uplands mountain belt, which evolved in response to northward-directed subduction and uplift of the newly created continent of Laurasia in the final stages of the Caledonian Orogeny (Middle Devonian) (Leggett *et al.*, 1979; Stone *et al.*, 1987). Resulting stress relaxation in the region during the final stages of deformation was accompanied by the contemporaneous emplacement of plutonic masses of granites and granodiorites (i.e. main summits in Carrick and Galloway) (Murphy, 1987). Continued subduction closed the Iapetus Ocean and terminated this episode of basin development by Late Silurian to Early Devonian times.

By the end of the Caledonian Orogeny the newly generated landmass created by the closure of the Iapetus Ocean was subsequently uplifted and subjected to subaerial erosion (Holland, 1981). Devonian rocks are thus part of a continental facies (as opposed to marine) termed the Old Red Sandstone which extensively outcrops in parts of Britain. Devonian strata have not been recovered offshore, but seismic evidence and extrapolation of onshore geology suggest that they are likely to locally exist (Jackson *et al.*, 1995). Within the Irish Sea Basin and peripheral onshore areas however only isolated outliers of Devonian rocks are currently preserved on Anglesey, the Isle of Man, in the eastern Lake District, southern Scotland, and in County Antrim.

The Irish Sea was the site of major deposition not only from Cambrian to Devonian times, but also during the Carboniferous, Permian, and Triassic, when each episode of basin development corresponded to a time of subsidence, sedimentation and deformation (Jackson and Mulholland, 1993). There is limited evidence for other episodes of deposition, although the products are commonly no longer preserved as has been particularly highlighted by Stone (1996) in south-west Scotland. During the Early Carboniferous, a tensional stress regime was established: this gave rise to a series of grabens and half-grabens which controlled sedimentation patterns throughout most of the Carboniferous Period (Gawthorpe *et al.*, 1989).

Carboniferous strata subcrop extensively beneath Quaternary deposits in the western part of the Irish Sea, and are considered to underlie a thick Permo-Triassic cover over most of the remainder of the area (Jackson *et al.*, 1995). Onshore, they crop out in Northern Ireland, the Republic of Ireland, southern Scotland, the Lake District, the Pennines, North Wales, and on the Isle of Man. Over most of the Irish Sea, the Carboniferous rests unconformably on folded Lower Palaeozoic rocks (Jackson *et al.*, 1995). During the Carboniferous Period, the British Isles passed through equatorial latitudes and endured a hot, wet climate. In the Early Carboniferous times, the region lay to the south of the Equator, but movement related to the collision of the Eurasia and Gondwana continental plates resulted in northerly drift, so that by the beginning of the Permian, the region lay in the northern hemisphere (Johnson and Tarling, 1985). The climate changed gradually and moist, subtropical conditions gave way to a desert environment in the Late Carboniferous and Early Permian (Jackson *et al.*, 1995). Throughout such times, large parts of the major Lower Palaeozoic massifs such as the Southern Uplands in the north remained above sea level. A wide variety of facies are thus present in the Carboniferous of the region, and are thought to occur widely in the West Irish Sea Basin. Terrestrial, fluviodeltaic and marginal-marine sequences occur onshore in the fault-bounded Solway Firth Basin, indicating a range of sedimentary environments (Jackson *et al.*, 1995).

Within the Irish Sea Chadwick (1985) has applied a schematic model for the development of the Permo-Triassic Basins of the British Isles. It is suggested from the implications of this work that thick Lower Permian sediments and, locally, extrusive igneous rocks are confined to narrow grabens and half-grabens in Northern Ireland, south-west Scotland and the southern part of the East Irish Sea Basin. These centres of deposition are arranged peripherally to the centre of the East Irish Sea Basin. Seismic evidence also suggests that Permian strata are also specifically present beneath the Triassic in the North Channel, Solway Firth and Kish Bank basins (Jackson *et al.*, 1995). In the Caernarfon Bay and Central Irish Sea basins, their presence beneath younger beds is more speculative (Barr *et al.*, 1981), and they are thought to have been removed by erosion over a wide area of the West Irish Sea Basin. The Permian succession in the Irish Sea is subdivided into Lower Permian, largely continental redbeds, and Upper Permian, mainly marginal-marine and evaporitic deposits. The former continental type sediments include dune sands, alluvial fans and playa lake material and fluvial sequences. Widespread marine incursion and shallow-water conditions prominent at the beginning of the Late Permian account for the latter cycles (Jackson *et al.*, 1995).

During the earliest Permian, topography was structurally controlled by north-north-westerly and north-easterly trending active faults. Local volcanic activity was contemporaneous with the earliest sedimentation though confined to the Early Permian; lavas and pyroclastic rocks occur in both Northern Ireland, south-west Scotland and possibly in the centre of the East Irish Sea (Hardman, 1992) and south Cumbria. Around the Irish Sea, the Lower Permian is restricted to narrow grabens and half-grabens; it invariably rests unconformably on the Carboniferous, and is overlapped by Upper Permian sediments, particularly on the edges of half-grabens and at basin margins (Jackson and Mulholland, 1993). During the Permian independently subsiding grabens and half-grabens acted as isolated depocentres; particularly in the case of the East Irish Sea and Cheshire Basins. The greatest thickness of Lower Permian strata is found in the rift basins on and towards the perimeter of the Irish Sea; these deposits appear to thin towards the centre of the East Irish Sea Basin. By contrast, the Upper Permian strata show a regional increase in thickness towards the centre of this basin (Jackson *et al.*, 1995). In later Permian and Early Triassic times, the East Irish Sea Basin became linked to the Cheshire Basin creating a single depositional environment. In the North Channel Basin over 1000 m of Lower Permian sedimentary and igneous rocks have been proved, and over 1400 m of sedimentary rocks are estimated to occur in the Solway Firth Basin (Jackson *et al.*, 1995).

Variations in thickness and lithology demonstrate that the East Irish Sea and Solway Firth basins subsided independently throughout the Permian; the Solway Firth Basin successions are dominated by relatively thick aeolian deposits during the Early Permian, whereas the northern East Irish Sea Basin is characterised by thick evaporites during the Late Permian (Jackson *et al.*, 1995). Lithological evidence also suggests that the Ramsey-Whitehaven Ridge, which had been active throughout much of the Carboniferous, continued to exert an influence on sedimentation by separating the East Irish Sea Basin from basins to the north (Jackson *et al.*, 1995).

Rapid regional subsidence continued further throughout the Triassic and into the Early Jurassic, punctuated by limited uplift of the Welsh massif, and some renewed faulting. The major Lower Palaeozoic highs of the Longford-Down Massif, the Southern Uplands Massif, the Lake District Massif, the Isle of Man, the Ramsey-Whitehaven Ridge, and the Welsh Massif, continued to influence sedimentation patterns during the Triassic (Jackson *et al.*, 1995). Triassic strata are thus widespread in the Irish Sea Basin, where they lie conformably upon, and locally overlap, Late Permian rocks. At a few localities in and around the area, they are overlain by Lower Jurassic strata, but the Triassic generally subcrops Quaternary sediments (Jackson *et al.*, 1995). Subsidence continued in all the major depocentres throughout the Late Triassic and into Jurassic times, particularly throughout the Liassic allowing for thick sediment accumulation (Dobson, 1977b). Whilst the deposition was widespread present day Lias Group sediments are limited mainly to the Solway Firth Basin, the Cheshire Basin, the Kish Bank Basin, the St. George's Channel Basin, and crop out onshore in Northern Ireland (Dobson, 1977b; Jackson *et al.*, 1995). On the periphery of the offshore area, Lower Jurassic rocks are preserved in Northern Ireland beneath the protective capping of early Paleogene lavas, and small outliers of Lower Jurassic sediments are preserved in the Cheshire Basin. Offshore in the East Irish Sea Basin, proven rocks of Early Jurassic age are restricted to an erosional outlier in the Keys basin (Jackson *et al.*, 1987). Other occurrences are inferred in the Solway Firth Basin, largely on the basis of seismic interpretation and regional considerations. Jurassic sediments are possibly present in the Central Irish Sea and Caernarfon Bay basins (Jackson *et al.*, 1995).

No Cretaceous rocks are known offshore other than small outcrops from beneath lavas of Northern Ireland. It is likely that thin Upper Cretaceous limestone and chalk were deposited extensively over much of Ireland and the Irish Sea area (Wilson, 1972, 1981; Cope, 1984). Proven offshore occurrence of Cretaceous strata in the area is restricted to a small outcrop on the west side of the North Channel Basin; this is a continuation of the Northern Ireland outcrop. On the basis of the distribution of flint erratics found on the eastern coast of Anglesey, Greenly (1919) suggested that a Cretaceous outlier might occur in the East Irish Sea Basin.

Following widespread marine inundation in the Late Cretaceous, the beginning of the Tertiary was a time of considerable change in the vicinity of the British Isles. Rifting and increased thermal activity associated with the opening of the north-east Atlantic in the late Palaeocene (56 Ma - Dewey and Windley, 1988) resulted in uplift and extensive volcanic activity on the continental margins, including northern and western Britain. To the south of Britain, the northward movement of Africa towards Europe produced the Alpine orogeny, which reached its maximum during the mid-Tertiary. Compression related to this intense tectonic activity resulted in uplift towards the north-west in part of Britain and the removal of a considerable thickness of Permian to Early Triassic sediment (Jackson *et al.*, 1987; Jackson and Mulholland, 1993). The only Tertiary sediments currently identified offshore are restricted to the Central Irish Sea, Kish Bank and St. George's Channel Basins, extending to the Celtic Sea in the south (Tappin *et al.*, 1994). The Pliocene now present may, however, be only a remnant of its former thickness, the irregularities of the upper surface testifying to the effects of the Quaternary (Dobson, 1977b).

2.1.3: Pleistocene and Holocene stratigraphy of the Irish Sea Basin

The Pliocene and early Pleistocene witnessed a period of erosion and uplift across the Irish Sea region (Dobson and Whittington, 1987), following which, extensive sedimentation occurred in the Mid- to Late Pleistocene. Detail concerning the thickness and distribution of Quaternary drift on the floor of the Irish Sea Basin is limited and is derived from the results of offshore seismic and drilling programmes (Wright *et al.*, 1971; Wilkinson and Halliwell, 1979). Both Pleistocene and Holocene deposits are recognised by their striking unconformable relationship with underlying pre-Quaternary strata and by their commonly occurring sub-horizontal stratification (Jackson *et al.*, 1995).

In general, drift thickness, including both Pleistocene deposits and Holocene marine sediment, averages less than 20 m over most of the Irish Sea Basin but there are few areas of any extent where bedrock is exposed on the sea floor (Thomas, 1985). Very thick local accumulations of Quaternary sediments range up to 300 m, but even the thickest sequences include major erosion surfaces, and nowhere has deposition been continuous (Jackson *et al.*, 1995). Extensive thick deposits (>100 m) are confined within a 25 to 80 km-wide trough roughly coextensive with the deepest waters of the Western Trough. Flanking this trough to the east and west respectively are the Eastern and Irish Platforms where Quaternary deposits are generally less than 50 m thick (Jackson *et al.,* 1995). Detailed investigations in certain areas, using closely-spaced geophysical traverses, have identified very considerable variation in both drift thickness and bedrock topography. Thus in the area of the basin stretching from the Isle of Man to the Lake District drift thicknesses locally reach 120 m in the base of channels incised into underlying bedrock. It is considered by Wingfield (1989) that these major incisions are of a glacigenic origin (Section 2.3.2.2). Many incisions on the platforms are filled with sediment up to 200 m thick in localised, enclosed depressions that are less than 5 km wide and up to 40 km long (Jackson *et al.*, 1995). Some incisions, both on the platforms and in the Western trough, are incompletely filled and are marked by elongate, enclosed bathymetric deeps (Jackson *et al.*, 1995).

Borehole records (Wright *et al.*, 1971; Wilkinson and Halliwell, 1979) and acoustic profiles (Belderson, 1964; Pantin, 1979) suggest that the Quaternary succession underlying the Irish Sea is relatively uncomplicated and consists of a tripartite sequence (Figure 2.3). Although there are variations from area to area the succession generally conforms to one of an upward transition from till, to proglacial sediment, to marine sediment, each bounded separately by a marked unconformity (Pantin, 1975, 1979). The rockhead is overlain over most of the area by a more or less continuous mantle of till a few metres thick. This is relatively uniform lithologically and comprises a stiff, reddish-brown mud with erratic pebbles (Thomas, 1985). Usually there are no signs of internal stratification from acoustic readings, but a faint diffuse banding may persist locally (Pantin, 1977). In some parts of the north-eastern and western sector of the Irish Sea

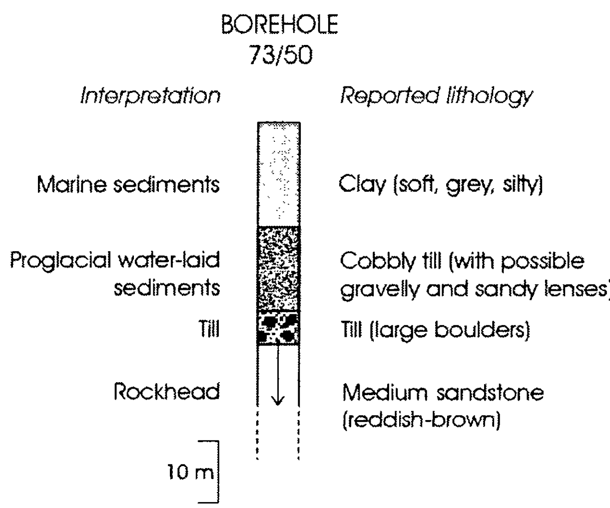


Figure 2.3 Diagram of Irish Geological Survey borehole 73/50 demonstrating a tripartite sequence of Quaternary sediment. (Modified from Pantin, 1977).

Basin the till comes within 1 m of the sea bed being directly overlain by thin marine sediments (Pantin, 1975, 1977). Till covering the rest of the region is overlain by a unit of well-bedded sediments, up to several tens of metres thick (Pantin, 1977).

Comparison of acoustic profiles, vibrocores and borehole evidence reveals that over much of the area, the unit immediately overlying the till consists of proglacial water-laid sediments, apparently deposited in a proglacial lagoon (Pantin, 1975, 1977). These sediments are characteristically muddy, with thin sand and coarse silt laminae which are frequently numerous, and occasional beds of coarser sediment up to cobble size (Thomas, 1985). Dropstones are common and thin intercalated till units, similar to iceberg dump structures or bergmounds (Thomas, 1985), also occur. Overlying the proglacial unit, and extending up to the present sea floor, is a series of marine beds* whose components (mud, sand and/or gravel) vary widely in amount depending upon horizon and locality (Pantin, 1975, 1977; Thomas, 1985). In localities where the marine beds are sufficiently thin to allow cores to reach the underlying proglacial deposits, the cores always reveal an abrupt, unconformable contact between the two units (Pantin, 1977). Where the beds are thicker, acoustic profiles generally indicate an upward transition varying from an abrupt discontinuity to a moderately rapid change-over (Pantin, 1977). The age of the marine beds is thought to range from Late Devensian through Flandrian to the present day.

Subsequent recognition of further correlative units gained from seismic profiles has since allowed Hession (1988) to divide the Quaternary deposits of the Irish Sea into six formations, as adopted by the British Geological Survey. In order of decreasing age these are: the Bardsey Loom, Caernarfon Bay, St George's Channel, Cardigan Bay, Western Irish Sea and Surface Sands Formations. These formations are seen to largely overlie one another in the Western, St. George's Channel and Celtic Deep Troughs, although nowhere are all the formations present. Lateral transition, including interdigitations, occurs between the Cardigan Bay Formation and the Western Irish Sea Formation thereby qualifying the interpretation (Jackson *et al.*, 1995). By analogy with the threefold division of the North Sea Quaternary sequence, where major incisions* (erosion surfaces) are dated as Late Elsterian, Late Saalian and Late Devensian (Cameron *et* al., 1987), Wingfield (1989) proposed that a crude event stratigraphy can be applied to the offshore Quaternary sequences of Britain. This framework, when applied to the sequences interpreted by Hession (1988) in the Irish Sea Basin produces the correlations illustrated in Figure 2.4.

BARDSEY LOOM FORMATION

This formation has been recognised as the oldest Quaternary strata and therefore is confined to the Western Trough in St George's Channel, where they are deeply buried. The deposits generally occupy shallow basins within rockhead some 3 to 10 km wide and consist of beds up to 50 m thick (Tappin *et al.*, 1994; Jackson *et al.*, 1995). No samples have been obtained from these deposits but their acoustic signature may be tentatively interpreted as rudimentary bedding within shallow channels, possibly indicating a fluviatile or shallow-marine

* The term "marine beds" is adopted here in a broad sense, and includes possible estuarine or intertidal sediments, in addition to deposits associated with a normal 'marine' environment. The term has not been used for any of the proglacial beds as it is thought inappropriate for the type of environment in which these sediments are envisaged to have been deposited.

* Major incisions are defined by Wingfield (1990) as eroded, enclosed depressions cut into the continental shelf or adjacent lowlands and attaining dimensions greater than 100 m incised depths and 2 km widths. Such major incisions have lengths of less than 30 km, widths up to 5 km and incised depth to 360 m (Wingfield, 1989).

environment (Jackson *et al.*, 1995). In the Celtic Trough, the Bardsey Loom Formation comprises beds of clay, sand, pebbly sand and gravel with layers of peat (Tappin *et al.*, 1994). The deposits have a sparse microbiota indicative of a cold environment (BGS Biostratigraphy Group), and given their position at the base of the Quaternary sequence they are tentatively assigned a pre-Elsterian age (Tappin *et al.*, 1994). This contention is further supported by amino-acid ratio determination.

CAERNARFON BAY FORMATION

This formation consists of four informal members: the Lower Unstratified, Bedded, Incision Infill and Upper Stratified members separated by a major erosion surface marking the first generation of major incisions (Tappin *et al.*, 1994; Jackson *et al.*, 1995). The Lower Unstratified deposit is up to 70 m thick, and occurs largely in the Western and Celtic Deep Troughs, but also on the platform margins in Caernarfon Bay, north-west of Holyhead. The member either rests upon the Bardsey Loom Formation with apparent conformity, or overlies a subhorizontal unconformity above pre-Quaternary rocks (Tappin *et al.*, 1994; Jackson *et al.*, 1995). The acoustic signature of this member implies that it is composed of unsorted material deposited in subglacial or ice-proximal conditions (Jackson *et al*, 1995). Since the erosion surface above this member is considered to relate to Late Elsterian glaciation (Wingfield, 1989), these sediments are believed to be of earlier Elsterian age.

The Bedded member comprises deposits up to 35 m thick in the St. George's Channel Trough and 70 m in the Celtic Deep Trough where it is more widely preserved (Tappin *et al.*, 1994). In the St. George's Trough the restricted remnants of this member either grade laterally into the Lower Unstratified member, or conformably overlie it (Tappin *et al.*, 1994). The depositional setting is unknown, but the member is truncated by the Late Elsterian erosion surface therefore immediately predates it (Tappin *et al.*, 1994). Within the Bedded member, and into the underlying older Pleistocene strata, and up to 150 m of pre-Quaternary bedrock, exist boat-shaped depressions more than 200 m thick (Tappin *et al.*, 1994). Post-dating or coeval with these eroded depressions rests the Incision Infill member which is characteristically more than 200m deep and considered to be of Elsterian age (Tappin *et al.*, 1994). The final member of the Caernarfon Bay Formation is the Upper Unstratified member which is well developed throughout the Celtic Deep and St. George's Channel troughs, as well as in central parts of Cardigan Bay on the Welsh Platform (Tappin *et al.*, 1994). In the Celtic Deep trough these deposits both underlie, and pass laterally into, the St. George's Channel Formation (Tappin *et al.*, 1994).

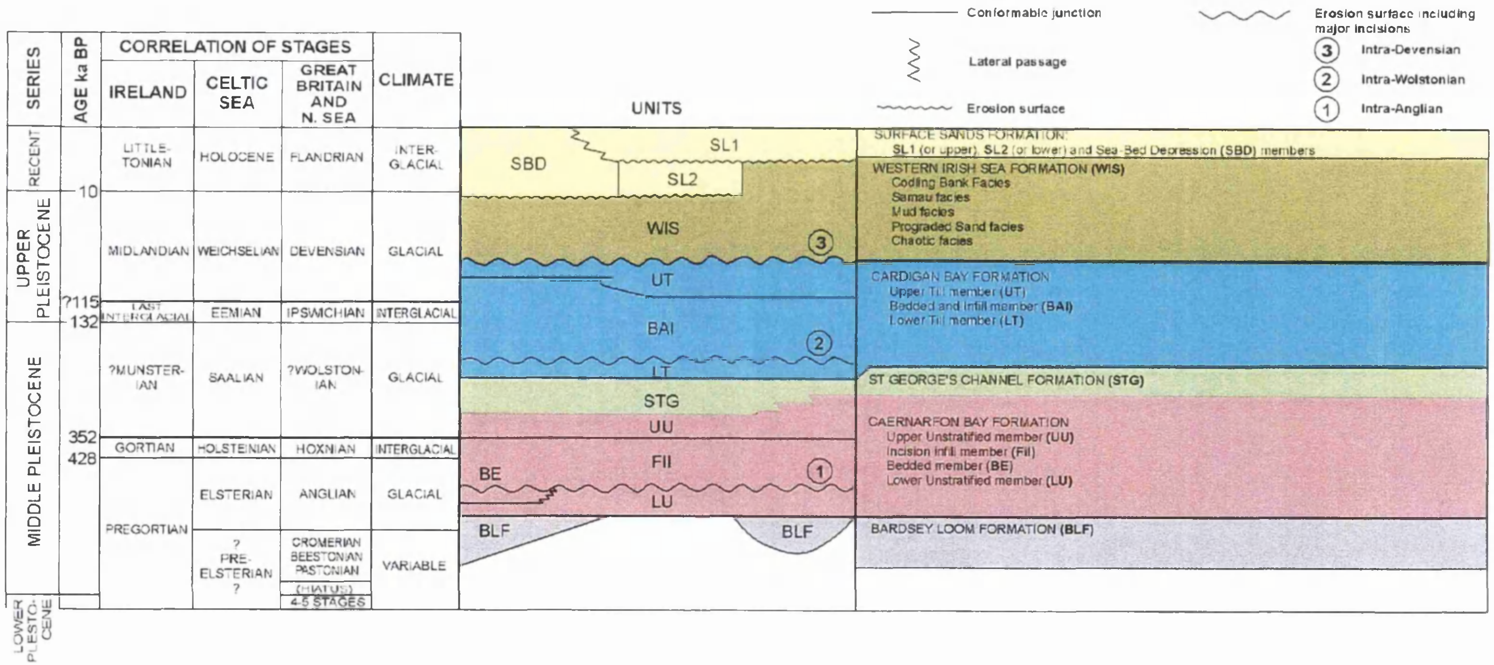


Figure 2.4 Quaternary stratigraphy in the Irish Sea Basin. (Modified from Tappin *et al.*, 1994).

ST GEORGE'S CHANNEL FORMATION

This unit occurs in the southern part of the Western Trough and exclusively within the Celtic Trough, but also extends on to the adjacent platforms, e.g. Caernarfon Bay (Tappin *et* al., 1994). It consists of deposits generally 30 to 65 m thick, however attains 125 m in the Celtic Deep. The formation has not been sampled within the main Irish Sea Basin although further south three boreholes through the St George's Channel Formation proved muds with minor shell debris and sporadic pebbles (Tappin *et al.*, 1994). Although these thick sequences have been previously described as temperate marine interglacial deposits of possible Ipswichian age (Jasin, 1976; Garrard, 1977; Whittington, 1980; Hession, 1988), it is currently disputed that the abundant macro- and microfauna are of boreal, cold waters that suggest arctic-like, glaciomarine, depositional conditions (Jasin, 1976; D.M. Gregory and R. Harland, written communications, 1985, 1987). In view of this information, and its position within the Quaternary sequence, Jackson *et* al. (1995) propose an early Saalian age for the St George's Channel Formation.

CARDIGAN BAY FORMATION

The Cardigan Bay Formation comprises Upper and Lower Till members between which lies a middle member, the Bedded and Infill member, composed of lenticular infills of major incisions (Tappin *et al.*, 1994; Jackson *et al.*, 1995). All three members can be found in sequence in the troughs alone, although the Upper Till member is widely represented on the platforms, as, to a lesser extent, is the Bedded and Infill member (Tappin *et al.*, 1994). The Upper and Lower Till units pass southward into thinner sands and gravels (Tappin *et al.*, 1994).

The Lower Till member ranges up to 90 m in thickness in the northern part of St George's Channel, where it outcrops only locally at the eastern margin of the Western Trough (Jackson *et al.*, 1995). Further south at the southern limit of the Celtic Deep Trough its thickness diminishes to less than 5 m (Tappin *et* al., 1994). The base of this member is an erosion surface with a gentle topographic variation of some 15 m in amplitude (Tappin *et al.*, 1994; Jackson *et al.*, 1995). There are only two borehole sections of the Lower Till member available. However, what has been uncovered appears to suggest that the deposit consists of subglacial lodgement till, passing southwards into ice-proximal glaciomarine gravels and sands (Tappin *et al.*, 1994; Jackson *et al.*, 1995). The member is presumed to have been deposited by Saalian ice prior to the formation of the next generation of major incisions into which the Bedded and Infill member was deposited (Jackson *et al.*, 1995).

Two distinct facies are present in the Bedded and Infill member, which do not necessarily occur together. The lower facies is at least 85 m thick in the eastern Irish Sea, and occupies erosional downcuts of various sizes, the largest of which are in excess of 75 m deep (Tappin *et al.*, 1994). Some of these incisions may however be from an earlier erosional event, in which case their infills belong to the Caernarfon Bay Formation. In detail, seismic profiles across these lenticular infills display similar acoustic characters to the Incision Infill member of the Caernarfon Bay Formation (Tappin *et* al., 1994). The lower part of the Bedded and Infill member is thought to represent the Saalian deglaciation episode, perhaps including partial infill during part of the Eemian Interglacial (Jackson *et al.*, 1995).

The Ayre Marine Silts found in boreholes on the northern part of the Isle of Man (Lamplugh, 1903; Smith, 1931) and offshore in Luce Bay probably form the uppermost part of the Bedded and Infill member in the northern half of the Irish Sea Basin. These are silts with a cold or boreal marine macrofauna overlying barren, gravelly sands. In the south, this upper portion of the member is similar to the St. George's Channel Formation comprising fine-grained silty sands and sandy clays (Tappin *et al.*, 1994). Sparse micro- and macrofauna indicating arctic-like conditions, combined with the stratigraphic position of the Bedded and Infill member, probably indicate a very late Saalian or early Eemian origin (Jackson *et al.*, 1995).

The Upper Till member crops out extensively at the sea bed in the Irish Sea, especially north of Anglesey and forms most of the sea bed of St. George's Channel and Cardigan Bay (Tappin *et al.*, 1994; Jackson *et al.*, 1995). It is found in numerous cores as a till comprising stiff or hard diamicton of clay with varying amounts of sand, gravel, shell and cobbles and boulders ranging up to 1 m in diameter (Tappin *et al.*, 1994; Jackson *et al.*, 1995). In seismic profiles, it appears to form a unit across most of the platforms and in the northern and southern parts of the Western Trough. The central part of the trough in the Manx Depression is, however, devoid of the Upper Till member, except as patches on raised topographic features (Jackson *et al.*, 1995). The member ranges from 5 to 35 m in thickness in many offshore boreholes. South of 52°N, it thins progressively as far as 51°20′N, where it is seen to wedge out (Tappin *et al.*, 1994). The Upper Till member is interpreted to be a subglacial lodgement till, the product of Late Devensian glaciation, and it is thought to be correlatable with the basal till of the 'Irish Sea Drift' onshore (Eyles and McCabe, 1989a). If this is so, it lends support to the hypothesis that the basal portion of the 'Irish Sea Drift' is subglacial in origin. It also correlates with the similarly deposited basal Devensian deposits on the Isle of Man (Eyles and McCabe, 1989a). Nearshore, both in Dublin Bay (Naylor, 1964) and Morecambe Bay (Knight, 1977), a number of boreholes reveal that the Upper Till member forms the basal Quaternary deposit, where it ranges from 2 to 56 m in thickness (Jackson *et al.*, 1995).

WESTERN IRISH SEA FORMATION

This formation crops out over much of the Irish Sea Basin and its sediments are similar in seismic character and geometry to the deposits of three previous older formations: the Caernarfon Bay, St. George's Channel and Cardigan Bay formations (Jackson *et al.*, 1995). The formation displays notable facies changes both vertically and laterally which are attributable to variations in sedimentation with time, or to sedimentary passages from proximal to distal settings in relation to sediment supply (Tappin *et* al., 1994). The deposits are locally massive in incision infills such as the Beaufort's Dyke in the North Channel although are mainly found in extensive east and west belts; otherwise they are developed in isolated patches (Tappin *et al.*, 1994).

Table 2.2 Facies types constituting the Western Irish Sea Formation with proposed depositional settings.

Facies recognised within Western Irish Sea Formation (decreasing age)	*Environmental depositional context*
Chaotic facies (dominantly gravels with muds, sands, cobbles and boulders)	Deposited during the Devensian in glaciolacustrine and glaciomarine, ice-proximal conditions.
Prograded facies (sands with subordinate muddy and pebbly parts)	Prodeltaic and glaciomarine facies, representing passage from the ice-proximal chaotic facies to a distal mud facies as Devensian ice retreated.
Mud facies (silts, with scattered pebbles and very sparse cobbles (dropstones))	A distal, glaciomarine deposit passing upwards into distal, marine deposits as the climate warmed at the end of the Devensian.
Sarnau facies (ridges covered by gravel, cobbles and boulders, and formed of clast-supported, clayey diamicton)	(1) Lateglacial median moraines of piedmont glaciers (Garrard & Dobson, 1974). (2) Remnants of Lateglacial sandur.
Codling Bank facies (forms the dissected remnants of a previously continuous mantle)	May compare with the Lateglacial, braided deposits described onshore in Leinster by Eyles & McCabe (1989a), or the succeeding diamicton they attributed to subaqueous ice-rafting.

Sources: Tappin *et al.*, 1994; Jackson *et al.*, 1995.

In the east belt, defined by the northern St George's Channel and North Channel, the Western Irish Sea Formation lies unconformably upon, and therefore entirely post-dates, the Upper Till member of the older Cardigan Bay Formation. The deposits are considered to be no older than Late Devensian, as they overlie till estimated to be of Late Devensian age (Jackson *et al.*, 1995). In the west belt, similar units are present, but they occur two-fold. There is a basal unit, the Western Irish Sea Formation B, comprising incision infills up to 100 m thick, with younger deposits up to 80 m thick (Jackson *et al.*, 1995). At the top of this lower unit there is a widespread erosion surface (unconformity) above which lies the Western Irish Sea Formation

A (Jackson *et al.*, 1995). Five facies are recognised within the Western Irish Sea Formation, each of which pertains to a particular environmental setting. From oldest to youngest these are the: Chaotic, Prograded, Mud, Sarnau, and Codling Bank facies (Table 2.2).

SURFACE SANDS FORMATION

Deposits of this formation are absent or extremely thin, generally less than 2 m thick over the greater part of the area, and commonly less than 0.5 m (Tappin *et al.*, 1994; Jackson *et al.*, 1995). However, in the nearshore and intertidal zones there are localised thicknesses in sandbanks of 20 to 40 m (Adams and Haynes, 1965; Haynes and Dobson, 1969; Knight, 1977; Pantin, 1978; Allen, 1990).

The Surface Sands Formation is divided into three morphological members: the Sea Bed Depression, SL1 and SL2 members (Tappin *et al.*, 1994; Jackson *et al.*, 1995). Although they all dominantly comprise sand the first two are in their upper parts products of present-day marine processes at the sea bed. Each member also includes sediments deposited in conditions very different from those of the present, including some shallower-water or subaerial deposits. In intertidal areas, muds make up an important, though subordinate, part of the Surface Sands Formation (Tappin *et al.*, 1994; Jackson *et al.*, 1995).

The Sea Bed Depression member comprises the partial or complete fill of hollows cut into deposits of the Western Irish Sea Formation A. Wingfield (1990) postulates that these hollows were exhumed as large kettle holes, some of which have an incomplete postglacial sediment infill, and thus still form enclosed bathymetric deeps. Elsewhere, many filled kettle holes are capped by deposits of other members of this formation, as in the Solway Firth and the east Irish Sea (Tappin *et al.*, 1994; Jackson *et al.*, 1995). Following Pantin's (1977, 1978) work in the east Irish Sea, the more extensively but thinner parts of the Surface Sands Formation are divided into two members: a lower SL (sedimentary layer) 2, and an upper SL1. The SL2 member is diachronous and comprises the deposits formed across a surface of erosion before and during the early Holocene (Lateglacial and postglacial) marine transgression after 10.2 ka BP (Tappin *et al.*, 1994; Jackson *et al.*, 1995). The SL1 member disconformably overlies an erosion surface across the SL2 member, or rests upon the Sea Bed Depression member or older strata (Tappin *et al.*, 1994; Jackson *et al.*, 1995). It represents the present-day mobile sediments, which include transitory and thin deposits, through accumulations up to 5 m thick in the east Irish Sea (Pantin, 1977, 1978), to active tidal sand ridges and giant sand waves up to 40 m thick. In the nearer-shore areas and inshore, the sediments pass landwards into intertidal sandy muds and saltmarsh organic clays (Naylor, 1964; Knight, 1977).

2.2: Pre-Devensian Stage

2.2.1: Pre-Devensian Glaciation

Of events predating the main Devensian stage of glaciation in the northern Irish Sea and peripheral margins we have very limited insight. Indeed, with the exception of a tentative chronology established for deeply buried tills lying below the northern part of the Isle of Man, which possibly date to the Anglian glaciation (the Lower Till and Lower Sand of the Basement Group; Thomas (1977)), subglacial erosion resulting from ice movement during the last glaciation appears to have left the Irish Sea Basin devoid of any pre-existing deposits (Garrard, 1977). Kerr (1982a) additionally notes a similar lack of pre-Devensian evidence in the Carrick and Galloway districts of south-west Scotland. This picture contrasts with the stratigraphic context which has emerged in the southern sector of the Irish Sea Basin where two major Irish Sea till units extensively prevail over much of the sea floor. These are separated by thick sequences of temperate marine interglacial sediments of possible Ipswichian age (starting *c.*128 000 yrs B.P.) (Garrard, 1977; Catt, 1981) (Section 2.2.2).

At Shortalstown near Wexford (south-east Ireland), Colhoun and Mitchell (1971) similarly describe an upper till ascribed to the Devensian and a lower till separated by a series of sands and beach gravels. Mitchell (1972) later connected these sands to the Ipswichian interglacial and the lower till to the Wolstonian glaciation. Garrard and Dobson (1974) also regard the upper till sheet of this tripartite sequence as being Devensian in age, and proposed limits for its extent in the vicinity of the southern entrance to St. George's Channel (Section 2.3.2). It is not possible though to date the older glacial event more accurately than the pre-Devensian (Garrard, 1977). Stratigraphically it underlies the marine sediments of Ipswichian age in the South Irish Sea although it has given rise to sediments which are similar in composition and appearance to those of the Upper Devensian unit, suggesting both ice sheets to have flowed along much the same lines and to have eroded similar underlying strata. Nevertheless, because of the absence of more conclusive data the possibility that the pre-Devensian cold stage, in fact, represents more than one older glaciation, cannot be ruled out (Garrard, 1977).

Selective erosion of a soft Tertiary basement, particularly in St. George's Channel, has influenced the accumulation of the two units of Irish Sea till. However in the east, glacial sediments are restricted to those of Devensian age (Garrard, 1977). It has been suggested by Garrard (1977) that this scenario is acceptable for the position of till of pre-Devensian age for it is characteristic of large ice sheets that, at least in their advancing stage, deposition predominates towards the outer margin and erosion in the interior. Thus, pre-Devensian deposits occur especially where they have been deposited in localised, overdeepened bedrock basins or troughs where advancing ice has failed to remove them (Thomas, 1985). Further relict glacial sediments, some showing evidence of weathering, are present only within certain restricted localities in other areas of the South Irish Sea Basin, and parts of the Celtic Sea and Bristol Channel (Garrard, 1977).

A series of surveys in the south Irish Sea by Blundell *et* al. (1971) led them to propose a sedimentary unit of Neogene age which existed beneath a depositional sequence interpreted as glacial and post-glacial. Dobson *et al.* (1973) made a suggestion that a connection existed between the Neogene deposits which exist in the south Irish Sea and an isolated occurrence of Neogene deposits in the Nymphe Bank area of the north-east Celtic Sea. However, recent investigations by Delantey and Whittington (1977) has prompted a reassessment of the Neogene deposits, based upon their interpretations of seismic profiles, borehole information and direct sampling. They argue that the deposits previously described as Neogene are in fact Pleistocene age (Delantey and Whittington, 1977). The approximate extent of these deposits in the south Irish Sea and the north-east Celtic Sea is indicated in Figure 2.5.

The pre-Devensian sedimentary and geomorphological ice-marginal features (e.g. end-moraines, kames, meltwater channels), which are universally accepted as evidence of a glacial limit, are rarely preserved, therefore it is imperative that we rely almost entirely on the stratigraphic and sedimentological evidence to delimit the extent of earlier ice sheets (Catt, 1981). This usually means assigning an ice margin to enclose all deposits thought to be of a certain age from their lithology or stratigraphic position. However, this is less reliable for the earlier phases of glaciation, for there is little or no evidence of direct glacial deposition (Catt, 1981). A number of regional syntheses of the Pleistocene stratigraphy of the Southern Irish Sea Basin as a whole have been conducted in order to ascertain the glacier limits of former glaciations.

A number of these reviews relate directly to the Pleistocene stratigraphy of the Isles of Scilly which lies some 40 km west-south-west of Land's End (Scourse, 1986). The first comprehensive account of a possible pre-Devensian ice limit was published by Mitchell and Orme (1967) who revised the original findings of Barrow (1906). Their interpretation of the stratigraphy on the Scilly Isles suggested that the widespread occurrence of outwash gravels formed the basis for an ice limit on Scilly. By mapping the distribution of the till and outwash gravel, Mitchell and Orme (1967) proposed an ice limit running through the northern islands in the group (Figure 2.6). South of this line, the glacially-derived sediments were not found. By comparing the basic stratigraphic sequence that they had established on the Scilly Isles with

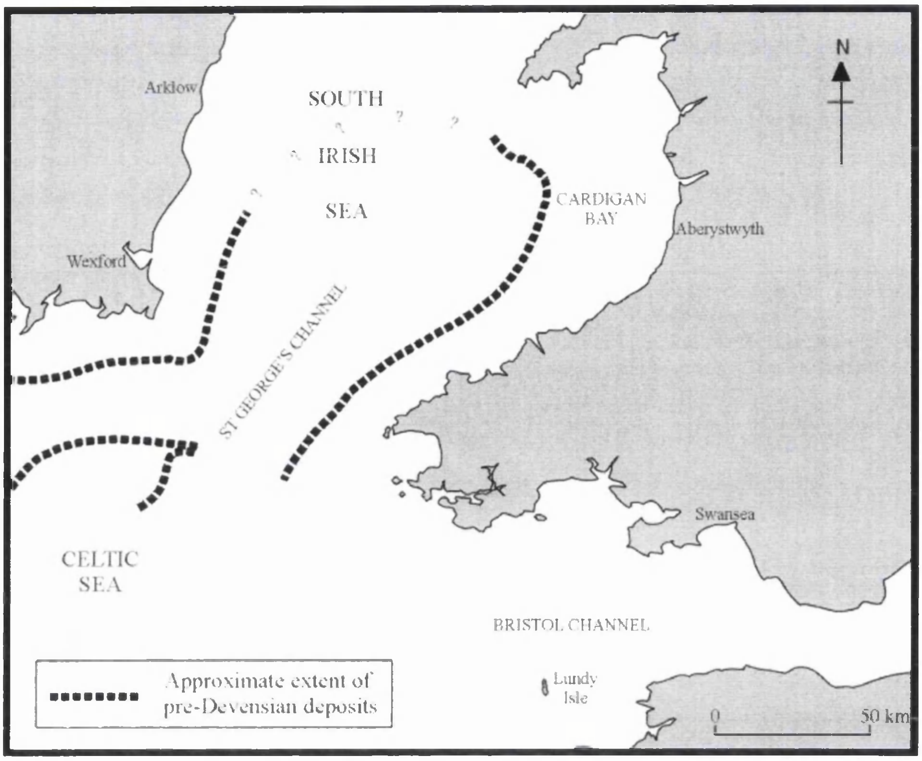


Figure 2.5 Approximate extent of pre-Devensian deposits as suggested by Delantey and Whittington (1977). Direction of Wolstonian ice movements is considered to have been from the north-east. (Modified from Delantey and Whittington, 1977).

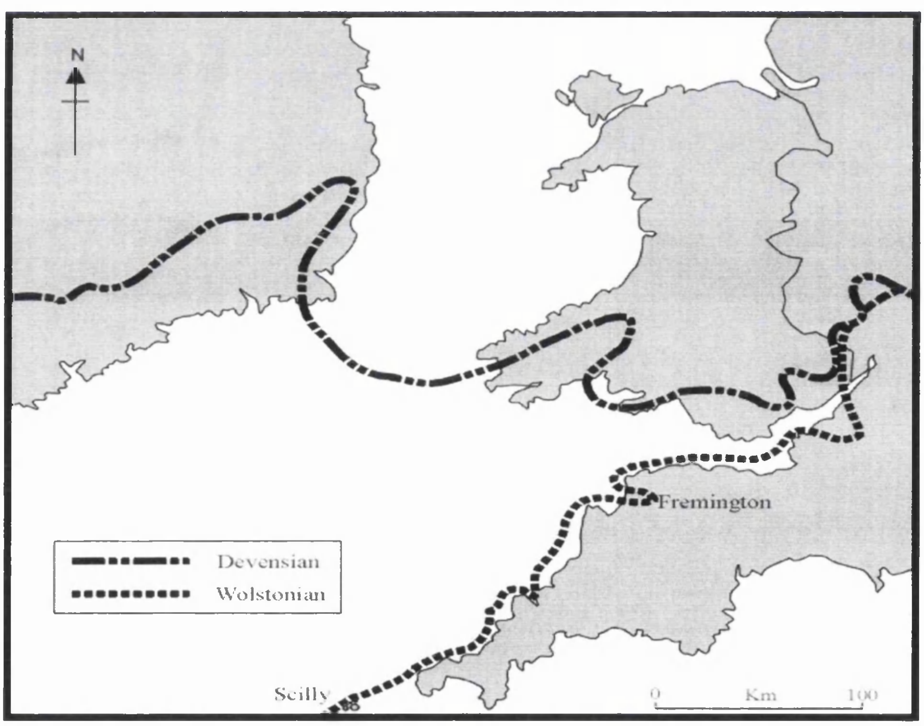


Figure 2.6 Postulated glacial limits in the southern Irish Sea Basin. Wolstonian limit derived mainly from Mitchell and Orme (1967) and Mitchell (1972). Devensian based on Charlesworth (1926a, 1928, 1929. 1966), Dwerryhouse (1923), Farrington (1944, 1954), Synge (1968), Bowen (1973b, 1981, 1986). (Modified from Scourse, 1986).

similar sections in south-west England and in Ireland, they claimed a Wolstonian age for the glaciation responsible for the deposition of the glacial material. A further basis for this dating was not the outwash deposit itself, but the age of the underlying 'main' raised beach in Scilly and West Cornwall which has been interpreted as Hoxnian (Mitchell and Orme, 1967). Wherever glacial deposits are observed overlying the raised beach they have been regarded as Wolstonian (if they lie outside the proposed extent of the Devensian glaciation). A Wolstonian age has been similarly assumed for the pre-Devensian glacial deposits left by Irish Sea ice on Lundy island (Mitchell, 1968) and various coastal areas bordering the Bristol Channel (Stephens, 1966; Edmonds, 1972; Bowen, 1977; Gilbertson and Hawkins, 1978).

A Munsterian (= Wolstonian) Irish Sea glacier moving westwards was also regarded as responsible for deposition of the shelly Ballycroneen Till, which overlies the 8 m raised beach at several sites on the south-east coast of Ireland (Wright and Muff, 1904; Mitchell *et al.*, 1973). The beach has been correlated with the Gortian Interglacial, which may be equated with the Hoxnian in England (Mitchell *et* al., 1973). It therefore emerges that the 'main raised beach' is a crucial stratigraphic marker (Bowen, 1981), especially as there is a lack of recognition of Ipswichian interglacial high sea levels in the current onshore stratigraphy above the proposed Wolstonian units. The potential is present, therefore, for raised beach deposits to be used as an interglacial datum for the lithostratigraphic successions that occur above and below them (Bowen, 1973a).

This approach cannot, however, be adopted for the northern half of the Irish Sea Basin for the coastlines of the Southern Uplands, the Lake District, North Wales east of the Llêyn peninsula, and the whole of the Irish coast north of County Wicklow, are devoid of raised beaches overlain by glacigenic deposits (Thomas, 1985). However in the southern Irish Sea Basin, at least, raised beach deposits are widespread around the margins (Thomas, 1985). Kidson (1977) infers that the parallelism in height of these raised beaches implies structural stability and has been taken to indicate the return of successive interglacial seas to levels only marginally different from those present today. Intrinsically the raised beaches cannot be dated, but as they are everywhere overlain by cold climate (including glacial) deposits which reflect one continuous cold period, hence the Devensian, it follows that the beach deposits are most logically related to the last (Ipswichian) interglacial (Bowen, 1973a). The fragmentary evidence present in the northern basin suggests, though, that the last interglacial marine datum is currently below sea level. At Ballure in the Isle of Man, for example, the only fossil beach deposits of suggested interglacial age have been detected through seismic investigations at a depth of -4 m O.D. (Thomas, 1985).

Kidson (1977) and Dobson (1977b) have respectively provided two explanations that could account for the apparent failure of the Ipswichian marine datum in the north to attain the same relative level as in the southern Irish Sea. First, although it is generally accepted that isostatic recovery from the Devensian glaciation is complete, there is very little hard evidence to confirm or deny this (Thomas, 1985). Therefore, the question of whether the depressed sea floor in the northern Irish Sea Basin has established equilibrium remains speculative (Kidson, 1977). The second explanation draws attention to the question of tectonic movement during the Quaternary along the boundary faults that have initiated the graben-structure of the Irish Sea Basin. Although the Pliocene-Pleistocene is thought to have been a tectonically quiet period in western Britain, loading by ice throughout successive glaciations could feasibly have accentuated or further promoted such movement. This would have inflicted unsuspected change upon the vertical transition of sea level between the Wolstonian and Devensian glaciations (Kidson, 1977).

Despite the recognised importance of the raised beach-tied stratigraphies, Scourse (1985) independently mapped the existing sections on the sixteen largest islands in the Scillies in order to establish a local lithostratigraphic framework. Twenty-nine absolute carbon-14 dates and two pre-existing thermoluminescence (TL) dates (Wintle, 1981) helped to provide an accurate chronology of events while inter-site correlations were strengthened by detailed palynology (Scourse, 1985). Scourse (1985, 1986) defines two lithostratigraphic models for the 'southern' Scillies and 'northern' Scillies; these areas can be regarded as 'extra-glacial' and 'glacial' respectively. If the evidence presented by Scourse (1985) is accurate it proposes that a thin lobate 'surge' of ice extended beyond the 'stable' margin of the Irish Sea ice sheet in St. George's Channel (Figure 2.6) resulting in glaciation of the northern part of the Isles of Scilly. This occurred during the Dimlington Stadial of the Late Devensian substage around 18 600 ± 3 700 (QTL 1d and 1f) yrs B.P. (Wintle, 1981) (Section 2.3.2.2). Such a working hypothesis of Devensian glaciation is not new; John (1970) and Synge (1977, 1981) both having raised the possibility in speculative discussion. The evidence of a Devensian glaciation used by Scourse (1985) clearly conflicts in a number of important respects with Mitchell and Orme's (1967) sequence of events. The criteria on which this model of the Devensian glaciation of the Isles of Scilly is based is, however, far from unequivocal, and the resultant interpretations are open to further modification (Scourse, 1985).

2.2.2: Pre-Devensian Interglacial

The record of pre-Devensian interglacial sediments is equally as sparse (Thomas, 1985). Aside from the great thickness of Ipswichian marine sediment described by Garrard (1977) no other comparable deposits have been discovered beneath tills in boreholes. The deduction seems again valid that these too have been denudated by Devensian glacial erosion. An indication of the extent of this erosion within the Irish Sea Basin is the widespread occurrence of an erratic shell fauna which is thought to partly derive itself from the reworking of pre-glacial Neogene deposits (Thomas, 1985). However, much of its presence is more probably owed to the destruction of interglacial sea-floor sediment which was covered by Devensian ice and incorporated into the basal till (Thomas, 1985).

Garrard (1977) identified the thick sequences of Ipswichian marine sediments underlying the Devensian Irish Sea till sheet in six boreholes in St. George's Channel. The thickness of the unit varied between 3 m and at least 56 m at depths below the sea-floor of between 9 m and 41 m. Where the base of these sediments was penetrated during coring the material was found to be composed of either bedrock or a lower Irish Sea till. Continuous seismic profiling across the investigation area revealed a conformable sequence of horizontally stratified facies resting upon a lower non-stratified unit. The stratified section, which is indicative of marine deposition, was laterally continuous with the pre-Devensian marine sediments seen in the boreholes (Garrard, 1977). Overlying the marine sediments was a further non-stratified unit which was believed to be Devensian Irish Sea till. The succession identified by a seismic (sparker) profile correlates with a borehole taken 8 km to the north-east of the survey line (Garrard, 1977):

<u>*Sparker Profile*</u> Lat. 52° 33.16′N Long. 04° 56.6′W		*Velocity*	<u>*Borehole 73/43*</u> Lat. 52° 33.68′N Long. 04° 51.07′W	
Water depth	65 m	1.5 km/s	Water depth	60.0 m
Non-stratified unit	41 m	1.8 km/s	Devensian Upper till	40.0 m
Stratified unit	24 m	1.8 km/s	Ipswichian Marine formation	26.5 m
Non-stratified unit	31 m	1.8 km/s	Pre-Devensian Lower till	10.5 m+
	161 m			137.0 m

The Ipswichian marine sediments are predominantly arenaceous, consist of a bimodal mixture of sands and gravels and contain thin layers of silt and clay (Garrard, 1977). Detailed particle-size analysis identified similarities to recent marine sediments on the present day floor of the Irish Sea, and also to those of known marine interglacial age exposed onshore (Garrard, 1977). Likewise, microfaunal analysis revealed a number of similarities to Ipswichian assemblages from sites in both Ireland and the United Kingdom (Haynes *pers. comm. to* Garrard, 1977).

Around the periphery of the Irish Sea Basin a number of current coastal sections reveal organic sediments beneath till deposits which have been provisionally assigned to an Ipswichian interglacial. At Wigton in the Solway Lowland (Eastwood *et al.*, 1968) a borehole found till and gravel overlying a clay deposit containing marine fauna, while in Lower Furness up to 8 m of organic sediment was identified by Kendall (1881) to lie between two separate tills. Pollen from organic silts beneath till in Scandal Beck, Cumbria also provides an Ipswichian age (Carter, unpublished) with carbon-14 dating providing minimum ages of 32 500 and 42 000 yrs B.P. (Shotton and Williams, 1971, 1973). The Ipswichian *Patella* beach of George (1932) has also been identified along the south coast of Gower, south Wales as a reliable marker horizon for the interglacial phase. The *Patella* beach formation lies on shore platforms ranging in height from just over 15 m O.D. to approximately level with the current mean high-water mark (3.6 m), but most commonly on a platform at 10 m O.D. This represents a former storm beach which characteristically consists of a resistant conglomerate, cemented by carbonate of lime with prolific local molluscan fauna (Bowen, 1977).

Table 2.3 Comparison of pre-Devensian sequence of events.

Bowen (1973)		*Other views**	
Event	**Status and (correlation)**	**Event**	**Status and (correlation)**
Beach formation	Interglacial (last/Ipswichian)	Beach formation	Interglacial (last/Ipswichian)
Complete glaciation	Glacial (indeterminate)	Complete glaciation Head formation	Glacial (Wolstonian)
		Beach formation	Interglacial (Gortian/Hoxnian)
		Glaciation of unknown extent	Glacial (Anglian)

Source: Bowen, 1973a. *Notes*: *This column is based, in various parts, on the views of: Mitchell, 1960, 1972; Synge, 1970; Mitchell and Orme, 1967; Stephens, 1966; and Watson and Watson, 1967.

2.2.3: Sequence of events

The most recent sequence of events proposed throughout the pre-Devensian (Bowen, 1973) reflects a more concise and simpler chronology than has previously been put forward in recent years (e.g. Mitchell, 1960, 1972) (Table 2.3). The first major event, represented by deposits correlated throughout the southern Irish Sea Basin, is the interglacial raised beach. The deposition of the pre-Devensian marine sediments is thought to have been in response to a major temperate episode of considerable duration. The sediment characteristics of the marine deposits are not those of glaciofluvial origin while an interstadial origin seems unlikely in view of their thickness and temperate fauna (Bowen, 1977). This and other evidence would therefore suggest a last (Ipswichian) interglacial age. This episode was preceded by complete glaciation of indeterminate detail and age. The raised beach event, terminated by deteriorating climate revealed by the palaeobotany at Newtown and Marros, was followed by a sheet-washing phase when colluvial silts were deposited prior to the onset of the Devensian period (Bowen, 1977).

2.3: Devensian Stage

2.3.1: Early and Middle Devensian Glaciation

The type site for the Devensian Glacial Stage in the British Isles is located at Four Ashes in Staffordshire, England where there are depositional units of Late, Middle and Early Devensian above an Ipswichian interglacial horizon (Shotton, 1977). With respect to these deposits, though, existing evidence for the presence of glacier ice is to be found only from the Late Devensian (Bowen *et al.*, 1986), a scenario that is typical of Devensian glacial evidence in south-west Scotland and other peripheral areas of the Irish Sea Basin. Indeed it has been generally accepted that in Britain, at least, major expansion of ice from mountainous dispersal centres such as the Southern Uplands into low-lying depositional settings did not occur on a major scale until after 25 000 yrs B.P. (Thomas, 1985). It corresponds that limited unequivocal detail has been established concerning the chronology of events or nature of depositional environments in the Irish Sea Basin prior to this expansion, except that there was a downturn into cold-climate conditions about 80 000 yrs B.P. (Thomas, 1985). Further to this it is apparent that the views of traditional workers (e.g. Wirtz, 1953; Mitchell, 1960, 1972) can now be considered as largely illusory in light of the fact that they were mostly based upon scant evidence which, on some occasions, conveniently supported an unreasoned opinion of Devensian glaciations.

Within the British Isles, an Early Devensian glaciation can only be demonstrated in Ulster where till units occur containing marine molluscs of oxygen isotope sub-stage 5e as the youngest faunal element (Bowen *et al.*, 1986). Elsewhere around the coast of Britain this is found not to be the case as glacial deposits contain fauna ascribed to both oxygen isotope sub-stage 5e and stage 3 of the Middle Devensian (Bowen, McCabe, Sykes and Harkness, *unpublished data*). Given an Early Devensian ice advance of unknown extent in Ireland, it is also thought to be unlikely that the mountains of Scotland, Wales and western and upland England escaped glaciation (Bowen *et al.*, 1986). A recurring theme, however, is that these areas have had any evidence of previous glaciations removed by Late Devensian ice.

Contrary to this, Sutherland (1981) suggested that the 'high level marine clays' known at a number of localities within a few kilometres of the present coastline of Scotland accumulated on a glacio-isostatically depressed crust during an Early Devensian expansion of the Scottish ice sheet. The implication from this was that the partial build-up of Scottish ice, perhaps because of its situation on the Atlantic seaboard and its small size, responded more rapidly in relative terms to the onset of the glacial phase than did the Laurentide or Scandinavian ice sheets (Sutherland, 1981). Amino acid D/L ratios from molluscs in these beds indicated an approximate age equivalent to oxygen isotope stage 5. It is possible that some of these were late stage 5 or earliest stage 4 in age, thus adding support to Sutherland's proposal of Early Devensian glaciation in Scotland. Furthermore the oxygen isotope curve of Shackleton and Opdyke (1973) showed a major build-up of world glaciers around 75 000 yrs B.P. with no subsequent period of deglaciation, and Ruddiman *et al.* (1980) also gave details of a major glacier build-up at *c.* 75 000 yrs B.P. Whether the sea penetrated the northern Irish Sea Basin during this episode is unclear for any record in the region has most likely been removed by subsequent glaciation. Sutherland (1981) implies from the model presented that marine transgression must have occurred prior to and consequent upon the loading of the crust by the Scottish ice (Section 2.8). From this, a scenario may be envisaged of both a fluctuating ice cover over the northern Irish Sea Basin and an isostatically raised sea level for much of the time interval between this and major ice expansion marking the onset of Late Devensian glaciation.

2.3.2: Late Devensian Glaciation

Abundant glacial sediments in the Irish Sea Basin have been unable to provide an exact depositional age although the extent, faunal content and sedimentary characteristics of the Irish Sea Drift all suggest it to be a product of the last (Late Devensian) ice advance in this region (Garrard, 1977). In the west, the till sheet overlies the temperate marine sediments considered to be of prior Ipswichian interglacial age. Meanwhile, the same glacial formation has been correlated in the eastern sector of the Irish Sea Basin with an onshore succession dated by several workers as Devensian (John, 1970; Bowen, 1973a, b). It has been considered, however, that during the deposition of an extensive till cover there was most probably some degree of sediment recycling of older glacial deposits (Garrard, 1977). The Irish Sea Basin is envisaged, therefore, to have been occupied at this time by a large ice stream (the Irish Sea Glacier) which was fed by converging ice streams radiating from ice dispersal centres in Scotland, Ireland, Wales and Northern England (Wright, 1937; Mitchell, 1972; Bowen, 1973b, 1974) (Figure 2.7; Section 2.4.2).

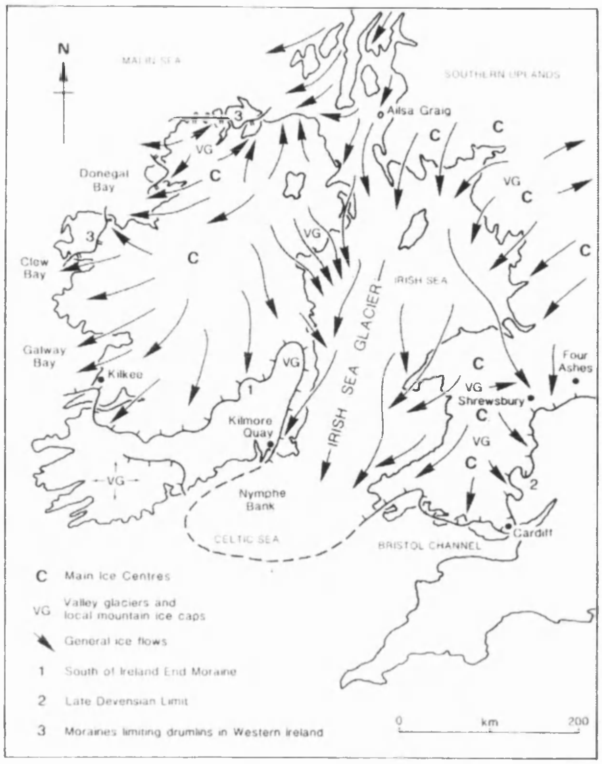


Figure 2.7 Late Devensian ice limits, ice dispersal centres and generalised flow lines in western Britain. Based on Charlesworth (1926a, 1928, 1929, 1966), Dwerryhouse (1923), Farrington (1944, 1954), Synge (1968), Bowen (1973b, 1981, 1986). The maximum extent of the Late Devensian ice along the southern approaches to St. George's Channel is not accurately known (See Section 2.3.2.2). (From Eyles and McCabe, 1989a).

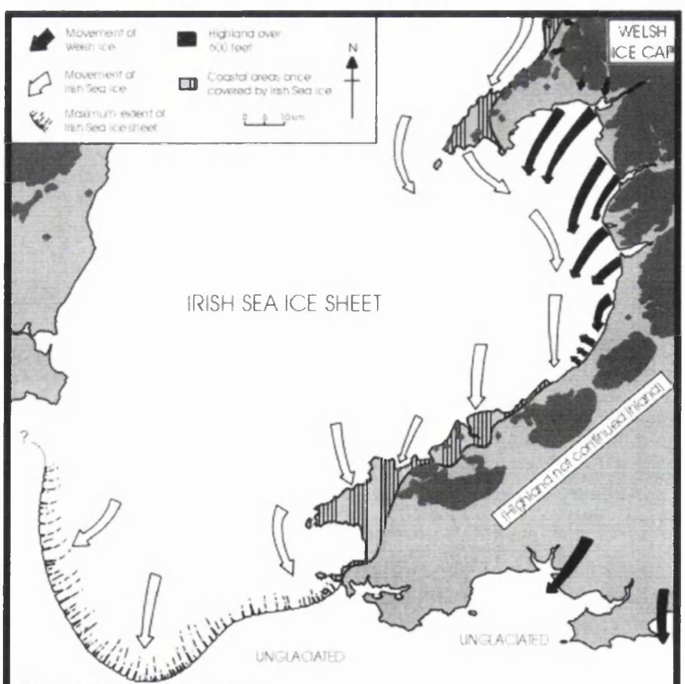


Figure 2.8 Final position of the Irish Sea ice during times of maximum Devensian glaciation, as proposed by Garrard (1977) on the basis of Devensian till coverage. (Modified from Garrard, 1977).

2.3.2.1: Late Devensian offshore stratigraphy

The term *Irish Sea Drift*, like its counterpart in eastern Britain, the *North Sea Drifts*, has been coined in the literature to describe what is known to be sedimentologically complex diamicton, mud, sand and gravel sequences along the margins of the Irish Sea Basin (Wright, 1937). Synonymous to this, the term *Irish Sea Till* is traditionally used as an omnibus to refer to any massive, fine-grained, laminated mud with varying clast contents and a variable content of foraminifera and molluscs (McCabe, 1987; Eyles and McCabe, 1991). These deposits are present across a large area of the offshore and are widely exposed around the Irish Sea coast and adjacent areas (Eyles and McCabe, 1989a). Fluvioglacial sediments are also common, especially in the South Irish Sea, but are never abundant (Garrard, 1977). Traditional work conventionally viewed the Irish Sea tills as 'basal tills' which were associated with terrestrially-based ice sheet models. Indeed Garrard (1977) suggested that lodgement till was the commonest and most widely distributed of the glacial sediments in the South Irish Sea. The complex nature of these deposits was attributed to 'advance and retreat' ice sheet cycles (McCabe, 1987). Emphasis has often been placed, in the past, on stratigraphy rather than detailed facies analysis and inadequate field methodology resulting in the splitting up of a natural depositional system into individual lithological units which were undeservedly provided with a formal status (e.g. Mitchell *et al.*, 1973; Colhoun, 1971, 1973; Davies and Stephens, 1978; Warren, 1985). Modern interpretations are tending towards the recognition of such deposits as heterogeneous assemblages of glacial and glacially-influenced marine sediments deposited in proximal and distal settings around retreating tidewater ice margins in varying water depths (Eyles and McCabe, 1991) (Section 2.5). In doing so, this has allowed greater definition of the larger depositional systems within such glaciated basins as the Irish Sea Basin.

One of the most diagnostic features of the Irish Sea till can be attributed to a high carbonate content of between 12% and 27% of the total sample weight. The carbonaceous constituent has been proven to exist as a fine-grained rock flour within the clay fraction which possibly represents the eroded remains of carbonate rocks, especially Carboniferous and Middle Jurassic limestones, and Permo-Triassic marls (Garrard, 1977). Other distinguishing characteristics of the till include a high degree of textural maturity and either a dark yellowish brown (10 Yr 4/2) or olive grey (5 Y 5/1) colour. The olive grey coloured till is believed, however, to be more greatly concentrated in the eastern sector of the Irish Sea Basin (Garrard, 1977). Textural information obtained by Garrard (1977) from the till at different depths around the Irish Sea Basin was found to produce a consistent range of readings. The fine-grained clay matrix constitutes between 70% and 90% by weight of the total sediments and may contain up to 20% sands and shell fragments (Garrard, 1977). Erratics have been found although are unusually small and appear to form an open framework within the clay matrix (Garrard, 1977). The abundance of Upper Palaeozoic, Mesozoic and Tertiary erratics which were collected are envisaged to have been scoured from the floor of the Irish Sea (Wright *et al.*, 1971; Dobson *et al.*, 1973). It has been inferred from this that there was most likely extensive continental glaciation from ice of northern origin (Garrard, 1977).

The thickness of Irish Sea glacial sediments, mainly recorded from borehole information in the South Irish Sea, varies between 5 m and 50 m and average 30 m to 40 m, whereas seismic profiles suggest they may locally reach 70 m (Garrard, 1977). As the till sheet progresses southwards it becomes progressively thinner (10 m off the West Pembrokeshire coast), where it eventually forms a narrow lobe of glacial deposits which protrude into the northern part of the Celtic Sea (Garrard, 1977).

2.3.2.2: Offshore extent of Late Devensian Irish Sea Ice Sheet

The maximum offshore Late Devensian limit of the Irish Sea Glacier at the southern end of the Irish Sea Basin is not well defined. Although there has, in the past, been some considerable dispute concerning the maximum limits attained by Late Devensian Irish Sea ice, a consensus view now concludes that it extended at least as far south as a line from County Wexford, Ireland to Pembrokeshire, Wales (Bowen, 1973a; Eyles and McCabe, 1989a; Scourse *et al.*, 1991) (Figure 2.6, 2.7). Former interpretations considering a Devensian extent were largely based upon the offshore interpretation of Garrard and Dobson (1974), as outlined in Section 2.2.1. They regarded the upper till sheet of a tripartite sequence as being of Devensian age, and thus formulated a southerly limit of this till sheet in the vicinity to St. George's Channel. Garrard (1977) later identified definite limits for the upper of the Irish Sea till sheets (Devensian) in the St. George's Channel area. From this, Garrard (1977) has provisionally related the final termination of the Irish Sea till sheet, due west of St. Govan's Head at a latitude of 51°34′N, to a theoretical ice limit (Figure 2.8). Bowen *et* al. (1986) have since provided an inferred Late Devensian ice limit at an approximate latitude of 51°30′N. South of these limits Garrard (1977) has noted only small and isolated outliers of glacial drift found as far south as the Bristol Channel. These consist of 'lag gravels'; small patches of erratic pebbles and cobbles with the fines winnowed out by strong tidal currents. The rate of sediment thinning and the sedimentary character of the till was thought to indicate that the Devensian aged deposits were unlikely to have been deposited much further south than such proposed limits (Garrard, 1977).

Pantin and Evans (1984) have reported from side-scan sonar occasional small 'mounds' of possible glacigenic material and scattered boulders on the floor of the Central and South-western Celtic Sea, which they see as reflecting individual iceberg 'dumps'. The occurrence of scattered boulders has also been noted by Hamilton *et al.* (1980). These sediments and morphological features occur between 300 and 500 km to the south-west of the suggested limits of Devensian material in the St. George's Channel (Scourse *et al.*, 1991). Beyond the defined continental shelf in this area there is thought to be further evidence for glaciomarine sedimentation in the deeper water (Scourse *et al.*, 1991) (Figure 2.9). Day (1959) also reported possible glaciomarine sediments at 48°39′N 10°35′W in 1419 m of water on the continental slope. Based on these tentative reports Scourse *et al.* (1991) conducted detailed analysis of 'glacigenic sediments' in the southern region of the basin following vibrocore sampling by the Continental Shelf Division of the Irish Geological Survey. They inferred that at around 49°30′N a change in depositional environment did occur, either a grounding line representing a transition from grounded to floating ice, or simply a change from proximal to distal glaciomarine settings. This transition occurred between -127 and -145 m OD. The most probable hypothesis deemed acceptable by Scourse *et al.* (1991) was for these sediments to have been deposited from rafted ice.

A proposed regional reconstruction of glacial events has been outlined by Scourse *et al.* (1991) in which offshore glacigenic sediments have been tentatively correlated with the Devensian till from the Scilly Isles. The basis for this association is that the Scilly till resembles the offshore glacigenic samples both lithologically (Scourse, 1985) and mineralogically (Catt, 1986). In this model, global stadial sea level is believed to have stood somewhere between 100 and 50 m below the present, with a post-rebound shoreline below present sea level (Scourse *et al.*, 1991). This limit is said to be consistent with the lack of a raised shoreline on the Scilly Isles or in Cornwall. It is perceived that such a shoreline could only have been generated by ice if it obtained an overall thickness greater than 250 m, in which case the Scilly Isles would most probably have been overridden anyway (Scourse *et al.*, 1991). Recent British Geological Survey coring and seismic evidence in the region between the continental shelf in the Celtic Sea and north of 50°N currently supports the original idea of Garrard (1977) for an ice lobe having advanced from the north-east, beyond the stable ice margin of the Irish Sea ice sheet (Scourse *et al.*, 1991). This inferred lobe is predicted therefore to have originated from immediately south of St. George's Channel from the Celtic Deep Trough (Figure 2.9). This theory is also concurrent with that of Synge (1981) who suggested that Late Devensian ice expanded westwards from St. George's Channel as a great lobe of floating shelf-ice that reached the Devon coast and the Scilly Isles.

Further evidence for an ice limit may be attained by mapping the geographical distribution of major glacial incisions (Wingfield, 1989). Current research regarding the origin of major incisions (Boulton and Hindmarsh, 1987; Boyd *et al.*, 1988; Wingfield, 1990), and lesser features of similar form (Mooers, 1989), suggests that these major incisions form under or at the margin of ice sheets although details as to how remain confused*. Thus the finding of a major incision may be taken to indicate the former presence of an ice sheet. Late Devensian incisions are known to be relatively easy to distinguish in that they lack cover deposits and their infills pass upwards to form the current seabed (Wingfield, 1989, 1990). In a majority of incisions the infill is incomplete and enclosed deeps prevail (Wingfield, 1989). Numerous Late Devensian major incisions and associated enclosed deeps between Britain and Ireland occur in specific localised areas spanning from the North Channel to the Celtic Deep (Wingfield, 1989). A proposed south-west outer limit to ice in St. George's Channel, the St. George's Lobe, occurs at 51°N (Wingfield, 1989) which is some 100 km north of the ice advance suggested to have occasionally occurred by Pantin and Evans (1984) and Scourse (1986), though more concurrent with the findings of Garrard (1977) and Bowen *et al.* (1986).

2.3.2.3: Onshore western extent - Southern Irish End Moraine

Onshore in the south of Ireland the maximum extent of the ice sheet is marked by a discontinuous moraine that occurs along the southern edge of the limestone lowlands allowing a stronger inference about the ice limit to be made (McCabe, 1987; Hoare, 1991). Specifically, the moraine complex can be traced north-eastwards from Counties Limerick and Tipperary to the western and northern slopes of the Wicklow mountains (Farrington, 1957a, b; Synge, 1973, 1981; Hoare, 1975). This feature was first discovered by Lewis (1894) and subsequently mapped in detail by Charlesworth (1928) from which point is was then referred to as the Southern Irish End Moraine (S.I.E.M.) (Figure 2.7). Further work by Farrington (1934) suggested that the last ice sheet from the Devensian did in fact surround the Wicklow mountains (south-east Ireland) to a height of 150-230 m. However it was later recognised that the stratigraphy examined by Farrington (1934) was more complicated than formerly realised. Synge (1973) used further evidence which included the concentration of granite boulders, arcuate lakes, outwash terraces and erratic distribution to reconstruct a comprehensive and decidedly more complex pattern of ice limits in this area than was first adjudged.

* Boulton and Hindmarsh (1987), Boyd *et al.* (1988) and Mooers (1989) consider that major glacial incisions form subglacially by the piping of sediment under high pressure water flow. To the contrary Wingfield (1990) proposes that each incision was created separately at a lowland or tidewater ice sheet margin as a jökulhlaup plunge pool, which cut back, and unroofed, into the ice sheet margin.

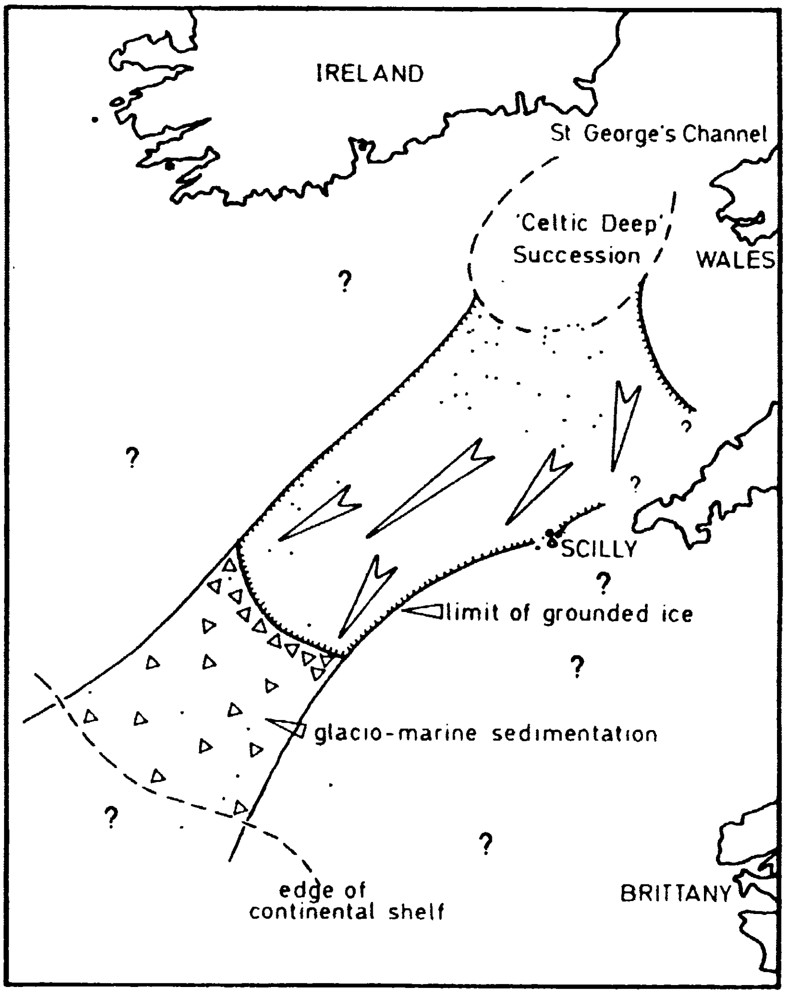


Figure 2.9 A reconstruction of the Celtic Sea ice lobe and glaciomarine terminus. Dots represent vibrocore sampling sites yielding glacigenic sediment. (From Scourse *et al.*, 1991).

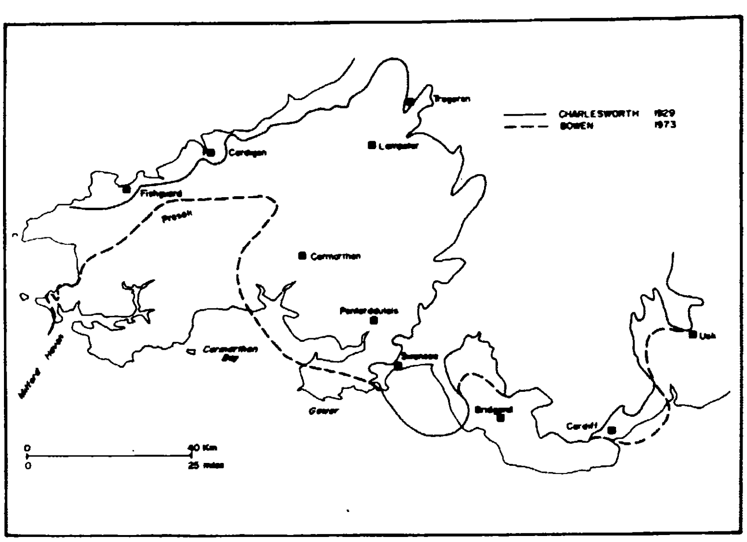


Figure 2.10 The South Wales End Moraine as proposed by Charlesworth (1929) and the current recognised Late Devensian limit of Bowen (1973a). (From Bowen, 1981).

The S.I.E.M. complex comprises mainly kame and kettle topography dominated by glaciofluvial sands and gravels, thus representing a major depositional ice limit. It has been suggested by McCabe (1987), however, that this complex may not necessarily demarcate the precise extent of the Late Devensian ice over Ireland. It is conceivable that the outer ice sheet limits may have been debris deficient, with large scale deposition concentrated some distance behind the main ice limit (McCabe, 1987). It is such that even where a well-developed moraine is known to exist, evidence has been found to imply that ice may have extended beyond the predicted margin. Indeed whilst the general position of this Late Devensian limit has been adopted by most workers (e.g. Mitchell, 1972; Synge, 1973; McCabe, 1985; Watts, 1985), a small number have nevertheless questioned its validity (e.g. Bowen, 1973b; Warren, 1979, 1985). A recurring problem appears to be that the chronology of glacial events have been largely identified from sequences which were formed in locations subjected to complex ice-marginal oscillations in response either to glaciological or climatic changes (McCabe, 1987). It is accepted though that the events identified are most likely to be of Late Devensian age (McCabe, 1987).

2.3.2.4: Onshore eastern extent - South Wales End Moraine

The first comprehensive Late Devensian limit in the eastern extremity of the Irish Sea Basin, coincident with South Wales, was presented by Charlesworth (1929) to the Geological Society of London (Figure 2.10). A downfall in the proposed extent was a lack of fundamental evidence, as maps and memoirs of the Geological Survey were only available south of a line drawn through approximately Carmarthen. Furthermore, to the north of this, Charlesworth was only able to base completed work on earlier papers by Jehu (1904), on Preseli (north Pembrokeshire), and Williams (1927), on south Cardigan (Cardiganshire). Neither of these publications had produced geological maps of the Pleistocene deposits (Bowen, 1981). The argument presented by Charlesworth (1929) was that the Irish Sea ice sheet would have dammed the mouths of local streams, causing proglacial lakes to develop. Ice-marginal overflow channels, which drained the lakes, would therefore serve to indicate the ice margin (Charlesworth, 1929). This criteria was therefore used as the main control for reconstructing some 72 km of the Irish Sea Ice Sheet margin (mainly in Preseli, North Pembrokeshire and Ceredigion, Cardiganshire), out of a total length of around 96 km (Bowen, 1981). The other 24 km was indicated by moraines, later to be reinterpreted as fluvioglacial features, which Charlesworth (1929) described as 'discontinuous', 'moderate in dimension' and 'rarely conspicuous'. The ice margin portrayed by Charlesworth (1929) coincided more or less with the southern limit of continuous drift and this association was used to extend the proposed limit to the east of Swansea. To the west this precept was disregarded and an arbitrary line inferred to be 'conjectural', 'uncertain' and 'very doubtful' (Charlesworth, 1929, p. 345) was drawn north to Tregaron, where a prominent end moraine occurs (Bowen, 1981).

George's (1932, 1933) work on the glacial and coastal Pleistocene deposits of Gower confirmed Charlesworth's limit on the western side of Swansea Bay. However Griffiths (1937, 1939, 1940), by examining the mineralogy and erratic content of glacial deposits west of Cardiff, confirmed the critical response of Jones (1929) who was familiar in detail with the local geology. Wirtz (1953) later suggested that only limited local valley glaciation occurred during the last glaciation, with the Irish Sea ice sheet impinging locally around Cardigan and Fishguard. In the 1960's, a lithostratigraphic approach to the problem of ice limits supported by other evidence was adopted, thus with less emphasis placed on geomorphology. In south-east and central South Wales Bowen (1970) related the Pleistocene geology to the lithostratigraphy of Gower, and the extent of Late Devensian Welsh ice was shown to be more extensive than envisaged by Charlesworth (1929) (Figure 2.10). In Preseli, John (1970) showed that the coastal lithostratigraphic succession was identical on both sides of Charlesworth's (1929) line, thus invalidating it as a significant ice margin (John, 1970). John (1970) argued that the Irish Sea ice sheet had reached south Pembrokeshire and possibly beyond. Subsequently John (1974) indicated that it terminated between Milford Haven and the south Pembrokeshire coast.

Further lithostratigraphic documentation of South Wales (Bowen, 1973a, b, 1974, 1977) as a whole followed on from the preliminary work of Bowen (1970). Using the last interglacial *Patella* beach as a stratigraphic marker, Bowen (1973a, b, 1974) was able to discriminate between glaciated and unglaciated areas by the presence or absence of glacial deposits in stratigraphic position above the raised beaches. It was inferred therefore that a 'older drift' glaciation had antedated the raised beach interglacial, and that the 'newer drift' glaciation (South Wales End Moraine) was the sole glacial event subsequent to that interglacial. The Late Devensian ice limit outlined by Charlesworth (1929) has nevertheless, despite continuous reassessment, survived with only minor modification in the area east of Swansea Bay. Bowen (1981) believes that, in retrospect, Charlesworth was over-concerned with the search for missing moraines to delimit the Late Devensian ice extent. It is now clear that the former ice margin is not marked by a continuous morainic ridge or topographic zone (Bowen, 1981). In light of this a number of problematic areas still exist (e.g. west and south-west Gower) although the overall outline currently proposed by Bowen (1973a) is thought to be secure.

2.3.2.5: Timing of Late Devensian ice sheet maximum

The precise timing of the Late Devensian maximum in the Irish Sea Basin is not well known, although it is generally recognised to have occurred somewhere around 22 000 yrs B.P. (Bowen and Sykes, 1988). Mitchell (1972) suggests that in North America there is good evidence that the main advance of Wisconsinan ice took place about 20 000 yrs B.P., and the first advance of ice down the Irish Sea may well be of the same age. This theory is somewhat meaningless, though, given that Dyke and Prest (1987) have indicated that there are various ages derived for the Late Wisconsinan maximum, arising from the long-standing problem of defining the Late Wisconsinan limit. Radiocarbon dates from raised glaciomarine studies in Ireland (Colhoun *et al.*, 1972; McCabe *et al.*, 1986) suggest that the Late Devensian glaciation peaked in this location between 24 000 and 20 000 yrs B.P. Definite maximum ages are provided by the 30000 yrs B.P. radiocarbon ages from Four Ashes, in Staffordshire (Shotton, 1967).

Further maximum age indicators may be obtained from amino acid D/L ratios of *in situ* shells incorporated into glacial drift. This method shows that the glacial drifts of south-west Dyfed (e.g. Abermawr and Banc-y-Warren), and north Gower, are Late Devensian in age because they contain shells of isotope stage 3 (Bowen and Henry, 1984). In Ireland, Bowen *et al.* (1986) suggest that the Late Devensian maximum probably occurred around 23 000 yrs B.P. However, if the inferences made by Scourse (1985) are accurate then an ice 'surge' as far as the northern Isles of Scilly may have occurred around 18 600 ± 3 700 yrs B.P. (Wintle, 1981). A similar date of 18 000 yrs B.P. is available from a carpal bone of a woolly mammoth from the Tremeirchion Caves of the Vale of Clwyd, north-east Wales (Rowlands, 1971), which have been sealed by Irish Sea till.

It is therefore apparent that there is great speculation over the exact age of maximum ice expansion thus at least for the meantime, until further maximum ages are obtained, a restricted period in which the maximum limit was attained is all that can be accepted. Relatively few maximum age determinations have arisen because of the lack of organic material available from what was an extremely harsh climatic environment immediately prior to, and during maximum glaciation (Bowen *et al.*, 1986).

2.4: Full glacial ice dynamics

2.4.1: Ice sheet geometry - impact of a deforming bed hypothesis

It has been recognised that former ice sheet profiles and ice dynamics can be tentatively reconstructed if their extent and length of flow lines are known and if glacier budgets and palaeotemperatures can be estimated (Boulton *et* al., 1977). Following from this, Boulton *et al.*

(1977) have presented the results of a modelling exercise for the whole of the Late Devensian British ice sheet which incorporated a modelled surface topography, distribution of mass balance velocities and ice-flow lines (Figure 2.11). A number of useful inferences have been drawn from these results with respect to the morphology, kinetics and thermal structure of the Irish Sea ice sheet during its full glacial maximum. For the Irish Sea Basin, the model postulated an ice sheet running approximately south-southwest from a maximum surface elevation of around 1800 m over the eastern part of the Southern Uplands, south-west Scotland. In the central area, basal ice was predicted to be cold whereas nearer to the margins of the ice sheet, temperate ice existed which accommodated high marginal ice-flow velocities in the range of 150 to 500 m/yr (Boulton *et al.*, 1977). The position of the outer margin of the Irish Sea ice sheet, at its maximum, was based on Charlesworth's (1928, 1929) synthesis as generally accepted by most former workers (Section 2.3.2).

Although this reconstruction may be seen as useful in terms of a first approximation of ice dynamics within the Irish Sea Basin, the modelling itself can be heavily criticised in a number of ways. First, Boulton *et al.* (1977) used Weertman's (1961) analysis to determine the surface shape and velocity distribution of the Late Devensian ice sheet over Britain and their own analysis to determine basal temperatures. They assumed that at its maximum the ice sheet was in steady-state equilibrium; the margin was stationary, ablation balanced accumulation, the glacier flow field was constant with time, and the ice sheet was in thermal equilibrium with the atmosphere and its bed. Given their estimated build-up period for maximum expansion of the Late Devensian ice sheet of *c.* 15 000 yrs B.P. (based on a mean accumulation rate of 0.1 m/yr) this is thought to be too long a period in which to establish steady-state conditions (Thomas, 1985). Furthermore, Boulton *et al.* (1977) admit that expansion may indeed have been stopped short by a rapid climatic amelioration thus preventing the ice sheet from reaching equilibrium conditions. Secondly, the model assumes Late Devensian advance across a terrestrial floor, which although likely, has by no means been proved (Thomas, 1985). Many workers, dating as far back as Lewis (1894), have raised the possibility that the Irish Sea ice sheet may have advanced into a marine margin as an ice shelf.

A third criticism is that the model is wholly based on a modern rigid-bed analogue (Thomas, 1985). Many theories developed to account for surface profiles and movement of ice sheets generally assume that the subglacial ice is in contact with a rigid (bedrock) surface, and that these two properties are dependent on the rheological status of the ice (Nye, 1957) and the thickness of a lubricating water layer at the ice-bed interface (Nye, 1957; Weertman, 1961). Whereas the assumption of a rigid bedrock substrate underlying modern glaciers is most

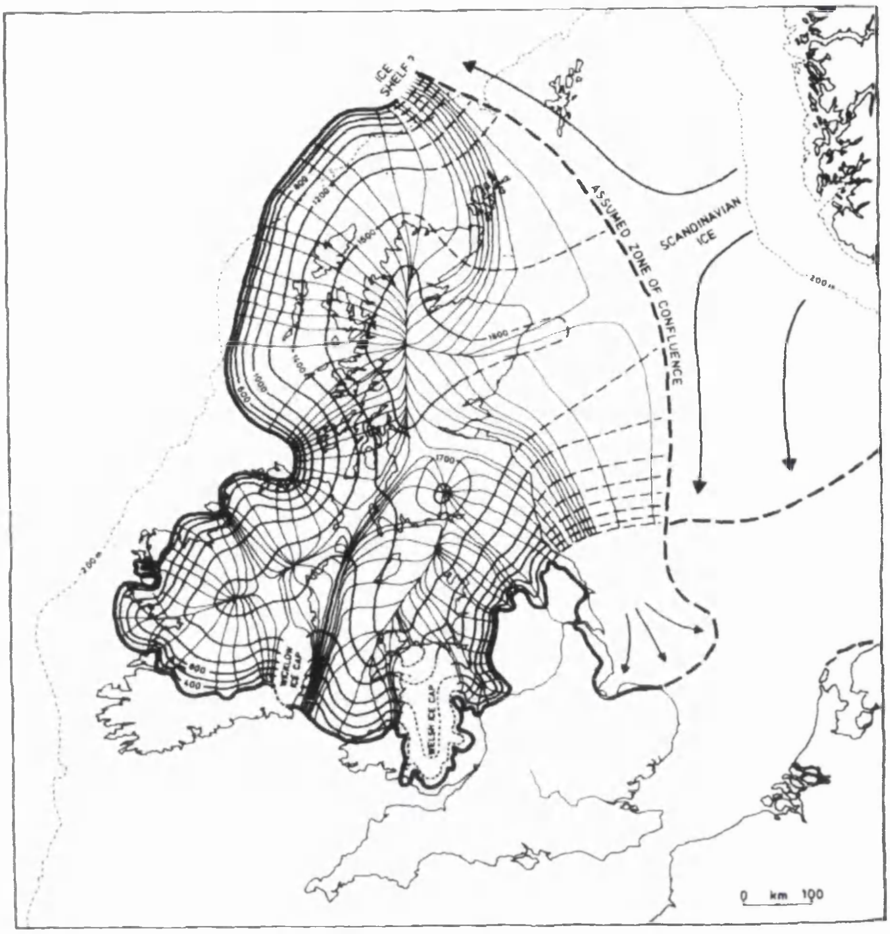


Figure 2.11 Modelled surface topography and flow lines of the Late Devensian British ice sheet (From Boulton *et al.*, 1977).

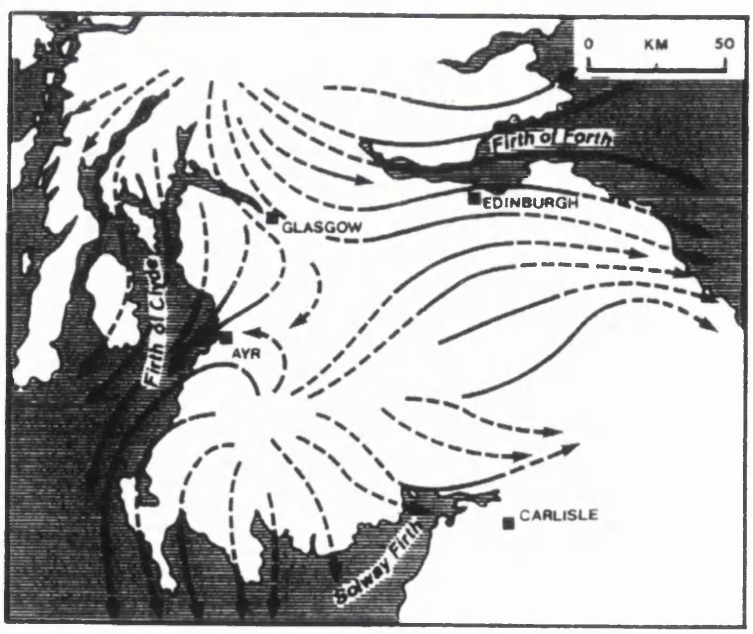


Figure 2.12 Directions of ice-flow in Southern Scotland as envisaged by Geikie (1901) (From Kerr, 1982b).

probably correct, there are many places where glaciers rest on beds which are partly or wholly made up of unlithified sediments which are potentially deformable (Boulton and Jones, 1979; Fisher *et al.*, 1985). Measurements beneath a contemporary temperate glacier sitting on till show that 90% of the basal movement of the ice can be attributed to deformation within the deformable till substrate rather than by internal flow or through movement at the ice-bed interface (Boulton and Jones, 1979).

A theory has been developed by Boulton and Jones (1979) whereby the profile of the glacier or ice sheet may be related to the hydraulic and strength properties of underlying deformable bed materials. In its simplest form, the theory predicts that the geological properties of the subglacial bed will influence the surface profile of a glacier or ice sheet, and that ice sheets moving over fine-grained, unlithified sediments (low cohesion and relatively impermeable) will tend to develop a relatively low profile (Boulton and Jones, 1979). Therefore, if a bed has a high hydraulic transmissibility, meltwater is readily discharged subglacially, the bed maintains its stability, and the corresponding ice sheet profile is a normal parabolic one. If the bed transmissibility is low, water pressures build up, the bed begins to deform, and a lower equilibrium profile will develop (Boulton and Jones, 1979).

A large proportion of the beds underlying the Pleistocene ice sheets which covered much of North America and north-west Europe are known to have contained deformable sediment. The Irish Sea Basin itself seems a potentially likely site to promote the scenario of Late Devensian ice sheet flow by means of a subglacial deforming substrate. This is based upon the assumption that prior to the Devensian expansion, the Irish Sea Basin was underlain, as now, by a thick sequence of soft glacial and interglacial marine sediments (Thomas, 1985). It has therefore been suggested by Thomas (1985) that the Irish Sea ice sheet would most likely have maintained a much lower equilibrium profile than predicted by the outcome of the rigid-bed model. The construction of a 'minimum' model for Late Devensian ice geometry in the Irish Sea Basin has therefore been documented by Thomas (1985) in which three components are identified with respect to the physiographic setting of the ice sheet.

Over the upland region of south-west Scotland and much or Ireland the ice sheet is envisaged by Thomas (1985) to have had a gradient and thickness commensurate with the source-area, rigid-bed analogue. The magnitude of ice thickness over the northern part of the northern Irish Sea Basin was therefore possibly in excess of 1500 m (Thomas, 1985). However, as the ice passed south into the basin proper, a lower-gradient thin ice sheet developed as a function of movement over the base of unconsolidated deformable sediment. Fisher *et* al. (1985), in modelling the effects of possible deforming beds on a Laurentide ice sheet reconstruction, infer a situation akin to this whereby flow lines crossing from a hard bed to a deformable bed cause the surface slope to change by a factor $(\tau_{soft}/\tau_{hard}) \sim {}^{1}/_{3}$ to ${}^{1}/_{10}$ and the direction of the flow lines to change dramatically. In low $\tau_o{}^*$ areas ice flow is thought by Fisher *et al.* (1985) to be sensitive to topography and is more readily diverted to low areas or around uplands whereas, in hard bed areas, the basal topography is thought only to partially control ice flow direction.

The deformable beds of the Irish Sea Basin, into which ice flowed from the peripheral upland masses, may therefore have induced relatively low surface slopes and promoted remarkable abrupt ice flow direction changes particularly in the transition zone between the potential unconsolidated basinal sediment and hard elevated bedrock. Thomas (1977) suggests that Devensian Irish Sea ice rose no higher than 180 m onto the margins of the Isle of Man. Therefore from this height, a very low gradient* of approximately 100 m in 300 km is indicated for the proposed limit of Devensian Irish Sea ice in Pembrokeshire and south-east Ireland (Thomas, 1977). This is an order of magnitude or more lower than the observed gradients of the Greenland ice cap (Paterson, 1969) and the reconstruction of both the Scandinavian (Flint, 1970) and the Laurentide ice sheets (Andrews, 1973). It is also below the anomalously low gradients reported by Mathews (1974) on some margins of the Laurentide ice sheet. At the southern end of the Irish Sea Basin the third component of the ice sheet identified by Thomas (1985) has been approximated to a thin flat sheet which questionably may have passed into a eustatically lowered sea level as a floating ice-shelf.

2.4.2: Ice dynamics in south-west Scotland

For an insight into the nature of Late Devensian ice dynamics in south-west Scotland attention has to be turned to the pioneering work of Charlesworth (1926a) for it is solely this investigation into 'The glacial phenomena of the Southern Uplands of Scotland' (Charlesworth, 1926a, p. 1) which bears any detailed reference upon the subject. In stating this, however, it would be correct to acknowledge that the direction of Charlesworth's work was most probably guided by the foundation work of Geikie in whose text 'The Scenery of Scotland' (1901) is included a generalised map showing the direction of ice flow in the Southern Uplands (Figure 2.12). Similarly influential were the contributions provided by Professor J.W. Gregory (1925), which are summarised in Section 1.4.

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* τ_o = yield shear stress (Nm^{-2}).

* This gradient estimate ignores a relative reduction in current altitudinal differences caused by a normal, but unknown, amount of isostatic depression at the time of maximum glaciation, an effect that might lower the gradients considerably (Thomas, 1977).

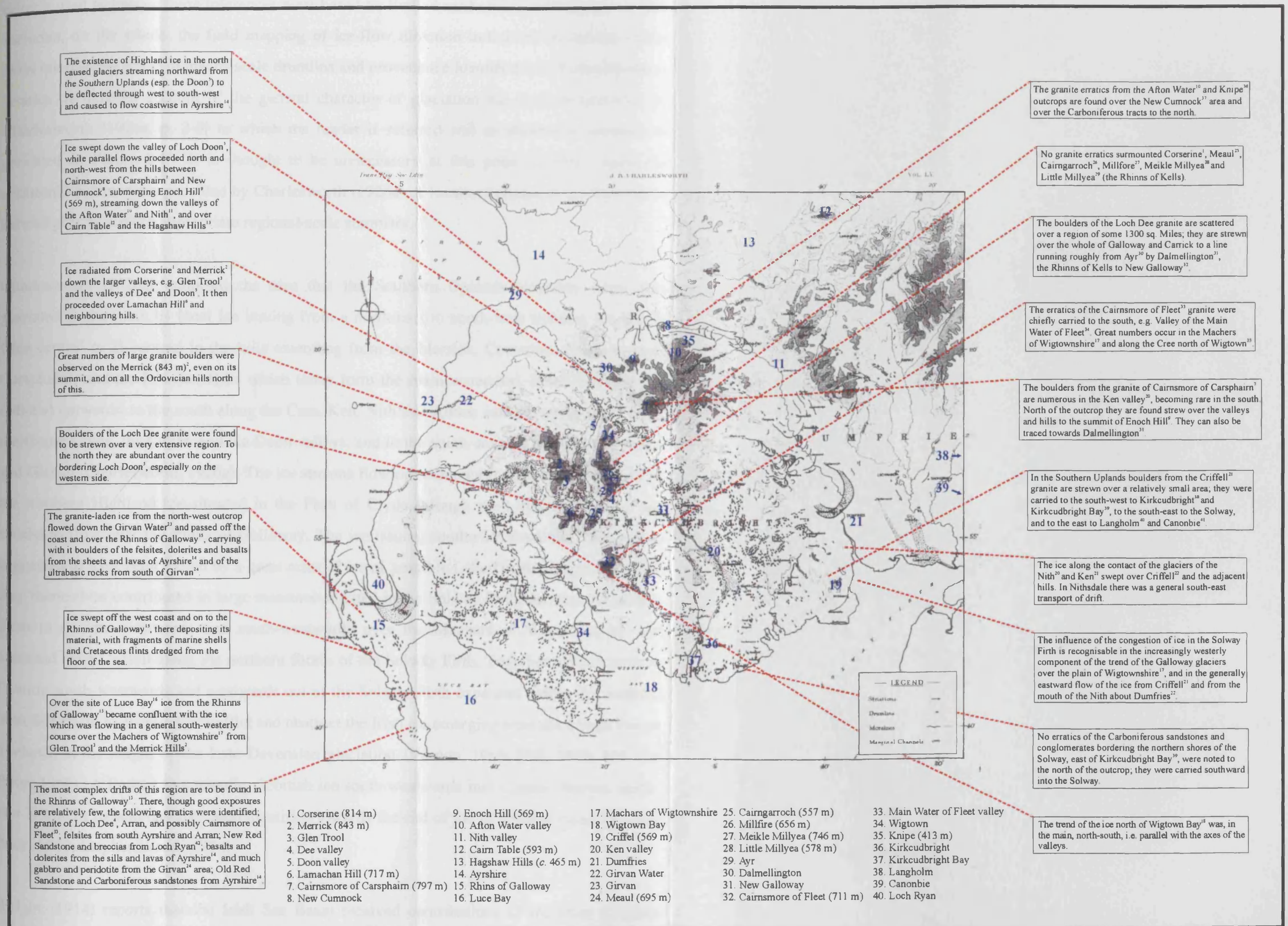

Figure 2.13 Summary of the general character of Devensian glaciation in south-west Scotland as documented by Charlesworth (1926a).

The significant aspects of Charlesworth's (1926a) work are clearly illustrated by a constrained geomorphological map (Figure 2.13) which attempted, among other things, to reconstruct the trend of advancing ice flow over the upland and coastal lowland of south-west Scotland. The criteria used to initiate these inferences were based on limited methods of field analysis which included, on the whole, the field mapping of ice-flow direction indicators on various scales from micro-scale striae to macro-scale drumlins and provenance identification of allochthonous erratics. A detailed account of the general character of glaciation has been documented by Charlesworth (1926a, p. 2-8) to which the reader is referred and an illustrated summary is provided in Figure 2.13. It is thought to be unnecessary at this point to reflect upon the additional commentary provided by Charlesworth (1926a) of localised phases of ice flow but to instead portray a more appropriate regional-scale summary.

Charlesworth (1926a) endorsed the idea that the Southern Uplands mountain range was glaciated in the main by local ice issuing from a north-east to south-west trending ice-divide ('ice centre' p. 2) centred in the hills extending from the Merrick, Corserine, Cairnsmore of Carsphairn, and the Lowther Hills which today form the main watershed. From this axis, ice radiated outwards, to the south along the Cree, Ken, Nith and Annan valleys, to the north along the Upper Clyde, the Lugar, and the Doon valleys, and to the south-west by the Stinchar valley and Girvan (Charlesworth, 1926a). The ice streams flowing to the north-west were deflected by the northern Highland ice situated in the Firth of Clyde through west into south-west, and caused to flow over the Rhins of Galloway. The ice issuing southward from the ice-divide is suggested to have been forced by a great mass of ice of possible Lake District origin (to which they themselves contributed in large measure) situated in the Solway Firth, thereby compelling them to deviate westwards and south-westwards over the Machars of Wigtownshire, and eastward from Criffell along the northern shores of the Solway Firth. Together, the ice streams flowing south-westwards and westwards out of the Solway Firth expanded across the northern Irish Sea Basin to eventually conflict and obstruct the Irish ice emerging from the Lough Neagh lowlands at the height of the Late Devensian glaciation (Vernon, 1966; Hill, 1968; Hill and Prior, 1968). A limited incursion by Scottish ice south-westwards into County Antrim, north-east Northern Ireland has also been identified towards the end of the Devensian stage (Hill and Prior, 1968; Stephens *et al.*, 1975).

Wright (1914) reports that the Irish Sea Basin received contributions of ice from all sides during the Devensian glacial period. "From Wales on the south, from Galloway [south-west Scotland] to the north, from the mountains of Cumberland and Westmoreland [north-east England] on the east, but above all from the great central ice sheet of Ireland, glaciers poured down into it on every side" (Wright, 1914, p. 50). Congestion was foreseen in the northern half of the Irish Sea Basin and therefore some form of outflow was deemed necessary to the south. It is proposed by Wright (1914) that the whole mass of amalgamated ice streams moved in a south-south-easterly direction over the Isle of Man, onto the northern shores of Wales, forcing back the local Welsh ice, and dividing it at this point into two streams. The less dominant of these two ice streams is believed to have flowed in a south-south-east direction over the plains of Lancashire and Cheshire in north-west England. Meanwhile, the other more powerful stream by far, continued south into St. George's Channel as far as the maximum offshore limit discussed in Section 2.3.2.2 (Wright, 1914).

2.5: Devensian Deglaciation - Glacioterrestrial Versus Glaciomarine

2.5.1: Regional deglacial chronology

The deglacial chronology of the Irish Sea ice sheet is poorly documented however it is generally believed that subsequent to its maximum peak around 22 000 yrs B.P. retreat was rapid. A few critical sites have been identified where dateable evidence of known stratigraphical position exists providing a tentative date of when Irish Sea ice receded from these areas. On the Llêyn peninsula, north Wales an investigation by Coope and Brophy (1972) into Late Glacial environmental changes provided a radiocarbon date of 14 468 ± 300 yrs B.P. (Birm 212; Shotton and Williams, 1971) obtained from a moss-rich layer above sand and gravel and marking the base of a Late Glacial sequence infilling a kettle hole (based on moss debris and seeds of *Caryophyllaceae*). The Llêyn peninsula is thought to have been ice free by at least this time with early deglaciation of the Irish coast by 17 000 yrs B.P. (Eyles and McCabe, 1989a). On the Isle of Man dates from organic sediments overlying the Devensian glacial sequence have been obtained which indicate that the Irish Sea ice sheet had cleared the island by at least 15 150 B.P. (Bowen, 1977; Dackombe and Thomas, 1991).

West Cumbria is thought to have been deglaciated by about 14 500 yrs B.P. (Pennington, 1978). The record provided by assemblages of insects including beetles (sensitive indicators of climatic change) and pollen preserved in organic sediment showed that rapid amelioration of climate took place after about 13 600 yrs B.P. marking the onset of the Windermere interstadial (Akhurst *et al.*, 1997) (about 13-11 ka B.P. according to Coope and Pennington, 1977). Pennington and Bonny (1970) have shown that the Lake District was deglaciated by at least 14 300 yrs B.P. (Pennington and Bonny, 1970) and a radiocarbon date of 12 940 ± 250 years (Q-643) at Robert Hill at the head of the Solway Firth provides a limiting date on deglaciation for this area (Bishop, 1963). Sutherland (1984) showed that the mouths of the fjords in north-west Scotland had been deglaciated by 13 000 yrs B.P. The Firth of Clyde was ice free by 13 020 yrs

B.P. (Bishop and Dickson, 1970) and by 12 600 yrs B.P. the inner parts of the Firth of Clyde had been deglaciated (Peacock, 1975, 1981; Peacock *et al.*, 1977, 1978). Gemmell (1973) has given a date of 12 500 yrs B.P. for Arran and by 12 000 yrs B.P. Moar (1969) believes peat was already forming in Dumfriesshire.

2.5.2: Terrestrial model - A retreating land-based glacial system

Conventionally, there has been a consensus view that most Devensian sediment sequences were produced by deposition from retreating land-based (terrestrial) glacial systems within the Irish Sea Basin. It subsequently follows on, within this school of thought, that fine-grained Irish Sea Tills are subglacial or glaciolacustrine in origin. The Irish Sea ice sheet is therefore pictured in this model as a fully terrestrial ice margin with large lakes dammed along its margins (e.g. Thomas and Summers, 1981, 1982, 1983; Thomas *et al.*, 1985; Thomas and Kerr, 1987; Thomas, 1989). The 'terrestrial' model has persisted in the literature for more than a century, since the recognition by Tiddeman (1872) of a Pleistocene Irish Sea glacier whose deposits consisted of terrestrial tills and outwash sediments. Recently, Thomas and co-workers have revisited this original interpretation and provided further evidence from localised depositional environments (Section 2.5.2.1), mainly consisting of sequential glacigenic sequences which validate this model. In doing so, emphasis appears to have been placed on the correlation, often over long distances, of generalised stratigraphies besides reconstructing environments of deposition.

2.5.2.1: Characteristic sediment associations and palaeontology in County Wexford, Ireland

In the low coastal cliffs around eastern County Wexford, Southern Ireland, Thomas and Summers (1982, 1983) have concentrated on a major Quaternary lithostratigraphic formation, the Blackwater Formation, whose deposits were laid down by ice advancing from the Irish Sea across the east and south-east of Ireland (Culleton, 1978). It is the stratigraphic sequences in this region which have recently been regularly cited as supporting evidence for the original concept of a retreating terrestrially-based Irish Sea ice sheet margin. Culleton (1978) describes the depositional pattern of the drift around Wexford as 'complex', varying from large areas of sand and gravel to flat till sheets, with an admixture of both types in places. Within the formation consist the more prominent Macamore, Screen, Ballinclash and Knocknasilloge members (Thomas and Summers, 1982, 1983) which have been separated on the basis of the physical and chemical characteristics in accordance with the Code of Stratigraphic Nomenclature (1961) (Culleton, 1978).

The Macamore member, or Irish Sea Till as it is commonly termed, has been interpreted as a lodgement till. It is suggested to have been emplaced by a small lobe of active, terrestrially based ice which pushed in from the Irish Sea, in the Devensian period, onto the coastlands of south-east Ireland, well to the south of the area considered here (Charlesworth, 1928; Culleton, 1978; Thomas and Summers, 1983). The deposits cover a large area in the east and south-east of County Wexford (Culleton, 1978). Contrary to this theory, however, Huddart (1977, 1981a) earlier tentatively suggested that this till may instead be of glaciomarine origin, thereby supporting a view first proposed by Lewis (1894) and latterly supported by Synge (1977) (Section 2.5.3). The mechanisms envisaged in this glaciomarine scenario involved deposition by means of melt from the base of a floating ice-shelf. Thomas and Summers (1983) strongly criticise this hypothesis, with caution, and considered it premature in view of both the difficulty in establishing adequate criteria for distinguishing glaciomarine sediment from terrestrial till (Mathews, 1980), and the profound implications for palaeoclimate, sea-level change and ice-sheet geometry that it carries.

The Macamore member is stratigraphically succeeded by the Screen member; a complex suite of pro-glacial outwash deposits that crop out along much of the east Wexford coast. The main concentration occurs as kames in the Screen-Blackwater area where they form a continuous belt of kame and kettle topography (Culleton, 1978). Occasionally though, the discontinuous Knocknasilloge member intervenes between these two units (Thomas and Summers, 1983). The Screen member is in turn succeeded by the limited Ballinclash Member, a till of Irish Sea origin but one differing from the Macamore in its lithological character (Thomas and Summers, 1982). Thomas and Summers (1982) describe a deposit overlying the Macamore Till which they believe differs significantly in lithological character from Macamore deposits seen elsewhere. As it is thought to differ also from other members of the Blackwater Formation this unit has been defined by Thomas and Summers (1982) as a separate Ely House member, named after the location of its type site.

The Ely House member is lithologically made up of a number of distinctive facies types, details of which are outlined in Thomas and Summers (1982). Of significance within the Ely House deposits were the presence of stratification, dropstone structures, clast and grain clusters, the close association with rhythmites, plus the occurrence of concretionary structures akin to those found only in glacial lakes, pointing strongly to a sub-aquatic origin (Thomas and Summers, 1982). Apart from the clast pattern, which showed a clear preferred orientation of a-axes parallel to regional ice directions, the sedimentary characteristics of these deposits certainly beared little resemblance to those of terrestrially based tills found elsewhere in the region. It has subsequently been proposed by Thomas and Summers (1981), in light of this evidence, that the Ely House deposits closely resemble the lithological description and characteristics of a waterlaid till described by May (1977) (equals waterlain till of Dreimanis (1979)).

Furthermore, an association has been made with the depositional model proposed by Gibbard (1980), that involves sedimentation by bottom melt beneath a floating ice-shelf near its grounding line. It is envisaged that this till unit was manufactured by debris release from the base of a floating ice-shelf that abutted a large glaciolacustrine basin covering much of the lower Slaney valley, north-west of Wexford (Thomas and Summers, 1983). This was accompanied by periodic release of sediment-laden meltwater from subglacial sources (Thomas and Summers, 1981). The sequence of subaqueous flow tills described at Ely House are believed to be stratigraphically equivalent to the Knocknasilloge member and likewise are, as previously indicated, seen to merge upwards from the underlying Macamore member and on up into the overlying Screen member (Thomas and Summers, 1983). Thomas and Summers (1983) argue, therefore, that a subaquatic lithofacies appears to persist at the same stratigraphic position throughout the Wexford area. Despite this, it is believed that the deposits do not pertain to one single proglacial lake system but rather to a number of discrete lakes of relatively small dimension and probable short duration (Thomas and Summers, 1983).

Aside from the stratigraphic context and depositional setting of the deposits, attention has additionally been focused upon the strong occurrence of microfauna within the Blackwater Formation, in particular the Knocknasilloge member. The environment of deposition of the Knocknasilloge member remains contentious and opinions range widely. Resolution of this issue is thought to be very important because the differing views are seen to radically affect crucial questions of glacial chronology, limits of ice advance, and the importance of eustatic and isostatic processes in the southern Irish Sea (Thomas and Kerr, 1987). Microfauna collected by Huddart (1981a) from the sequence at its type site at Knocknasilloge were found to contain a high concentration of *Elphidiella hannai* which prompted the initial suggestion that the fauna, and hence surrounding deposits, were *in situ* and that they may have survived in an Ipswichian interglacial 'refuge' in the Irish Sea. Although this species has been discovered in North Sea Early Pleistocene deposits Thomas and Kerr (1987) argued that it did not occur in the Ipswichian deposits of Selsey or the supposed Ipswichian deposits of the Somerset levels (Kidson and Heyworth, 1973; Kidson *et al.*, 1974) or Shortalstown (Colhoun and Mitchell, 1971). They suggest that Huddart's (1981a) interglacial marine theory seems unlikely and it is more probable that along with other taxa typical of the Early Pleistocene, such as *Lenticulina*

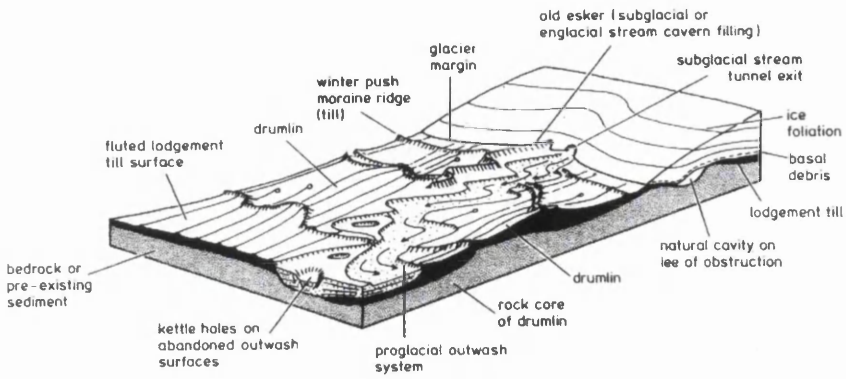

Figure 2.14 Schematic diagram illustrating the formation of the subglacial/proglacial sediment association and landsystem. Lodgement till and subglacial and proglacial outwash sediments are associated. A simple stratigraphy of outwash on till is produced by a single glacial episode of advance and retreat. (From Boulton and Paul, 1976).

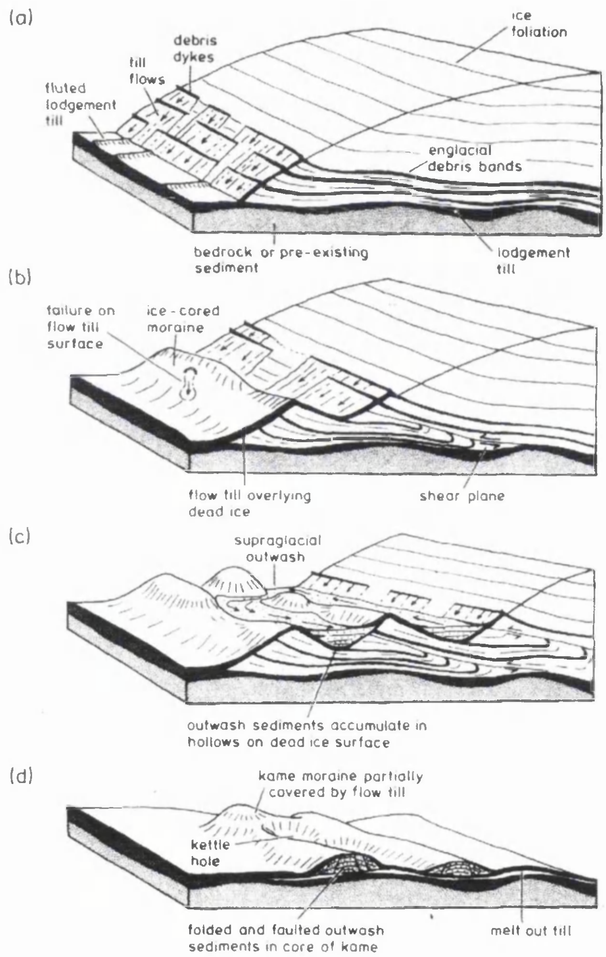

Figure 2.15 Schematic diagram illustrating the development of the supraglacial sediment association and landsystem. (a) Flow tills form on the surface of the retreating glacier from thick sequence of englacial debris. (b) Till cover inhibits ablation of underlying ice which is left behind during glacier retreat as an ice-cored moraine ridge. Supraglacial flow till is still active. (c) Outwash from the active glacier is forced to flow between ice-cored ridges. Tills flow into outwash systems, but are generally disaggregated until outwash dries up, when flow till forms a capping. (d) Dead ice melts, thus reversing the topography and leaving melt-out till in its place. The kame moraines which result are primarily formed of outwash, but are often capped by flow till. The kame sediments show collapse structures (From Boulton, 1976).

and *Nodosaria*, *Elphidiella hannai* was derived from pre-existing deposits on the floor of the Irish Sea.

Independent palaeontological evidence presented by Thomas and Kerr (1987) further suggests that the microfauna were not *in situ* for they contain a mixed assemblage of forms, drawn from different depths, salinity and temperature environments, and include a number of species that are characteristic of, or which became extinct, in various stages of the Early and Middle Pleistocene. In addition, evidence of transportation and reworking was shown and the overall vertical distribution of species did not conform to the pattern of variation described by Jasin (1976) from a type interglacial. Despite this contradiction Huddart (1981b) has since entertained the theory that the Knocknasilloge fauna are *in situ* and could represent Late Devensian glaciomarine conditions (Section 2.5.2.3). Nevertheless, Thomas and Summers (1981, 1983) still maintain that the environment of deposition of the Knocknasilloge member, and with that the Irish Sea till as a whole, was glaciolacustrine.

2.5.2.2: Associated subglacial/proglacial/supraglacial landsystem assemblages

In view of the support for a terrestrial-based Irish Sea ice sheet Thomas and Summers (1982, 1983) propose a sequence of events that account for the nature of the stratigraphic successions typified in the coastal region of County Wexford. These have been closely related to two principal sedimentary associations (the Landsystems approach; Section 3.3.1.4) proposed by Boulton (1976) which characterise deposition by large terrestrial ice sheets in lowland areas; the *subglacial/proglacial sediment association* in which the dominant till is lodgement till generated by temperate-based glaciers where debris transportation occurs in a relatively thin zone (Figure 2.14), and the *supraglacial sediment association* in which melt-out and flow tills occur associated with supraglacial kamiform outwash and proglacial outwash superimposed on lodgement till, thus revealing a frequently complex tripartite (till-outwash-till) sequence (Figure 2.15). The former sediment association, in its basic form, produces a simple stratigraphic couplet whereby lodgement till is overlain by proglacial outwash (Thomas and Summers, 1983). This couplet is shown in Boulton's (1976) model to be associated with a landsystem which incorporates till sheets, minor push moraines and drumlinised forms, locally overlain by occasional eskers, kames and outwash systems (Thomas and Summers, 1983). Thomas and Summers (1983) draw similarities from this assemblage to an area to the south of the Screen Hills which displays many of the characteristics of the subglacial/proglacial sediment association and landsystem.

South and south-east of Wexford Town, Thomas and Summers (1983) isolate a featureless lowland area which is underlain by a thick sheet of Macamore member and limited outwash. This is suggested to have been generated from lodgement processes during the advance of the Irish Sea ice but that on stagnation the ice receded rapidly to uncover a till plain supporting very little outwash. Thus, this area has been identified as a clear representation of essentially a subglacial sediment and landsystem assemblage (Thomas and Summers, 1983). In contrast, to the north and north-west of Wexford Town in a zone peripheral to the southern margin of the Screen Hills, has been identified an area also underlain largely by Macamore member but one much diversified by isolated kame and outwash forms on its surface. This area is suggested to be more akin to a proglacial sediment and landsystem assemblage evolving in a situation where the ice margin retreated in a less rapid fashion during which significant volumes of outwash were laid on top of the Macamore till (Thomas and Summers, 1983). In both modes of ice retreat it is envisaged by Thomas and Summers (1983) that temporary lakes were impounded in irregularities on the subglacial/proglacial surface, thereby accounting for the Knocknasilloge member. The existence of such a lake, dammed against the stagnating ice margin, has already been proposed above by Thomas and Summers (1983) (Section 2.5.2.1).

The supraglacial sediment association is believed by Boulton and Paul (1976) to be produced by cold-based sub-polar or polar glaciers which transport debris through a considerable thickness of basal and englacial ice (Figure 2.15). From this, it is suggested that the resulting landsystem is characterised by a complex of kame moraines, kettle basins and pitted outwash surfaces showing much reversal of topography (Boulton and Paul, 1976; Thomas and Summers, 1983). A contemporaneous analogy is drawn by Thomas and Summers (1983) to the Screen Hills; a type example of this sediment and landsystem assemblage.

2.5.3: Glaciomarine model - A collapsed ice sheet margin

Many early investigations carried out on Irish Sea Till (Portlock, 1843; Mellard-Reade, 1874, 1883; Brady *et al.*, 1874; Goodchild, 1874; Shone, 1878; Mackintosh, 1870, 1881; Lewis, 1894; Munthe, 1897; Smith, 1912) varyingly interpreted the deposit as either glaciomarine muds or glacially-thrust and transported rafts. Indeed, a number of similarities have been drawn by workers to contemporary glaciomarine deposits of Spitsbergen and Alaska (Eyles and McCabe, 1989a). Nevertheless, subsequent workers, with exception to a few (e.g. Synge, 1977; Huddart, 1977, 1981a (Section 2.5.2.1)), have dismissed such schools of thought in favour of the model outlined in Section 2.5.2 in which the deposits are viewed as glaciolacustrine deposits containing reworked fauna (Thomas and Summers, 1981, 1982, 1983; Thomas and Kerr, 1987). An implicit assumption, acting as a fundamental component in this model, is the existence of a terrestrial Irish Sea ice sheet, and therefore a glacio-eustatically lowered sea-level (Eyles and McCabe, 1989a). The 'validity' of the terrestrial model has been questioned recently by Eyles and McCabe (1989a) and the concept of marine deposition reintroduced. It is suggested that rather than representing deposition at a time of eustatically depressed sea-levels, most of the deposits of Late Devensian age in the Irish Sea Basin reflect glaciomarine conditions with isostatically raised relative sea-levels (Eyles and McCabe, 1989a).

Evidence cited to support this concept evolved from a revised interpretation of the Pliocene and early Pleistocene microfaunal components found within the Irish Sea Drifts; some of which are believed to be *in situ* and thus providing a mixed assemblage of both *in situ* and *reworked* taxa (Eyles and McCabe, 1989a). They are seen to exist largely as well-preserved but low-diversity boreo-arctic foraminifera in contrast to other shallow temperate species (e.g. *Elphidium crispum*, *Quinqueloluclina*) which are commonly damaged as a consequence of reworking, despite being more robust than the cold water forms (Eyles and McCabe, 1989a). Significantly, many fine-grained Irish Sea Drift units also contain substantial numbers of well-preserved ostracods (Thomas and Kerr, 1987) which do not survive reworking. The palaeoenvironmental significance of such mixing of reworked and *in situ* faunas is thought to be readily explained by deposition from suspended sediment plumes pumped from subglacial sources into the marine environment (Eyles and McCabe, 1989a). The foraminifera species have been noted, though, to vary in their relative abundance from site to site. Overall, the concept encapsulated in this glaciomarine model is that *in situ* cold-water ostracods and well-preserved boreo-arctic foraminiferal microfaunas are strongly associated with the Irish Sea Drifts. Other temperate species are clearly derived from earlier deposits and do not reflect the proposed nature of deposition (Eyles and McCabe, 1989a).

Eyles and McCabe (1989a) stress, however, that facies investigations (Section 3.3.1) should complement microfaunal studies in order to identify whether fine-grained deposits are *in situ* or are glacially-transported. By concentrating on detailed facies and biofacies examination of drift exposures of Late Devensian age along the coastal zone of the Irish Sea Basin, Eyles and McCabe (1989a) propose that the sediment sequences reveal a complex record of subglacial and glaciomarine sedimentation at and beyond a tidewater margin. Implications from this evidence are that the deglaciation of the Irish Sea ice sheet was controlled primarily by high relative sea levels as a result of flooding of a glacio-isostatically-depressed foreland basin (Eyles and McCabe, 1989b)

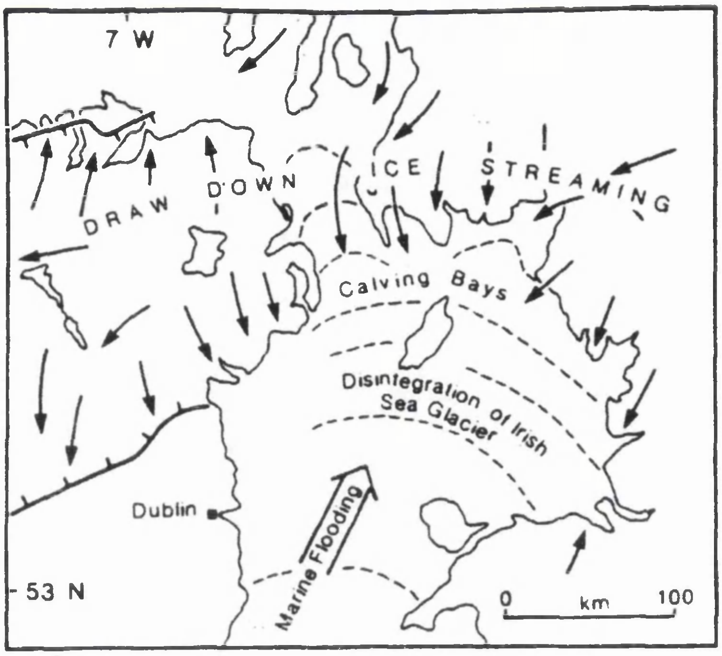

Figure 2.16 Disintegration of the Irish Sea Glacier as proposed by Eyles and McCabe (1989a) by calving, fast ice flow and drawdown. (From Eyles and McCabe, 1989a).

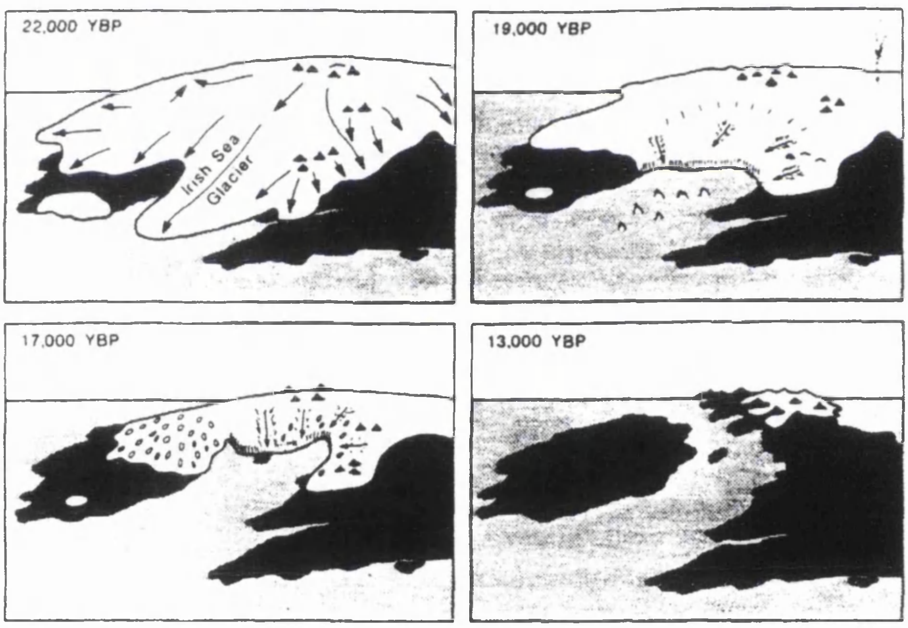

Figure 2.17 Schematic sketches illustrating Late Devensian environments in the Irish Sea Basin 22-13 ka B.P. *13 ka B.P.*: Final retreat of fjord glaciers in Scotland cessation of glaciomarine deposition (Clyde Beds) and start of rapid crustal rebound. Late glacial beaches are cut across glaciomarine sequences. A low marine stillstand caused by rapid isostatic recovery prior to the postglacial rise in sea-level allows accumulation of peat offshore. *17 ka B.P.*: Generation of fast ice flow, drumlinisation of interior portions of ice sheet (the 'Drumlin Event' in Ireland) and deposition of large morainal banks at tidewater margin and stratified glaciomarine complex offshore. Deposition of mud drape in areas well beyond limit of grounded ice (e.g. Southern Ireland and southern England). *19 ka B.P.*: marine incursion along St. George's Channel, as a result of glacio-isostatic subsidence followed by drawdown of ice margin and retreat of Irish Sea Glacier as a tidewater ice margin. Exposure of scoured subglacial topography across central part of basin. *22 ka B.P.*: maximum extent of Irish Sea Glacier coeval with glacio-eustatic sea-level lowering. (From Eyles and McCabe, 1989a).

The sequence of events envisaged by Eyles and McCabe (1989a) in their glaciomarine model are instigated by the initial 'unzipping' of the Irish and mainland ice streams which were formerly confluent along the central portion of the Irish Sea Basin (Figure 2.7). Deglaciation of the Irish Sea Basin is viewed in this model to have been triggered by rising relative sea-level, locally by as much as 180 m, as the basin underwent glacio-isostatic subsidence. This allowed marine waters to flood along the central trough from St. George's Channel sometime after 22 000 yrs B.P. creating a calving bay ice margin along the central part of the basin (Eyles and McCabe, 1989a) (Figures 2.16, 2.17). The growth of a large British ice sheet would have inevitably imposed a substantial load on the Earth's crust causing it to become downwarped (Walcott, 1970; Andrews, 1978, 1982; Quinlan and Beaumont, 1981, 1982; England, 1982, 1983). True to this fact Andrews *et al.* (1973) estimated that the equilibrium isostatic depression in the northern Irish Sea Basin may have been in excess of 500 m during maximum glaciation.

A fundamental point raised by Eyles and McCabe (1989a), however, is that in order for the Irish Sea Glacier to have advanced into the southern Irish Sea Basin it would be necessary to have had sea-level at or below the current level. This is primarily because of areas of deep water, such as occurring in the structurally-deepened troughs along the North Channel and St. George's Channel/Celtic Sea, and the high energy tide-dominated conditions which are both likely to have inhibited grounded ice advance (Eyles and McCabe, 1989a). The possibility of a relatively low gradient of the Irish Sea Glacier, as discussed in Section 2.4.1, would render the ice sheet margin incapable of moving into deep water. Southward extension as a substantial ice shelf is also thought unlikely given the storm and tide-dominated character of the basin, the lack of pinning points in the central part of the basin and the fragility of ice shelves (Eyles and McCabe, 1989a). Dating of Mid Devensian sea levels along coral coastlines does indicate, though, that sea levels during the period 34 000 to 23 000 B.P. (i.e. during the growth of the last British Ice Sheet) were at least 35 m below modern levels (Bloom *et al.*, 1974; Harmon *et al.*, 1978, 1983). Furthermore, the oxygen isotope record and global sea-level changes are almost in phase (Chappell and Shackleton, 1986) and support low Mid Devensian sea-level stands (Chappell and Veeh, 1978). Thus, Eyles and McCabe (1989a) believe it is likely that the Irish Sea Glacier expanded across a basin with water depths lowered by a minimum of 35 m.

In a review of ice dynamics at a calving bay margin Hughes (1987) states that instability at marine margins of an ice sheet is a direct consequence of total uncoupling of ice from the underlying bed, to produce the 'pulling power*' of a floating ice shelf. This need not occur abruptly at the marine margin but may typically be initiated in marine ice streams, progressively giving the stream flow a concave long profile ('downdraw') (Hughes, 1987). The pulling power of marine ice streams is much greater because they move faster than terrestrial ice sheets, and the reduced basal shear stress along the ice stream allows the pulling force to reach far into a marine ice sheet (Hughes, 1987). The most destabilising factor at the margin of a marine ice sheet is when the grounding line retreats downslope into an isostatically depressed peripheral moat, and hence a deeper water column (Hughes, 1987). This has an effect of increasing the pulling power exponentially, following from the work of Weertman (1957) who showed that the pulling force increases as the square of ice thickness at the grounding line.

The geographic extent of subsidence around the downwarped margin of an ice sheet cannot be easily predicted but downwarping is known to extend for a distance up to twice the radius beyond the ice margin (Walcott, 1970). If a radius of about 500 km is taken for the British ice sheet (e.g. Boulton *et al.*, 1977) then the peripheral moat would include a substantial portion of southern Ireland, southern England, the south-western approaches and the English Channel (Eyles and McCabe, 1989a). From this, Eyles and McCabe (1989a) envisage that as ice retreated further back into this isostatically-downwarped moat, instability and rapid retreat was hastened as a result of progressively enlarged 'collapsing' of the Irish Sea ice sheet. Ice would become drawn to the zones of the fast ice flow, generating strongly convergent flowlines from which Eyles and McCabe (1989a) anticipate enhanced subglacial erosion in the form of tunnel valleys and drumlins (Section 2.6.2.1).

Overall, it is stated that ice-flow direction indicators within the Irish Sea Basin and surrounding areas (e.g. striae, drumlins, tunnel valleys) record fast centripetal ice flow into what would have been a rapidly collapsing marine portion of the Irish Sea ice sheet through the means of enhanced calving (Eyles and McCabe, 1989a). Other feedback mechanisms are thought to exist which would maintain the rapid velocity of an ice stream, together combining to produce the Jakobshavns Effect*. These mechanisms include ubiquitous crevassing, rapid rates of summer

* 'Pulling power' is defined by Hughes (1987, p. 184-5) as "...the product of the pulling force and the ice velocity. The pulling force is gravity acceleration acting horizontally on a mass of ice having a sloping surface and a buoyant bed. It increases as the square of the fraction of ice thickness supported fully by basal water pressure. Pulling power is greatest at the heads of ice streams, where surface slope is a maximum and where converging flow causes a rapid downslope increase in ice velocity".

* Hughes (1987, p. 185) defines the 'Jakobshavns Effect' as a "hypothetical mechanism for enhancing downdraw when an ice stream drainage basin has a large ablation zone, as observed in Jakobshavns Isbrae, located at 69°10′N, 50°10′W on the west-central margin of the Greenland Ice Sheet (Hughes, 1986). It postulates that summer meltwater pouring into surface crevasses, and subsequently reaching the bed, should increase basal water pressure and thereby allow the pulling power of the ice stream to reach further into the ice sheet".

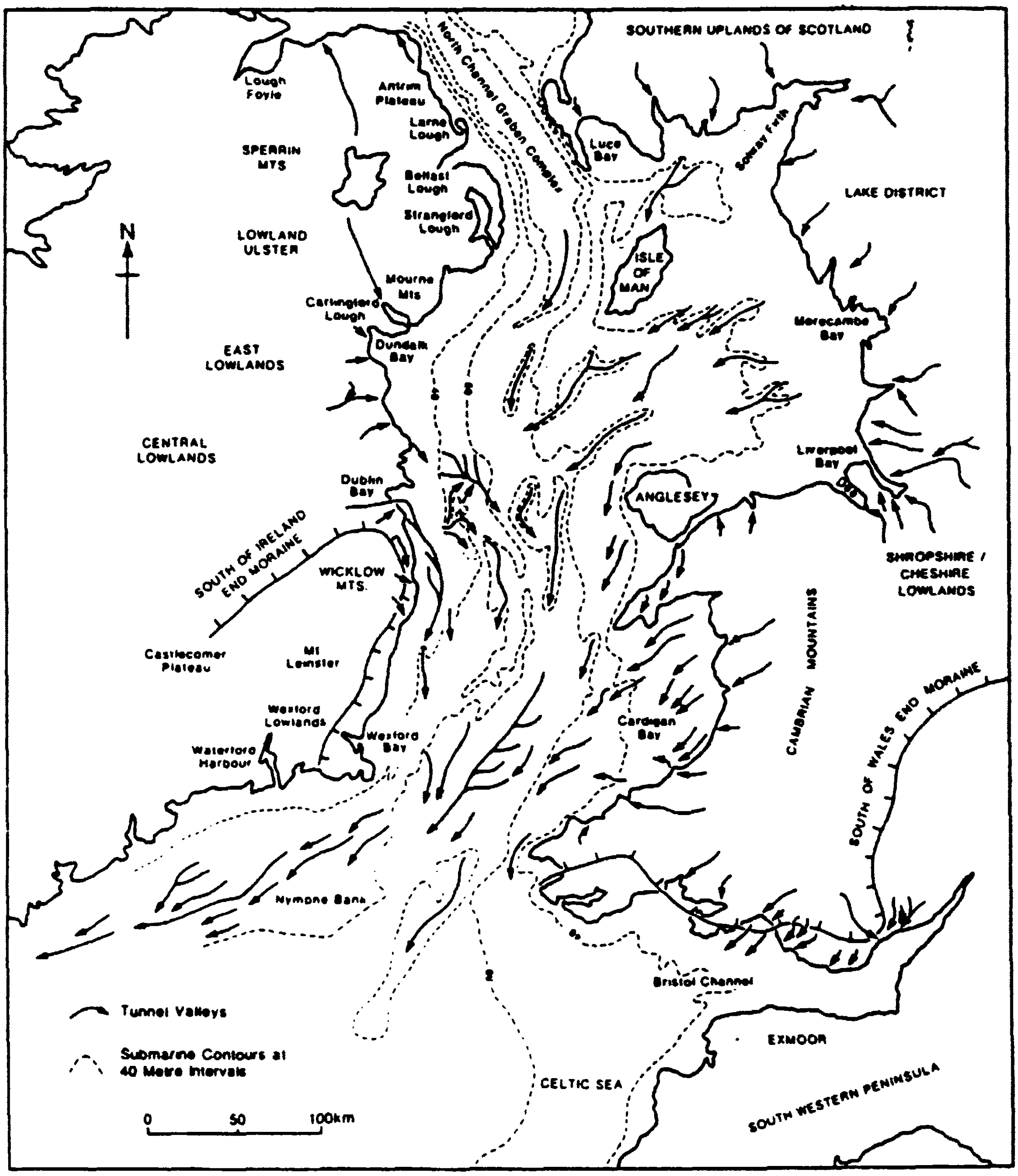

Figure 2.18 Subglacially-eroded tunnel valleys cut across bedrock on the floor of the Irish Sea Basin and adjacent coastal zones. Based on Admiralty Bathymetry charts, Institute of Geological Science charts, Dobson, Evans and Whittington (1973), Whittington (1977), Garrard (1977), Dobson (1977a), Blundell, Griffiths and King (1969), Garrard and Dobson (1974). (From Eyles and McCabe , 1989a).

melt over the crevassed surface, transport of supraglacial meltwater through crevasses to warm the ice internally and lubricate the ice-bed interface beneath an ice stream (causing an increase in basal sliding), and rapid iceberg calving promoted by tidal flexuring of a floating terminus (Hughes, 1987; Eyles and McCabe, 1989a).

Continued rapid ice retreat, following the 'unzipping' event, is thought to have revealed a scoured topography, in places showing a dense network of subglacially-cut tunnel valleys (Figure 2.18) which later were infilled by stratified proglacial glaciomarine sediments fed from temporarily stabilised tidewater ice margins (Eyles and McCabe, 1989a). Around the margin of the Irish Sea Basin, ice-contact glaciomarine sediments, represented in the form of large morainal bank complexes, are likely to have been deposited where the ice margin stabilised in shallow water (Eyles and McCabe, 1989a). Much of this sediment is thought to have been delivered to the ice margin by the tunnel valleys which drained the interior of the ice sheet (Eyles and McCabe, 1989a). As is discussed in Section 2.6 an encompassing theme has been developed by Eyles and McCabe (1989a) whereby within their glaciomarine model lies a generalised transitional relationship, on a geographic and genetic basis, between typically distinct depositional systems. They maintain that observation of these individual systems, identified from outcrop and subsurface studies and constrained by radiocarbon age dating and amino acid studies, strongly reflects the former position of the margin, mode of ice sheet retreat and deglacial ice dynamics.

2.5.4: Ice sheet readvances and margin oscillations in the north-east Irish Sea Basin

During the retreat from the Late Devensian maximum of the Irish Sea ice sheet, marginal ice readvances in the Irish Sea Basin were possibly common and have been claimed by various workers in Kirkham (Gresswell, 1967), the Isle of Man (Thomas, 1976, 1977, 1984a, 1985), Wexford (Huddart, 1977, 1981b; Thomas and Summers, 1983, 1984), Low Furness (Huddart *et al.*, 1977) and in Cumbria (Trotter, 1929; Trotter *et al.*, 1937; Huddart *et al.*, 1977; Huddart, 1991, 1994; Nirex, 1994). Thomas (1985) suggests that in all cases these readvances simply represent short-distance, localised oscillations of the snout. It is further claimed that they have no significant stratigraphic bearing and certainly do not represent ice sheet fluctuations driven by climatic change (Thomas, 1985b). Nevertheless, there have been many workers over the last century who have presented evidence to suggest that readvances did occur, many on a wide scale.

It is beyond the scope of this section to represent and discuss all documented readvances within the Irish Sea Basin during the deglacial phase (e.g. Gosforth Oscillation*; Trotter *et al.*, 1937). Instead, attention has been focused on the main readvances which are reputed to have occurred within the north-east sector of the Irish Sea Basin, in particular over south-west Scotland, north-west England and surrounding offshore areas.

2.5.4.1: The Scottish Readvance

A major area of controversy concerns the occurrence and status of several readvance phases of the main Late Devensian ice sheet which have been recognised across the Solway lowlands and the coastal area of west Cumbria. Opinions differ not only on the mode of deglaciation in this area (i.e. glaciomarine or terrestrial) but also on the presence or absence of evidence for glacial readvances. A recurrent theme in the Devensian glaciation of the northern Irish Sea basin is the concept of a major, so-called 'Scottish' readvance episode that occurred at a late stage in deglaciation (Huddart, 1994). The term 'Scottish Readvance Glaciation' was first established by the British Geological Survey in the 1920s to describe the final distinct episode of readvance of the main Late Devensian Irish Sea ice sheet from southern Scotland onto the Cumbrian lowland (Trotter, 1924, 1929; Trotter *et al.*, 1937). The concept of a Scottish readvance originates from the outcome of an attempt to subdivide Quaternary sequences occurring in western Cumbria (Trotter, 1929). A simple 'tripartite' stratigraphy was adopted for the glacigenic sequences which constituted a 'Lower Boulder Clay', 'Middle Sands and Gravels' and an 'Upper Boulder Clay', although an 'Upper Sands and Gravels' also occurred widely (Trotter, 1929; Hollingworth, 1931; Trotter and Hollingworth, 1932).

Deglaciation of the main Irish Sea ice sheet left widespread deposits of glaciofluvial sands and gravels in the form of kames and eskers, proglacial sandur, deltaic sands and gravels and glaciolacustrine silts and clays (Akhurst *et al.*, 1997). However, in a zone of about five to ten kilometres inland from the coast, these deposits show indications of having been overridden by later readvance(s) of ice. It is reported that they are capped by red, sandy diamictons, the 'upper till' (Upper Boulder Clay), which are locally extensive but laterally discontinuous on a regional scale (Trotter *et al.*, 1937). The 'upper till' has been variously interpreted as a deposit of a major readvance of Scottish ice (Trotter, 1929; Trotter *et al.*, 1937; Huddart *et al.*, 1977; Huddart, 1991, 1994; Nirex, 1994) or being a drape of glaciomarine mud deposited by meltwater plumes discharging from tidewater glaciers (raised glaciomarine deltas) during deglaciation (Eyles and McCabe, 1989a). Eyles and McCabe (1989a) envisaged high relative

* The reader is directed to Nirex, 1997. Sellafield geological and hydrogeological investigations: The Quaternary of the Sellafield area. Nirex Science Report, No. S/97/002 for a summary of information relating to the Gosforth Oscillation proposed by Trotter *et al.* (1937).

sea-levels (>150 m above O.D.) to account for the reported elevation of these deposits (Eastwood *et al.*, 1968). Huddart (1994) claims, however, that within the glacigenic sediments at this elevation there are no indisputable biostratigraphical or sedimentological indications of contemporaneous marine influence (Section 2.6).

Since the 1920s numerous workers have drawn many different, generalised lines for the marginal limits of this readvance (Figure 2.19). It can be deduced from this that the precise ice-marginal limits of the readvance are difficult to establish. This has resulted from the lack of a terminal morainic landform and deposition of only a thin till by the readvance ice, especially in the Carlisle Plain where the ice is thought to have spread out as a thin lobe into the topographically low terrain (Huddart, 1991). Trotter (1929) reported that the readvance extended over the Carlisle Plain up to the 400 m contour in the Brampton region, with the upper till plastered on the underlying deposits and no substantial modification to pre-existing landforms. These factors combined with the absence of any end moraine, progressive thinning of the upper till and preservation of the Middle Sands towards the outer limit, suggested that the readvance was short and of no great intensity (Huddart, 1991). Trotter and Hollingworth (1932) later traced the recessional deposits from the readvance and recognised that it impinged on the present day coast at St. Bees, Cumbria.

Trotter's (1924) limits in west Cumbria, the Carlisle Plain and the north coast of Ireland were extended by Charlesworth (1939) as the 'north-east Ireland – Isle of Man – Cumberland' moraine. Subsequently Charlesworth (1939) revised this line to correlate the Bride Moraine on the Isle of Man with the Carlingford Moraines of Ireland. Synge (1952) recognised a moraine which he traced from Brampton, south-west to St. Bees Head, out to sea and on to the Isle of Man in the Bride Hills. This ice limit was correlated with the Irish Kells Moraine, as both marked the southern limit of drumlin belts. Likewise, Mitchell (1960, 1963) identified moraines on the Irish Sea bed and linked these with moraines at Gormanstown (County Meath, southern Ireland), the Bride Hills and St. Bees. Penny (1964) suggested a line delimiting the Scottish Readvance in the Cumberland lowland which was merely the mapped feather-edge of the 'Upper Till'. Further south, a limit was also suggested by Gresswell (1967) who correlated the Kirkham Moraine with the Bride/Kells Moraines. Saunders (1968) suggested that the limit of the Scottish Readvance should be brought even further south and equated with the Dinas – Trevor – Brynkir Moraine in Llêyn, north Wales.

On the contrary, some workers have dismissed the need to invoke any glacial readvances during deglaciation of the Irish Sea ice sheet (e.g. Pennington, 1970; Evans and Arthurton, 1973; Sissons, 1974). Mitchell (1972) suggests that "many of our much fought over 'advances' and 'retreat-stages' are probably largely illusory, and depend as much on personal whim as on field evidence". The main source of argument presented by those opposed to a readvance is that the morphological evidence is unreliable (Richmond, 1959; Flint, 1970), for it is argued that a large morainic form may mark a readvance, a recessional stillstand, or even a terminal maximum. Likewise, the fundamental sedimentological criteria adopted by most workers is the stratigraphic position of the 'Main Glaciation' basal till and the Scottish Readvance till which provided an initial basis for assigning the two tills to distinct and separate periods (Huddart, 1991). Boulton (1972) suggests that the upper till need not automatically be of basal derivation but could feasibly have originated by ablation, flow or from an englacial deposition. Although the two till units are separated by 'Middle Sands' it does not mean that the sequence denotes a threefold ice advance-retreat-advance sequence (Boulton, 1972).

Furthermore Boulton (1977) also demonstrated, from work carried out in north-west Wales, that multiple till successions may arise from a single glacial episode without the need to invoke any readvance of the margin. Moreover, the point made by Huddart (1991, p. 159) appears valid: "the fact that there is an upper till in Cumbria will not stand by itself as evidence for a readvance". It is further believed that in many cases, where evidence does exist to suggest some form of overriding, this need not necessarily indicate a readvance on a wide extent (Thomas, 1985). Work carried out by Thomas (1976, 1977, 1984b) on the Isle of Man and in Ireland (Thomas and Summers, 1983, 1984) demonstrated that marginal readvance during the retreat of the Irish Sea ice sheet was most likely to have been endemic.

It is clear, therefore, that there exists a background of uncertainty as to the limits of a Scottish Readvance in northern Britain and even whether there is any evidence for such a readvance. Nevertheless in a re-evaluation of the concept, Huddart *et al.* (1977) argued that contrary to the views of many objectors there is substantive morphological and stratigraphic evidence to support a broadly correlative readvance limit between Ireland, the Isle of Man and Cumbria at some time between *c.* 13 000 and *c.* 15 000 yrs B.P. The approach used by Huddart was one which appeared to be reliable. It was based on a study of both landforms and stratigraphy in order to establish the depositional environment of any particular sedimentary association (the Landsystems approach; Section 3.3.1.4). In particular, this work established well defined proglacial depositional environments (Huddart, 1977). The purpose behind this was that if these proglacial morphostratigraphic units could be shown to have been associated with an advancing ice sheet rather than a retreating one and their marginal limits mapped, then the validity and extent of the readvance could be established (Huddart, 1977). The following depositional

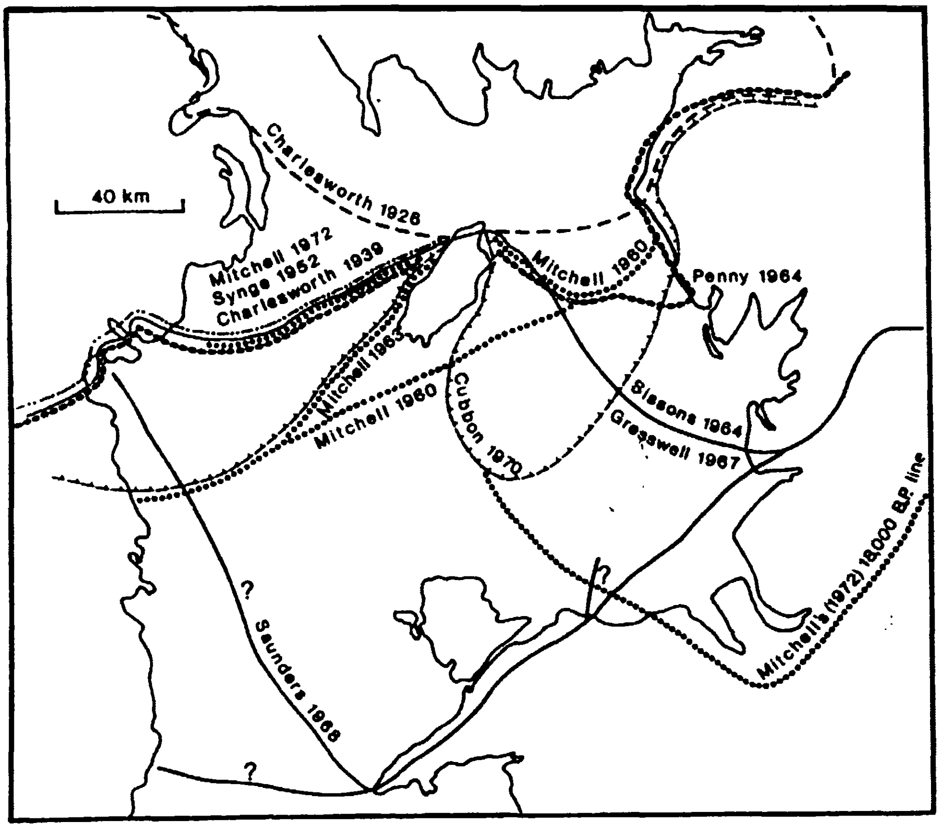

Figure 2.19 Postulated limits for the 'Scottish Readvance'. (From Huddart *et al.*, 1977).

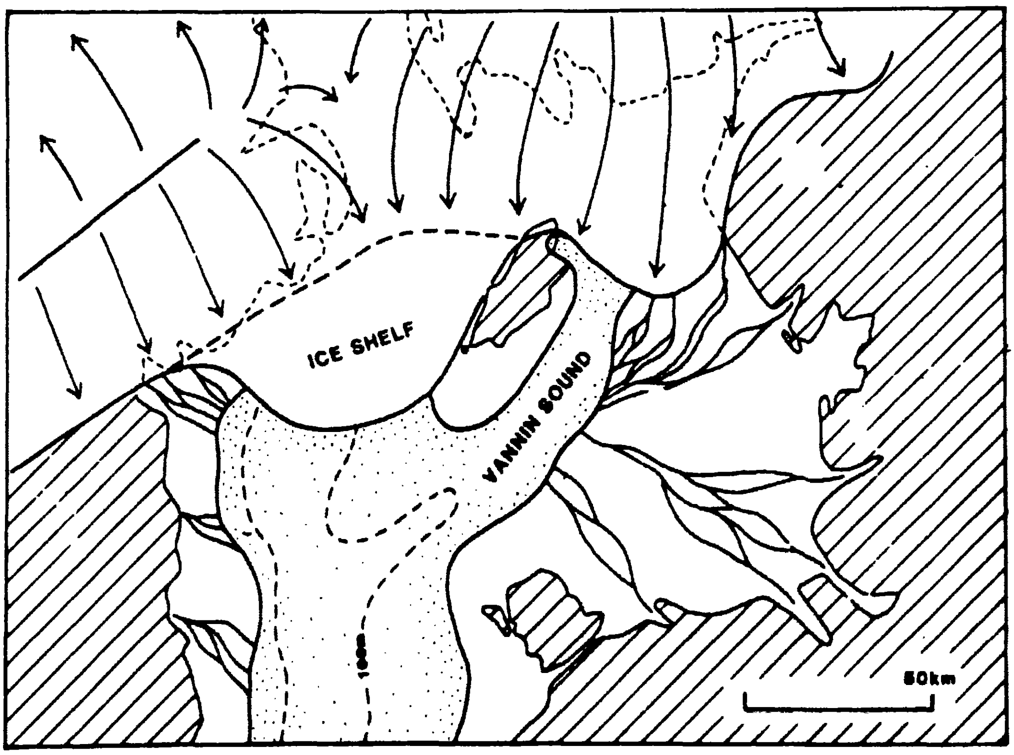

Figure 2.20 Tentative reconstruction of the northern Irish Sea basin around 15 000 years B.P. (From Thomas, 1985).

environments, reviewed in Huddart (1991), have been associated with a readvance of 'land-based' Irish Sea ice in the Cumbrian lowlands (the 'Scottish Readvance):

i. proglacial lacustrine deposit with overlying till in the eastern Carlisle Plain (Huddart, 1970).
ii. subglacial till in the western Carlisle plain (Huddart, 1970).
iii. proglacial sandur at Broomhills in the eastern Solway lowlands (Huddart, 1970).
iv. subglacial esker at Thursby in the eastern Solway lowlands (Huddart, 1973).
v. proglacial lacustrine deposit at Holme St. Cuthbert in the Solway lowlands (Huddart and Tooley, 1972).
vi. proglacial sandur at Harrington in west Cumbria (Huddart and Tooley, 1972).
vii. proglacial sandur and end moraine at St. Bees and between Nethertown and Seascale in west Cumbria (Huddart and Tooley, 1972; Huddart, 1977).
viii. proglacial lacustrine deposit in lower Wasdale (Huddart, 1970).
ix. proglacial sandur and end moraine along the Black Combe coastal lowland (Huddart and Tooley, 1972; Huddart, 1977).

Following the work of Huddart (1970, 1971a, b, 1991, 1994), Huddart and Tooley (1972) and Huddart *et al.* (1977), the concept of the 'Scottish Readvance' has gained renewed support from the more recent recognition of evidence for glaciotectonic deformation and glacial over-riding in the Whitehaven district, Cumbria (Nirex, 1993, 1995, 1997). This is seen to best effect along the narrow coastal strip affected by the Scottish Readvance where the style of glaciotectonic deformation is described as proglacial, as represented by a Glaciotectonic Thrust Terrain (Nirex, 1995). There, deposits have been subjected to severe proglacial thrusting, folding and faulting as revealed in the coastal section through the St. Bees Push Moraine (Nirex, 1993, 1995). A distinctive geomorphology is defined by elongated kettleholes and subparallel ridges, aligned north-east to south-west (Akhurst *et al.*, 1997).

It is currently believed, by those in support of the readvance model, that the whole northern Irish Sea basin was not deglaciated during the interval between the 'Main Glaciation' and the readvance as initially suggested by Trotter (1929). Additionally, there appears to be no evidence for the mechanism for such a readvance (Huddart, 1991). It is perceived, however, that a climatic change may have resulted in a greater snow accumulation during a period when the Southern Uplands was still ice covered. Correspondingly the Irish Sea ice sheet was reactivated and advanced an unknown distance onto the Cumbrian lowlands. This readvance was a major readvance stage in the general deglaciation of the Irish Sea basin (Huddart *et al.*, 1977).

2.5.4.2: Reconstruction of the northern Irish Sea Basin around 15 000 years B.P.

Offshore work by Pantin (1975, 1977, 1978) in the area between the Isle of Man and Cumbria has revealed proglacial deposits lying above till and beneath Holocene marine sediment. Pantin demonstrated that these proglacial sediments were deposited in a cold-water, low-salinity proglacial lagoon, termed the Vannin Sound, which was connected to the open sea. It is envisaged that this sea lay to the west, between the Isle of Man and Ireland, where the Irish Sea is deepest. The sea was partially separated from the lagoon by some sort of barrier and Pantin suggests that this may have been a floating ice-shelf possibly divided into two by the Isle of Man.

Independent of this work, Thomas (1976, 1977, 1985) has described a similar suite of deposits from the Dog Mills, Isle of Man ('Dog Mills Series') some 20 km north-west of Pantin's (1975) Vannin Sound. These deposits comprise alternating sequences of sand and mud, passing northwards into outwash sandur sediments fronting the Bride Moraine (Thomas, 1985). The depositional environment represented by these deposits is thought to be one of cold, low-salinity, estuarine-intertidal conditions with a current-swept open-water element (Thomas, 1977). Thomas (1985) suggested that the deposits of the Vannin Sound could conceivably be united with those from Dog Mills on the grounds that the latter represent either the shallow-water equivalent of the former.

Following on from this notion, Thomas (1985) suggested a tentative reconstruction of glacial conditions in the northern Irish Sea basin at approximately 15 000 yrs B.P. (Figure 2.20). To the east of the Isle of Man a narrow passage of the sea extended northwards to abut directly against an ice lobe margin running from the Bride Moraine towards Cumbria. Both to the east and west, where the ice margin rose against the bedrock highs of the Isle of Man and Cumbria, the margin was terrestrial and large volumes of proglacial outwash accumulated. To the west of the Isle of Man the ice margin is considered to have extended to the Kells moraine of eastern Ireland, with which the Bride Moraine has been commonly correlated (Charlesworth, 1939; Synge, 1952; Penny, 1964; Sissons, 1964; Gresswell, 1967; Mitchell, 1972; Stephens *et al.*, 1975; Stephens and McCabe, 1977; Section 2.5.4.1).

Huddart (1991) proposes that this reconstruction has implications for the validation of a Scottish Readvance model and states that the deposits described in Section 2.5.4.1 are "the result of this terrestrial deposition" (p. 166-67). Huddart (1991) admits, however, that the deposits found in the Cumbria lowlands do not appear to be the equivalent of the Dog Mills Series but nevertheless infers a similar proglacial depositional setting.

2.5.4.3: The Highland Readvance

In an early investigation, Charlesworth (1926b) described a kame-moraine extending along much of the northern edge of the Southern Uplands from the east coast south of St. Abb's Head to the Clyde above Lanark and the vicinity of New Cumnock, occurring again at Stranraer and in the Rhinns of Galloway in the west. This feature was termed the 'Lammermuir–Stranraer Moraine' and interpreted by Charlesworth as marking the former limit of a major readvance of Highland ice ('Highland Readvance') from the north. Charlesworth (1926b, p. 25-26) described the feature as:

> "great morainic belts...without doubt, the most conspicuous and easily recognisable and determinable of the drift formations. Over much of their extent they present the appearance of a rolling belt of country in which countless hillocks and hollows, ridges and troughs, succeed each other rapidly and tumultuously. They exhibit the bold, billowy relief of the swell and sag topography, with rapidly shifting curves and knobby and choppy surface, or a more subdued relief and gently undulating scenery. In general, they are kettle-moraines, pitted with numerous kettle-holes of varying depth and size, marked by small lakes, marshes, bogs, or meadows. These features show a tendency to parallelism with the course of the moraine".

In their constitution the moraines were seen to be variable, generally consisting of 'water-worn sands and gravels, the coarser and finer material preponderating in different places' and showed a special tendency to stratification (Charlesworth, 1926b).

Overall the impression was given by Charlesworth of a somewhat chaotic composition with many genetically different formations, occasionally 'decipherable', but usually and for the most part 'inextricable'. Nevertheless, Charlesworth states that despite their composition, variety of form, their topography and topographic relations these 'great accumulations' may be regarded as 'true moraines, in width, length, and form'. The term kame- or kettle-moraine was adopted by Charlesworth (1926b) to comprehensively describe a feature which was envisaged to have formed marginally to the ice limit demarcating its position at what was viewed as an important stage in glaciation. It is beyond the scope of this section to describe, in detail, the course of the moraine throughout southern Scotland. However, the reader is instead referred to Charlesworth (1926b, p. 27-44) where an in-depth description of the course of the moraine is provided together with an inferred nature of localised ice retreat. It is thought applicable, nevertheless, to describe the course of the moraine over the Galloway district, and to discuss the interpretations which have been developed to account for its existence.

The kame-moraine, as it exists, is described by Charlesworth (1926b) as sweeping round the head of Loch Ryan as a broad, hummocky belt, south of Stranraer. It then spreads under the grounds of Castle Kennedy [NX 111608] onto the foot of the hills east of Loch Ryan, where it becomes much less conspicuous as it is traced northward. Although it is described as a 'morainic belt' Charlesworth states that it is possible to single out individual ridges. Moraines corresponding to those at Stranraer were further observed in the Rhins of Galloway where one joins the Stranraer moraine, near Leswalt [NX 025640], and runs by Lochnaw Castle [NW 992627] to Balgracie [NW 987608], and south to Portslogan [NW 987587]. It appears, however, that the Lammermuir-Stranraer Moraine is entirely absent from the hillsides between Loch Ryan and New Cumnock to the north-east. It is believed by Charlesworth (1926b) that this break reflects the confluence, at this period, of the Highland Ice with the glaciers centred in the western part of the Southern Uplands. Further evidence which led to this hypothesis was the presence of a southern limit of Highland erratics in Ayrshire and Lanarkshire which were roughly coincident with the line of the kame-moraine. It is inferred from this that at an earlier pre-readvance stage the confluent Highland and Southern Upland ice masses cleaved somewhere in the Central Valley, well north of the moraine under discussion, the exact position being unknown and possibly indeterminable (Charlesworth, 1926b).

Following a retreat by the corresponding ice masses to the north and south, again of uncertain extent, a readvance occurred during which the Highland ice, except in the coastal areas, attained its farthest southern limit (Charlesworth, 1926b). At the period of maximum readvance, the Highland ice over South Ayrshire, west of New Cumnock, was still confluent with the local glaciers of the Southern Uplands. Moraines of this stage were in consequence absent and therefore accounted for the gap in the kame-moraine as outlined. Around Loch Ryan, however, a free edge of the Northern Ice again occurred and so the moraines are as 'conspicuously displayed as on any other section of the ice-front' (Charlesworth, 1926b).

Despite the attractiveness of this detailed hypothesis it is important to highlight this explanation as one which appears to be largely based on a number of assumptions, thereby potentially throwing into doubt the credibility of the reconstruction. In essence, the investigation fails to lend itself to the principle of Ockham's razor which is based on the fact that entities should not be multiplied beyond necessity.

2.5.4.4: The Southern Upland Readvance

In a subsequent paper, published at the same time, Charlesworth (1926b) further correlated what could be termed a 'Southern Upland Readvance' with the western portion of the kame-moraine marking the 'Highland Readvance' limit. During the retreat of the Southern Uplands ice it is believed that the ice split up into a series of more or less independent glaciers which occupied the major valleys in the region. The mode of recession of the ice during this period was investigated by Charlesworth (1926a) using moraines and outwash fans, developed in connection with all the major valley glaciers, as evidence of retreat positions. The former valley glaciers investigated occurred in the Cree, Ken, Nith, Annandale, Upper Clydesdale, Doon and Stinchar Valleys (Figure 2.13). The moraine positions and their corresponding descriptions are documented by Charlesworth (1926a, p. 8-17). An attempt was made to correlate the moraines of these widely separate regions in order to establish the limits of the glaciers at various stages of retreat. The Cree Glacier was chosen as the standard of comparison, as Charlesworth felt that here the moraines were quite distinct, sharply defined, and well distanced, and the relief of the bed of the glacier relatively simple. There were five stages of recession observed in connection with this glacier, each named after the locality at which the best moraines were found, which were subsequently used to cover all the corresponding moraines of the valley glaciers of this region: (a) the Monreith Stage; (b) the Kirkcowan Stage; (c) the Newton Stewart Stage; (d) the Minnoch Stage; (e) the Corrie Stage.

Charlesworth (1926a) latterly reflected upon the notion that some of these stages, possibly all, may have marked the limits of readvance of the Southern Upland ice. The clearest evidence for such a readvance was connected with the Kirkcowan stage, particularly in the region of the Cree glacier. Here, drumlins to the west of Wigtown Bay were noted to exhibit a striking discordance of trend, being divisible into two quite distinct systems, the drumlins of the one set being aligned north-east to south-west, or approximately so, while those of the other set ran north-northwest to south-southeast, or in other terms almost at right angles to the first drumlin set. The former set extended from a line trending roughly from Kirkcowan to Garliestown westward over the whole of the Machers to the eastern shores of Luce Bay. The latter set was confined solely to the east of this area (Charlesworth, 1926a).

In a closer examination of the distribution of these features, Charlesworth noted that the two sets of drumlins were mutually exclusive and that the line of delimitation was accurately coincident with the line of the outermost moraine of the Kirkcowan stage. Therefore, it was inferred that these two sets of drumlins, the one within, the other outside the moraine, demonstrated different directions of ice-flow over the two regions. Charlesworth (1926a) raised the point that it might be assumed that these two series were produced simultaneously. He proposed, however, that the complete absence of drumlins exhibiting intermediate trends rejected the view of successive formation of the features. Charlesworth (1926a) concluded, therefore, that it seemed necessary to envisage that the drumlins within and outside the moraine were the products of two 'distinct glaciations' separated by a retreat of uncertain magnitude.

The Lammermuir-Stranraer Moraine has, therefore, come to be applied to the combined limits of the Southern Upland and Highland ice at this stage. The moraine has subsequently been accepted as an important stage by a number of workers (e.g. Goodlet, 1970; Greig, 1971; Mitchell *et al.*, 1973) concerned with the glacial retreat stages in Scotland and has been variously correlated with stages in continental Europe. Sissons (1967) has cautiously noted, though, that despite the widespread acceptance of the Lammermuir-Stranraer stage the conclusions of almost all local workers other than Charlesworth are opposed to the existence of such a stage. Sissons (1963, 1964, 1967) had initially equated the Perth readvance to the Kirkcowan Stage. However, in a review which casts doubt upon the validity of a date supporting the concept of a Perth readvance, Sissons (1974, p. 314) states that "...there is at present no satisfactory evidence in Scotland that the decay of the last ice sheet was interrupted by a significant readvance". Cutler (1979), working in the Machars, carefully mapped the features suspected of being a readvance moraine and concluded that they were in fact the product of an ice standstill during final deglaciation rather than a readvance 'moraine', as previously interpreted by Charlesworth (1926b). Further support for this conclusion has been provided by Sissons (1974, 1981) and Price (1975, 1977).

Contrary to this, Huddart (1999) showed how a detailed sedimentological investigation of similar types of landforms in Nithsdale, Dumfriesshire need not imply a theory of glacier stagnation or a readvance. Huddart (1999) suggested that the origin of kames in this region are in fact the result of supraglacial trough fills associated with the decay of ice-cored ridges in a marginal environment. The glacier associated with such landforms is believed not to have been stagnant but instead fed meltwater and sediments to an ice front to the north of Dumfries, while to the south and south-east meltwater fed sandar systems in the Lochar Water and Nith valleys (Huddart, 1999). Overall, this recent investigation emphasises the obvious confusion which exists in interpreting this type of landform, particularly on occasions where sediment exposure is not good and conviction is instead placed on field mapping. The current work of Huddart and co-workers may have implications for the re-interpretation of similar glaciofluvial deposits in Galloway and Carrick.

Final retreat of ice from south-west Scotland is thought to have probably resulted in the ice becoming increasingly restricted to valley floors and basins. A consequence was that the ice became increasingly isolated from external sources and nourishment, hence its motion became increasingly sluggish until in most areas it became stagnant (Sissons, 1967).

2.5.4.5: The Loch Lomond Readvance and final decay of the ice sheet

It is widely recognised that following the decay of the Late Devensian ice sheet in Britain a short period of climatic deterioration occurred when glaciers built up on a much more restricted scale in Scotland, the Lake District and Wales (Sissons, 1979). This period is known as the Loch Lomond Stadial and is believed to have occurred between *c.* 11 000 and 10 000 yrs B.P. (Gray and Lowe, 1977). Over a wide area of the Scottish Highlands the limits of the Loch Lomond readvance have been mapped by numerous workers (e.g. Sissons *et al.*, 1973; Sissons, 1977). However, few detailed studies of a similar nature have been carried out in the Southern Uplands of south Scotland. Geikie (1901) was first to recognise that, following ice sheet decay, local glaciers developed in the hills surrounding the Loch Doon mass, while Jolly (1868) identified the end moraine by Loch Dungeon. Later, Charlesworth (1926a) drew attention to the moraines on the lower slopes of Kirriereoch and Merrick, equating them with his final 'Corrie Stage' (Section 2.5.4.4) and thus associated them with the initiation of small corrie glaciers in the Southern Uplands at this time.

Although limited evidence is available throughout the Southern Uplands, Charlesworth (1926a) states that the Loch Lomond readvance can only be detected where the direction of the ice-flow during the 'corrie stage' differed from that of the preceding stage, causing the corrie moraines to have a different composition from the other moraines of the same area. Furthermore, it is thought to be only those mountains which have the greatest elevation which attained the best developed corrie moraines (Charlesworth, 1926a). In the lower hills, the corrie glaciers are perceived to have extended for only a limited distance. Charlesworth (1926a) also noted that the corrie moraines were best developed in the corries facing north or east, e.g. in the Merrick (descending to 500 feet (Sissons, 1981)) and Kells ranges, the Lamachan Hills, Cairnsmore of Fleet, Cairnsmore of Carsphairn, and the Lowther Hills. In the centre of the Southern Uplands Sissons (1976) records the former existence of glaciers up to a few kilometres long as indicated by hummocky moraines, especially near the Loch Skene corrie. However, in the hills that surround the Loch Doon basin, high ground is not as extensive and consequently only small glaciers are thought to have developed in a few sheltered locations (Sissons, 1976).

Recently, Cornish (1981) has carried out a detailed reconstruction of glaciers of the Loch Lomond Stadial in the western Southern Uplands. Cornish infers the former existence of 11 glaciers near the summits of Merrick, Mullwharchar and the Kells range indicating a renewal of glacial conditions following the decay of the Late Devensian ice sheet. Reconstruction was based upon evidence from terminal, lateral and hummocky moraine and the distribution of erratics. The glaciers were all nourished on high ground (the majority in corries) with the lowest altitude occurring at 295 m. The firn line altitude, at its maximal extent, was calculated using an equation provided by Sissons (1974), and the values derived ranged from 328 m. OD to 670 m OD, with a mean value of 495 m.

2.6: A Conjectural Classification of Irish Sea Basin Depositional Systems

In Section 2.5.2.2 it was discussed how Thomas and Summers (1982, 1983) assigned a sequence of events to account for the nature of the stratigraphic successions typified in the coastal region of County Wexford, based on the assumption that these were derived from a land-based ice sheet. In turn, the generalised modes of deposition recognised by them were related to two principal environments which alone characterised deposition by terrestrial ice sheets in lowland areas. In a further attempt to identify a generalised stratigraphic relationship between individual depositional environments within the Irish Sea Basin, Eyles and McCabe (1989a) have since developed an encompassing theme whereby within their glaciomarine model (Section 2.5.3) lies a transitional relationship between what are argued to be distinct depositional settings* which are individually linked with a sequence of depositional events during the last deglacial phase (*c*. 20-14 ka B.P.).

2.6.1: Application of a depositional systems approach

The purpose behind the depositional systems approach is that it should provide a more realistic method of interpreting stratigraphic successions than the conventional 'layer cake' technique. A detailed account of the methodology is outlined in Section 3.3.1 however a general overview shall be provided at this point in order to discuss the development of the approach. Application of the latter has tended to concentrate solely on the vertical succession of lithological types, involving bed-to-bed correlations, and as a result has occasionally over-emphasised them (Eyles and McCabe, 1989a). This approach has been accommodated in the past primarily because Quaternary sediments have been commonly pictured as having accumulated by vertical aggradation. The inadequacies of this simple notion of stratigraphy have, however, been clearly demonstrated by comprehensive high resolution subsurface investigations of large sedimentary basins (e.g. Fisher and McGowen, 1967; Frazier, 1974; Casey, 1980) which provide detailed pictures of the broad-scale, three-dimensional stratigraphy of large sedimentary basins (Eyles and McCabe, 1989a). Mitchum *et* al. (1977), Brown and Fisher (1977), Winter (1979) and Miall (1984) have all demonstrated the importance of lateral accretion during deposition and stress the importance of recognising spatial changes in facies types and thickness resulting

* The term 'depositional setting' is used here by the author in the sense that it is thought to more aptly reflect a mode of deposition at a specific geographical location. It is preferred to use the phrase 'depositional system' as an umbrella term to incorporate a variety of depositional settings over a widespread region.

from sediment units developing out across a basin. This factor is particularly relevant when dealing with glacial environments which are viewed as possibly the most complex of all depositional environments, especially on a basin-wide context (Eyles and McCabe, 1989a).

The basis of the depositional systems approach is detailed facies analysis in a particular depositional setting, combining outcrop and subsurface data, followed by identification of depositional sequences and their three dimensional geometry and arrangement within the larger depositional system (Eyles and McCabe, 1989a, p. 313). Eyles and McCabe (1989a) stress that inherent in the approach is the recognition of the major depositional controls that have dictated the history of sedimentation (e.g. sea-level, tectonics, climate). Several Late Devensian depositional settings have been recognised in the Irish Sea Basin, through the application of this approach, and together they are seen to define a distinct event stratigraphy for the basin that describes the sequence of deglaciation and records changes of relative sea-level caused by ice sheet loading and crustal downwarping (Eyles and McCabe, 1989a). The following section does not attempt to provide a comprehensive account of all the major depositional settings within the Irish Sea Basin as this would prove to be lengthy. However, it is intended that a number of critical field sites will be examined which have been used to derive the present classification system adopted by Eyles and McCabe (1989a). In doing so, it should be brought to the attention of the reader that a number of sites worked upon by avid supporters of the glaciomarine concept have been subject to criticism on the basis of perceived erroneous interpretations. Areas where conflicting arguments exist are discussed in Harris and McCarroll (1990), Harris (1991), Austin and McCarroll (1992), McCarroll and Harris (1992) and Wingfield (1992).

2.6.2: A provisional classification of depositional systems

The schematic diagram represented in Figure 2.21 provides a provisional classification, as interpreted by Eyles and McCabe (1989a), of Late Devensian depositional settings/systems, within the northern and central Irish Sea Basin identified from outcrop and subsurface studies, as discussed above, and constrained by radiocarbon dating and amino acid studies. In summary, the diagram reflects the stratigraphic relationship between two major depositional systems, subglacial and glaciomarine, in which the former deposits, consisting of drumlins and other streamlined landforms, are pictured as 'interfingering' along the current coastal margin with glaciomarine deposits. A marked feature identified on the coastal margin are large morainal bank complexes, consisting of arcuate belts of hummocky topography overlying ice-contact glaciomarine stratigraphies, on the seaward side of drumlin belts (Eyles and McCabe, 1989a). These pass offshore into what have been classified as stratified glaciomarine complexes infilling tunnel valleys and overlying earlier subglacial sediments in the Irish Sea Basin (Eyles and Eyles, 1984; McCabe *et al.*, 1987; Eyles and McCabe, 1989a, b, 1991; McCabe and Haynes, 1996). Eyles and McCabe (1989a) note that a marked feature aiding this classification is that many depositional settings still retain much of their original geomorphological integrity, both onshore and offshore, as a result of the comparatively small time frame considered in what is viewed as a short-lived deglaciation lasting no more than a few thousand years

2.6.2.1: Subglacial depositional setting

Although subglacial deposits within the Irish Sea Basin are, in general, poorly understood (Eyles and McCabe, 1989a) limited drill core data presented by Garrard and Dobson (1974), Garrard (1977), Delantey and Whittington (1977) and Whittington (1977) has revealed complexes of coarse and fine-grained sediments and till. Eyles and McCabe (1989a) have inferred from the character of offshore subglacial 'heterogeneous' sediments that they were deposited at or beyond the ice margin, and subsequently overrun by the ice sheet as it expanded across the Irish Sea Basin. Further to this, they picture this advance as occurring over a partially flooded basin which maintained a wide range of water depths.

Along the coastal margins of the Irish Sea Basin insufficient evidence is currently available in which to portray the widespread occurrence of the subglacial depositional system. Eyles and McCabe (1989a) suggest that non-deposition, and possible removal, in these areas is most likely to be attributed to steep bedrock relief along many coastal stretches, such as in Wales, coupled with intense erosion in high energy proglacial marine settings. In contrast, however, prolific expanses of onshore subglacial landforms are prevalent throughout low relief coastal margins, in particular the coastal margins of Ireland, Wales, north-west England and south-west Scotland. Drumlins are the dominant subglacial landform assemblage and have been noted by many workers (e.g. Charlesworth, 1926a; Hollingworth, 1931; Hill, 1973; Hill and Prior, 1968; Huddart and Tooley, 1972; Stephens and McCabe, 1977) to overlie strongly ice-moulded bedrock. Moreover, their combined presence is seen to record centripetal Late Devensian ice flow into the Irish Sea Basin as earlier discussed in Sections 2.4.2 and 2.5.3.

A wide range of subglacial facies associations have been described, modelled and interpreted in Northern Ireland (e.g. Dardis and McCabe, 1983; Dardis *et al.*, 1984; Dardis, 1985; Dardis and McCabe, 1987) to which the reader is referred for detailed accounts. In sum, most investigations revealed a complex variation of sequences both in composition and internal structure, but common to many was the presence of stratified sequences. Facies modelling has been used in the majority of cases to resolve the complexity of the units into corresponding

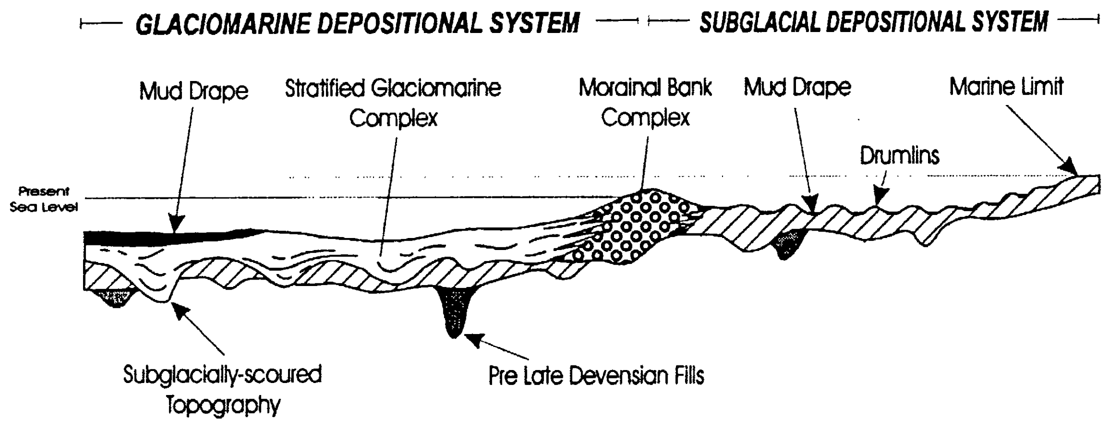

Figure 2.21 Late Devensian Irish Sea Basin depositional systems indicating the generalised stratigraphic relationship between the subglacial depositional system and the glaciomarine depositional system. (Modified from Eyles and McCabe, 1989a).

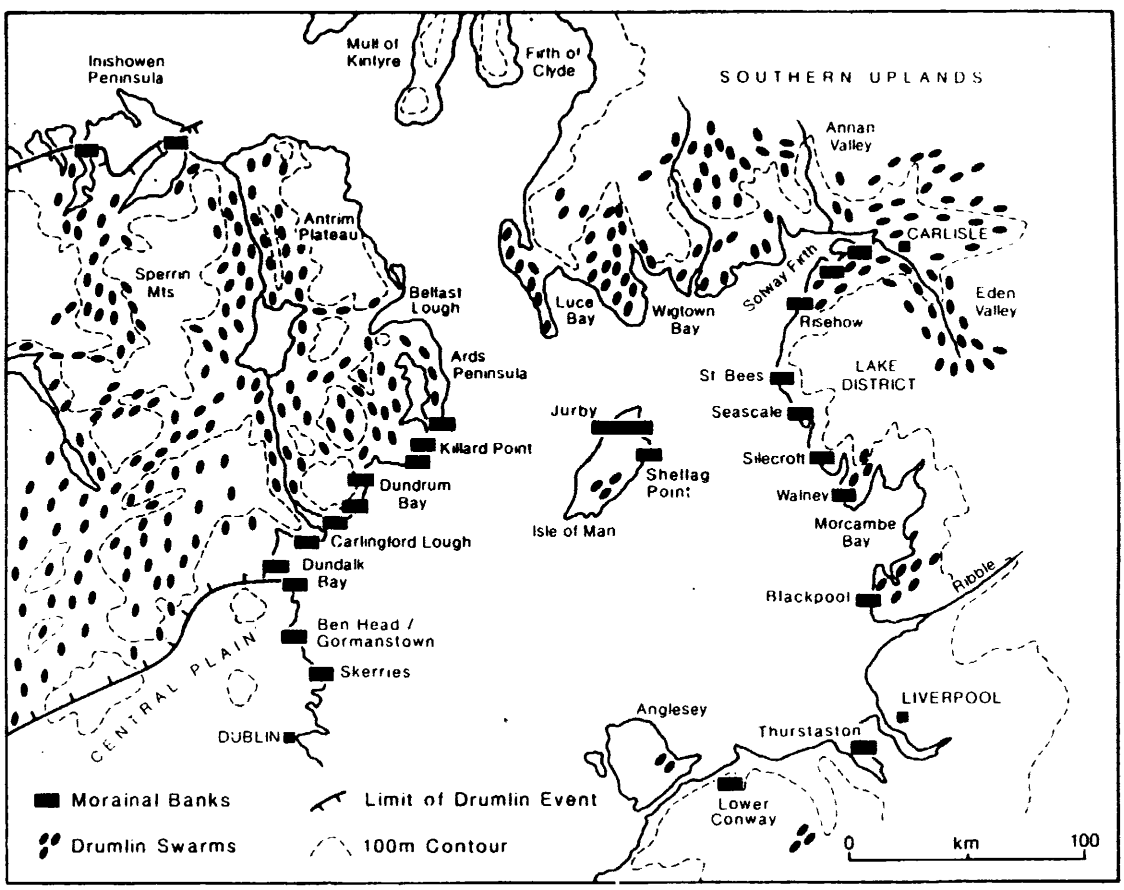

Figure 2.22 Distribution of drumlins and ice-contact morainal banks in the Irish Sea Basin. (From Eyles and McCabe, 1989a).

facies associations which were seen to reflect either a particular depositional process or palaeoenvironment. Five major facies have been identified from drumlins in northern and western parts of Ireland (Dardis and McCabe, 1983, 1987; Dardis *et* al., 1984; Dardis, 1985; McCabe, 1985). In these studies it was concluded that drumlins cannot be viewed from classical models of erosion or deposition (McCabe, 1987).

It is shown to be evident from facies analysis that drumlin formation can instead be linked to a series of temporal, sequential and spatial changes at or near the ice-substrate interface (McCabe, 1987). Dardis (1985) suggests that neither subglacial sediment deformation, squeezing of detritus into basal ice cavities or basal deposition is responsible for drumlinisation. Instead, Dardis prefers to invoke the role played by hydrologic processes in drumlin evolution in the final deglacial phase. It is suggested by Dardis and McCabe (1983), Dardis *et al.* (1984) and Dardis (1985) that the infilling and blockage (sedimentation) of interconnected subglacial drainage networks, resulting in a reduction in hydraulic transmissibility at the ice-substrate interface, constitutes an integral part to drumlinisation. It is considered that large quantities of meltwater become stored at the ice-substrate interface, with resulting high basal meltwater pressures initiating a total uncoupling, leading to uplift of the ice (cf. Röthlisberger and Iken, 1981). In this scenario, drumlinisation is pictured as occurring under surge-type flow conditions associated with periods of high subglacial meltwater evacuation, during which time streamlining is prevalent (Dardis, 1985). An extension of this idea is that large volumes of subglacial sediment, involving either a deformable bed or at least erosion below fast moving ice streams, are delivered to the ice margin (Eyles and McCabe, 1989a).

The spatial distribution of drumlins has been analysed at several scales. Drumlins tend to be concentrated in fields, often numbering up to several thousand individuals where associated with large former ice sheets (e.g. Laurentide ice sheet) (Vernon, 1966; Shilts *et al.*, 1987; Aylsworth ans Shilts, 1989a, b). Drumlin fields commonly form broad bands, aligned either transverse or parallel to former ice flow directions. Transverse fields may reflect zones of drumlin formation behind former ice margins, whereas longitudinal drumlin fields may mark the position of fast ice streams or glacier lobes in former ice sheets (Wright, 1957, 1962; Rose and Letzer, 1977; Jones, 1982; Piotrowski and Smalley, 1987; Dyke and Morris, 1988; Patterson and Hooke, 1996). In the British Isles, drumlins are most common in lowland areas or in large valley systems which act as conduits for the excavation of ice from the Late Devensian ice sheets (e.g. Hollingworth, 1931; Clapperton, 1970; Mitchell, 1994). Drumlin distribution within a swarm may appear to be regular, but is more likely to be random (Smalley and Unwin, 1968). However, apparent non-random distributions have also been identified which have been explained as reflecting one of banding of bed properties (Smalley and Piotrowski, 1987; Smalley and Warburton, 1994) or the influence of pre-existing variations in small-scale topography and substrate properties (Boulton, 1987).

Several studies have assessed characteristics such as drumlin spacing, density and distribution, and their variation within and between fields (e.g. Jewtuchowicz, 1956; Gravenor and Meneley, 1958; Vernon, 1966; Doornkamp and King, 1971; Hill, 1973; Crozier, 1975). Drumlin spacing, defined as the perpendicular distance between adjacent drumlins (Reed *et al.*, 1962), may exhibit normal or multi-modal distributions for individual fields, and is highly variable between fields. Menzies (1979) invokes that such inter-field variability suggests that the glacial processes responsible for drumlin formation differ according to location. Drumlin density, the number of drumlins per unit area, varies between regions but may also vary within an individual field (Reed *et al.*, 1962; Vernon, 1966; Hill, 1973). This is suggested to possibly decrease or increase in a downglacier direction (e.g. Fairchild, 1929; Charlesworth, 1957; Smalley and Unwin, 1968; Hill, 1973). Evidence has also been presented which suggests that drumlins within a swarm may not be synchronously deposited as is often assumed. Synge and Stephens (1960), Vernon (1966) and Hill (1971) have suggested time-transgressive (diachronous) deposition of drumlins within individual drumlin fields in Ireland. What is, however, generally accepted is that a common orientation between drumlins almost certainly pertains to a single flow event responsible for their development and that cross-cutting relations, where in existence, indicate a response to changes in the overall shape of the ice sheet (discussed fully in Section 3.2.1.2).

In the north and west of Ireland, the almost continuous drumlin presence on low ground is well constrained and this has been used in the past, to identify a Lateglacial 'Drumlin Readvance' (Synge, 1968, 1969). In preliminary work, in the western parts of the county of Mayo, Synge (1968) was struck by the apparent weak effects of glaciation on the north-west region of the county. A prominent end-moraine was subsequently identified and mapped across north Mayo, from Ballycastle to Mulrany (Synge, 1968). This is regarded as marking a later phase of renewed activity, possibly due to a climatic deterioration (McCabe, 1973; Davies and Stephens, 1978) or to a surge (Dardis *et al.*, 1984; McCabe, 1986), following the 'main glaciation' with drumlin formation being associated with the active advance of this ice (Synge, 1968). Retreat moraines inside this limit are limited due to what is thought to have been a rather inactive mode of retreat, and instead are replaced by extensive tracts of dead-ice topography and streamlined bedforms (Synge, 1968). Estimates of the age of drumlin-forming ice vary from 19 000-13 000 yrs B.P. (Stephens and McCabe, 1977) to 16 000 - 15 000 yrs B.P. (Synge, 1977). Glaciomarine muds deposited penecontemporaneously with marine delta-moraines at the limit of the Drumlin Readvance in northern County Mayo were radiocarbon dated to 16 940 and 17 300 yrs B.P. (McCabe *et al.*, 1986).

2.6.2.2: Glaciomarine depositional setting

MORAINAL BANK COMPLEXES

Prevalent along the eastern seaboard of Ireland and the Isle of Man are large moraine complexes up to 150 km^2 in area, with less extensive deposits preserved along the higher relief Lancashire and Cumbrian coast of England (Eyles and McCabe, 1989a) (Figure 2.22). These marked topographic features consist of elongate masses of sediment which are believed to reflect ice-contact marine deposition at the retreating margins of tidewater glaciers (Eyles and McCabe, 1989a). It has been documented that primary sediment delivery to morainal banks can be derived from several sources (Powell, 1981; Powell and Molnia, 1989; Powell and Domack, 1995): (a) Supraglacial debris can slump or fall down the ice margin cliff into the water; (b) Englacial and subglacial debris layers melt out below the water line, releasing detritus directly into the water; (c) Debris is dispersed over a wider area by iceberg melt-out and overturn; (d) Unfrozen sediment can emerge from beneath the glacier margin, either as the result of squeezing or as the output from subglacial deforming layers (Andrews, 1963; Smith, 1990; Benn, 1996); (e) Vast quantities of sediment can be delivered to the ice margin by subaqueous meltwater discharge. In addition, sediment can be deposited on the proximal side of morainal banks by subglacial processes such as lodgement and/or deposition from a subglacial deforming layer (Holdsworth, 1973; Barnett and Holdsworth, 1974; Benn, 1996). At oscillating ice margins, morainal banks can also be developed or modified by ice-push or thrusting (Powell, 1981; Boulton, 1986). It is understandable, therefore, that previous and current investigations indicate that a wide range of glaciomarine depositional styles do exist. This has promoted the development of a number of depositional models which are regarded purely as local summaries alone, and thus fail to simulate scenarios in other settings (e.g. McCabe *et al.*, 1984, p. 727; McCabe *et al.*, 1987, p. 481).

The term ice-contact morainal bank was originally adopted by Powell (1981) to describe relatively small bouldery ridges at the foot of tidewater valley glacier termini along Alaskan fjords. The term has been broadened by Eyles and McCabe (1989a, p. 316) to incorporate what they describe as "arcuate belts of hummocky, undulating topography, sometimes several kilometres in width, marking the former positions of ice sheet margins reaching tidewater". Each complex is typically associated with bedrock highs and relatively shallow water and is thought to possibly record a temporary halt or re-equilibration (Andrews, 1973; Hillaire-Marcel *et* al., 1981) of the Irish Sea Glacier margin (Eyles and McCabe, 1989a). The concept of ice-contact marine deposition, which reflects a temporary stabilisation of the retreating ice margin, has been derived from a number of detailed investigations.

Eyles and Eyles (1984) identified a subaerial exposure of Devensian marine and glaciomarine sediments found on what was interpreted as a raised marine foreland on the northern part of the Isle of Man in the central Irish Sea Basin. These sequences are thought to represent the largest exposure of emergent Upper Pleistocene glaciomarine deposits so far identified on the British continental shelf (Eyles and Eyles, 1984). Across the foreland occurs a large arcuate push ridge up to 100 m high and 0.6 km wide (the Bride moraine) which identifies a major ice limit. Thomas (1977) has speculated that this moraine is a 're-equilibration moraine' (Andrews, 1973; Hillaire-Marcel *et al.*, 1981) resulting from the advance of a formerly floating ice margin that became grounded. The push-moraine ridge is reported to divide the stratigraphy of the foreland into offlapping stratified and massive sandy diamict assemblages that occurred subaqueously over a basement of glaciotectonised marine sediments to the north, from a coarsening-upward marine sequence of pelagic muds with dropstones, sand, and gravel to the south (Eyles and Eyles, 1984). Further descriptions of sections and associated structures have been described by Slater (1931), Thomas (1977), Thomas and Summers (1984) and Eyles and McCabe (1991). A push-ridge–subaqueous-outwash (submarine) depositional setting (Figure 2.23) has been suggested whereby processes of suspension deposition, density underflow, ice rafting, variable traction current activity, and sediment gravity flow were operative in the vicinity of the grounded marine ice margin of the retreating Irish Sea Glacier (Eyles and Eyles, 1984). It is thought that the presence of large arcuate submarine banks offshore, having a relief up to 30 m and aligned parallel to the Bride moraine, might suggest that the construction of morainal banks was later repeated during ice recession (Eyles and Eyles, 1984).

Detailed studies have shown that morainal bank complexes are closely related to surging, drumlinisation of the bed and the delivery of large volumes of subglacial debris to ice-frontal morainal banks. It has been stressed by Dardis (1985), Dardis and McCabe (1983), Dardis *et* al. (1984), McCabe (1986) and Boulton and Hindmarsh (1987) that subglacial transport of unlithified over-pressured sediments is an important mechanism during drumlinisation. This idea may be extended to incorporate a deformable bed, or at least fast flowing ice streams, whereby large volumes of subglacial sediment are delivered to the ice margin. As previously mentioned, however, morainal bank stratigraphies are normally not floored by subglacially-deposited tills because it is believed that such sediments were either not deposited, or were

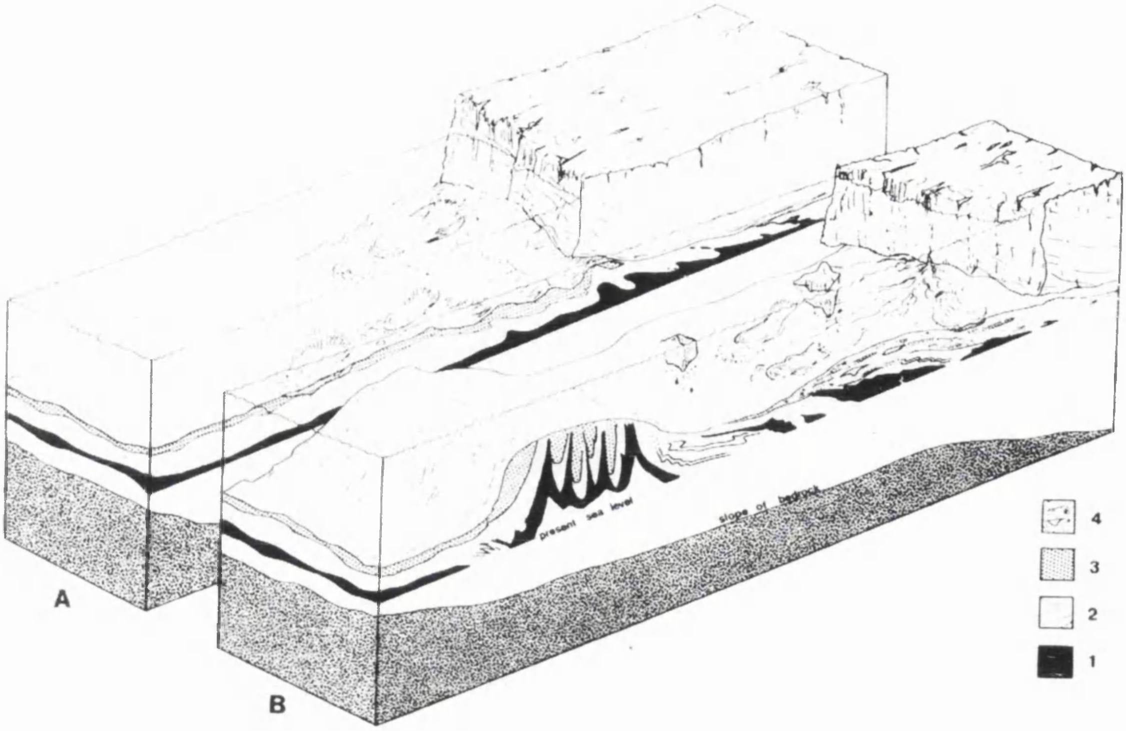

Figure 2.23 Simplified schematic model for deposition of "push-ridge–subaqueous-outwash" depositional system as exposed on the east coast of the Isle of Man. Suspended sediment plumes are omitted. 1=Fine-grained diamict assemblage; 2=shallow marine sandstone; 3=gravel; 4=coarse-grained diamict assemblages. Below sea-level, gravel, sandy diamicts, and shallow-marine sand and silt rest on a sloping rock surface (Smith, 1930) (From Eyles and Eyles, 1984).

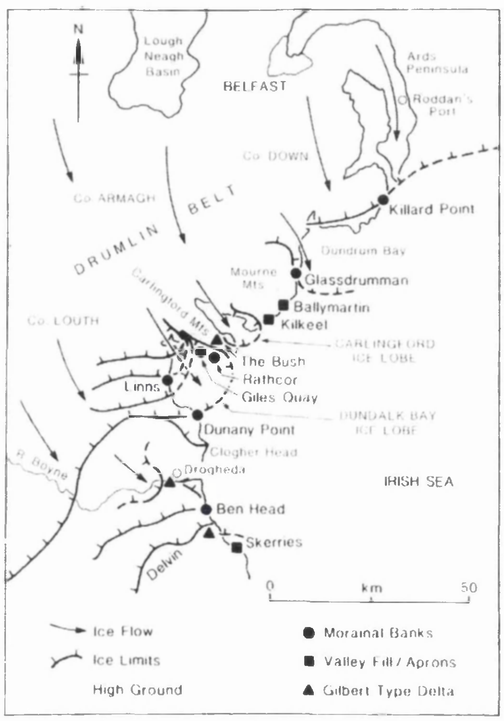

Figure 2.24 Morainal banks on the margins of the inland drumlin belts in north-east Ireland. (From Eyles and McCabe, 1989a).

scoured out by high ice velocities and meltwaters as the ice margin retreated (McCabe *et al.*, 1984; Eyles and McCabe, 1989a).

McCabe *et al.* (1984, 1987) and McCabe (1985, 1987) have described in detail the development of several ice-contact morainal banks along the eastern seaboard of Ireland, particularly along a 50 km stretch between Killard Point and Dunany Point in east central Ireland (Figure 2.24). In this area, tidewater sequences are restricted to the southern coastal margin of an extensive belt of drumlinised terrain. Typically these sequences consist of complexly interbedded deposits (diamictons, stratified gravels, sands and muds) which have, until recently, been explained in terms of land-based, terrestrial models. However, recent reappraisals of some of these sequences indicate that the deglacial palaeoenvironment was far more complex than initially perceived. McCabe *et al.* (1984) have attributed the sequences of the moraine at Killard Point, formed at about 16 000 yrs B.P., to submarine remobilisation and transport of glacigenic debris from an ice sheet/ice shelf grounded on the isostatically depressed coastal lowlands of County Down, Northern Ireland, at a time of relatively high sea level. The Killard Point moraine is one of a series of coastal ridges which are believed to mark former ice-marginal positions during northwestward retreat of Late Pleistocene ice in east-central Ireland (Stephens and McCabe, 1977). The moraine morphology is described as one of 'subdued hummocks and gentle billows with no marked crest orientations (local relief <15 m)'. The general absence of distinct kettle holes and ridges is noted to contrast markedly with moraine morphology farther inland. It is inferred, on the basis of raised-beach features, washing limits, and regional drapes of red marine clay (20-25 m O.D.), that such subdued topography resulted from erosion during high, Lateglacial sea-levels (Stephens and McCabe, 1977).

The Killard deposits are described as sedimentologically complex and may be grouped into three major lithologic associations: interbedded diamictons (="tills"), gravel, and sand (McCabe *et al.*, 1984). Such complexity has been attributed to remobilisation and transport of glacigenic debris from a grounded ice sheet/ice shelf junction into deeper water (McCabe *et al.*, 1984) (Table 2.4). The deposits found at Killard Point have been described as 'unfossiliferous' which in all depositional settings, where subaqueous material exists, often provides confusion over whether they represent deposition into a large lake rather than the sea (Miall, 1983). Nevertheless three suppositions, which have evolved from similar scenarios, have been used by McCabe *et al.* (1984) to support a glaciomarine origin in this instance:

i. Retreat of the ice margin in the Irish Sea Basin was accompanied by high sea levels (Stephens and McCabe, 1977; Synge, 1977).

ii. Presence of ice barriers across certain coastal strips prevented penetration of marine waters inland at a time of high, Lateglacial sea level (Stephens and McCabe, 1977; Synge, 1977).

iii. Marine fossils occur in associated but deeper-water glaciomarine sequences which formed in ice-marginal situations similar to Killard Point (Colhoun and McCabe, 1973; Stephens *et al.*, 1975). The absence of marine populations at the locality is thought probably to relate to a number of factors: (1) the length of time for species migration towards recently deglacierised areas (Shackleton, 1987); (2) the cold, turbid conditions that would have prevailed; (3) the lack of sunlight penetration (Carey and Ahmad, 1960); and (4) meltwater input.

Table 2.4 Lithological associations in the Killard Point moraine, eastern seaboard of Ireland.

Lithologic Association	*Major Lithofacies*	*Minor Lithofacies*	*Interpretation*
Gravel	BGmm, BGmg BGcg, BGcm PGmm, PGcm	Fm, Fd, St	High-density sediment-gravity flows
Sand	Fl, Sh, Sm	Fm, Ssd	Turbidity currents and sediment-gravity flows of low to intermediate viscosity
Diamicton	Dmm, Dms	Fm, Fd BGcg (i)	Remobilisation of glacigenic debris. Slumping and cohesive debris flows.

Source: McCabe *et al.* 1984. *Summary of glaciomarine lithofacies types*: **Diamicton (D)** (Poorly sorted, mud-sand-gravel admixture) Dmm: Matrix-supported, massive, structureless. Dms: Stratified, occasionally graded, matrix-to-clast-supported, evidence of winnowing. **Gravel (BG**=boulder/cobble grade or PG=pebble grade) BGmm: Matrix-supported, massive, unstratified, matrix from granules to pebbles. BGmg: Matrix-supported, stratified. BGcm: As BGmg, but clast-to-matrix supported. BGcg: As BGmg, but clast-supported. **Sand (S)** (Fine to medium grade) Sm: Massive, unstratified. Sh: Parallel laminations or partings. St: Cross-stratified. Ssd: Soft-sediment deformation. **Mud (F)** (Fine-grained units, mainly silt and clay) Fm: Massive mud, unstratified. Fl: Clay, silt, and sand laminae, alternating beds. Fd: Laminated pebbly mudstone or laminated mud with clasts.

The largest (150 km^2) tract of proglacial marine sediments in the Irish Sea Basin is exposed in coastal sections up to 70 m high and nearly 20 km long in the Screen Hills of southeast Ireland. This complex deposit provides a detailed example of proximal to distal facies variations within a large ice-contact 'braid-delta' (McPherson *et al.*, 1987, 1988) prograding into the marine environment from a grounded ice margin. The Screen Hills comprise a large area of rolling hummocky relief, close to the former maximum extent of Irish Sea ice in the basin, underlain by a coarsening-upward succession of mud, sand, gravel and diamict facies. The hummocky terrain is the result of glacitectonism (Thomas and Summers, 1984) involving the development of ice-thrust ridges combined with the large-scale gravitational loading of gravels into underlying marine muds (Eyles and McCabe, 1991).

Lowermost marine muds exposed along the coast are massive, commonly weakly-laminated, contain dropstones and a mixture of reworked cold water molluscs *in situ* and foraminifera (Macmillan, 1964; Thomas and Kerr, 1987). This mud unit has been traced south of the Screen Hills into the Wexford area (Culleton, 1977, 1978) where at Ely House it is overlain by stratified facies containing large amounts of ice-rafted debris. As noted in Section 2.5.2.1, these facies were interpreted as glaciomarine by Lewis (1894), Huddart (1977) and Synge (1977) but were later identified as lodgement and waterlaid till by Thomas and Summers (1983). To the north, the marine muds have been observed to pass into prodelta laminated muds and sands with abundant ice-rafted debris (Eyles and McCabe, 1989a). The presence of wave-ripples, reversing palaeocurrents and trace fossils all testify to a marine (possibly tidal) influence in a shallow water environment (Eyles and McCabe, 1989a). These shallow water facies can, themselves, be traced northwards into thick (50+ m) southward dipping units of sand and gravels which correspondingly interfinger with, and pass laterally into, thicker braided-river gravel sequences displaying unidirectional current indicators to the south (Eyles and McCabe, 1989a). In all, there is clear indication that a complete transition has arisen from marine to subaerial settings as revealed from the type and distribution of facies along the coastal margin of the Screen Hills, from marine muds in the south passing north into delta front and braid plain deposits. This is consistent with a large southward prograding glacier-fed braid delta (Thomas and Summers, 1983, 1984).

STRATIFIED GLACIOMARINE COMPLEXES

Across stretches of the Irish Sea Basin, a thick (sometimes over 100 m) well-stratified sediment drape infilling a scoured subglacial topography is displayed. The occurrence of this complex was first documented by Whittington (1977) following an investigation across part of the Irish Sea Basin, using a continuous seismic profiling technique, which identified this particular unit among three other distinct units making up the subsurface deposits. The lowest unit is that of pre-Pleistocene bedrock which is known to be of varying age (Whittington, 1977). The second unit is subglacially-deposited till, with the interface between it and the underlying bedrock reflecting erosional activity. The softer lithologies display considerable overdeepenings while the more resistant lithologies are believed to exist as upstanding blocks (Whittington, 1977). This unit has been correlated from the Kish Bank area into Cardigan Bay and the South Irish Sea where boreholes have shown the unit to consist of Devensian Irish Sea till (Eyles and McCabe, 1989a). The third unit, of particular importance here, is characterised by internal strongly reflecting horizons along general persistent horizontal beds as further discussed below (Whittington, 1977). Where the topography is rugged this well-stratified sediment follows the topography, slumping into topographic lows and drape thinning over high points (Whittington, 1977). Overall, no erosional phases have been recognised within this unit and therefore it is attributed entirely to a depositional origin. The topmost unit consists of the banks and other sand wave bodies, and recent muds and silts which infill some hollows (Whittington, 1977).

The base of the third depositional unit was mapped using contours at 20 m intervals and revealed that these extensive, well-stratified sediments infill a complex scoured valley system (Whittington, 1977). Seismic records across the western and eastern platforms of the Irish Sea Basin have shown that this topography consists of narrow (<2 km), steep-sided sinuous valleys that are locally overdeepened to as much as 130 m OD and are over 100 km in length (Eyles and McCabe, 1989a). The valleys are now identified as tunnel valleys cut by subglacial meltwater and debris under high hydrostatic heads (e.g. Grube, 1983; Ehlers, 1983; Ehlers *et al.*, 1984; Cameron *et al.*, 1987; Boulton and Hindmarsh, 1987). They form an extensive rectilinear network, possibly reflecting block faulting patterns. In places they have been traced inland as buried bedrock-floored 'preglacial' channels for many tens of kilometres (e.g. Lancashire, Dublin and south Wales coasts) (Jones, 1942; Gresswell, 1951, 1964, 1967; Griffiths *et al.*, 1961; Bowen and Gregory, 1965; Crampton, 1966; Blundell *et al.*, 1969; John, 1972; Howell, 1973; Garrard, 1977; Culver and Bull, 1979). Substantial quantities of sediment filling the offshore valley system is thought to suggest that they drained from the coastal margins of the Irish Sea Basin to the central trough (Eyles and McCabe, 1989a). Eyles and McCabe (1989a) suggest that the most probable source is a tidewater ice margin anchored around the coastal margins of the Irish Sea Basin delivering large volumes of sediment to the marine environment.

Coastal exposures of Late Pleistocene sediments deposited after 19 000 yrs B.P. at Bray, near Dublin, Ireland, have provided a window into the stratified seismic infill of a subglacially-cut tunnel valley (Eyles and McCabe, 1989a, b). Sections located close to a steep sidewall of a tunnel valley show gravel and cobble-filled multi-storey U-shaped channels cut and filled by powerful subglacial meltwaters driven by high hydrostatic head (Eyles and McCabe, 1989 a, b). Gravels are blanketed by poorly sorted ice-proximal glaciomarine sediments (muds and ice-rafted debris, subaqueous outwash, sediment gravity flow facies) that Eyles and McCabe (1989a, b) believe record the pumping of large volumes of subglacial debris along the tunnel valley to a tidewater ice sheet margin. Cutting and filling of very large tunnel valleys, such as this, through the means of a dynamic hydraulic system, have been linked to areas of fast ice flow within ice sheets (Boulton and Hindmarsh, 1987).

RAISED MARINE DELTAS AND VALLEY INFILL COMPLEXES

The coastal margins of the Irish Sea Basin reportedly display clear examples of marine deltas graded to former high sea levels in the Irish Sea (Figure 2.25). The best example of high level

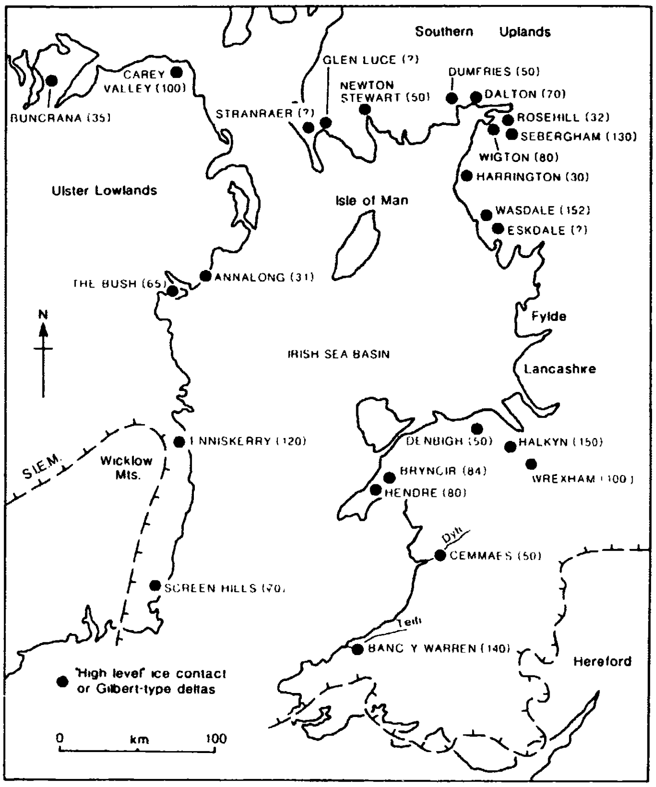

Figure 2.25 High level raised delta deposits around the margins of the Irish Sea Basin. Numbers in parentheses refer to elevations in metres above sea level. In most cases these features are located in coastal valleys and are thought to represent the positions where ice streams re-equilibrated and anchored after rapid wastage in open glaciomarine conditions. (From Eyles and McCabe, 1989a).

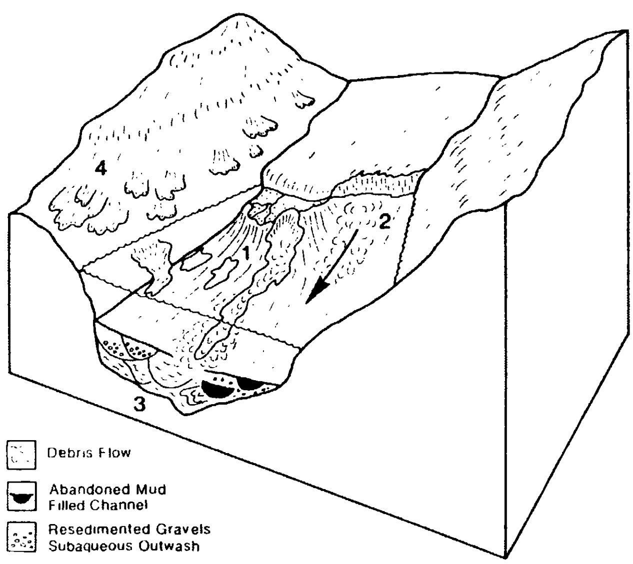

Figure 2.26 Sedimentation along a high relief coastal margin during rapid deglaciation. (1) Unstable valley-head delta subject to downslope failure. (2) Calving tidewater ice margin issuing subaqueous outwash and debris flow facies. Turbulent flows erode and fill channels with sediment gravity flow facies. (3) Multi-storey nested channels cut and filled by sediment gravity flows; mud filled channels record abandonment and infilling with plume muds. (4) Subaerial mass wasting. (From Eyles and McCabe, 1989a).

deltas are thought to be those at the southern exits of cols located along the topographic spine of the Llyn peninsula in North Wales. In this area, a number of extensive 'flat-topped kames' have been recorded by Saunders (1968) forming part of a discontinuous line of fluvioglacial deltaic deposits that have been interpreted as the record of a succession of glacial lakes ponded between ice and the peninsula. It has been inferred that the cols along the spine of the peninsula probably acted as meltwater channels from ice impinging on the northern flanks of the peninsula and subsequently fed large Gilbert-type deltas and sedimentologically-complex spreads of sand and gravel to the south. Well-preserved delta surfaces at diminishing heights of 84, 70 and 49 m OD (Saunders, 1968) have been interpreted as marine in origin thereby providing clear indicators of crustal rebound and falling relative sea-level.

To the south of these deltaic complexes there has been no geomorphic evidence reported of ice margins that may have dammed lakes against the peninsula (Eyles and McCabe, 1989a). Instead, sediments have been discovered at a lower elevation revealing a composition of laminated and massive muds and sands that drape the underlying bedrock and which are probably of a prodelta marine origin (Eyles and McCabe, 1989a). Similar topographic settings and high-level marine delta deposits have been reported on the margins of the Lake District (Eastwood *et al.*, 1968) and the Wicklow Mountains (Farrington, 1944).

Huddart and Tooley (1972) and Bowen and Lear (1984) have noted that along the high relief coastal margins of Cumberland and Wales are sections which show ice-contact glaciomarine complexes preserved in confined bedrock valleys. These valley infill complexes are thought to be the result of downvalley resedimentation of unstable piles of heterogeneous sediments from the margins of glaciers reaching tidewater (Eyles and McCabe, 1989a; Figure 2.26). The offshore extent of such bedrock valleys is unknown but it is suggested that they may represent the onshore portions of subglacial tunnel valleys mapped offshore (Eyles and McCabe, 1989a). Excellent exposures have been documented from sites in central Wales, on the northern Llyn Peninsula and at St. Bees in Cumberland (De Rance, 1871a, b; Smith, 1912, 1931; Synge, 1963, 1964; Saunders, 1968; Whittow and Ball, 1970; Huddart and Tooley, 1972; Huddart *et al.*, 1977). Examples of valley fill complexes exposed in confined topographic lows have also been found along the north Welsh coast (Synge, 1963, 1964; Saunders, 1968, 1973; Whittow and Ball, 1970).

MUD DRAPE

Across the Irish Sea Basin, thick distal glaciomarine muds are judged to have accumulated diachronously as the ice margin withdrew north towards the central Irish Sea Basin and stabilised along coastal margins (Eyles and McCabe, 1989a). This interpretation stems from the

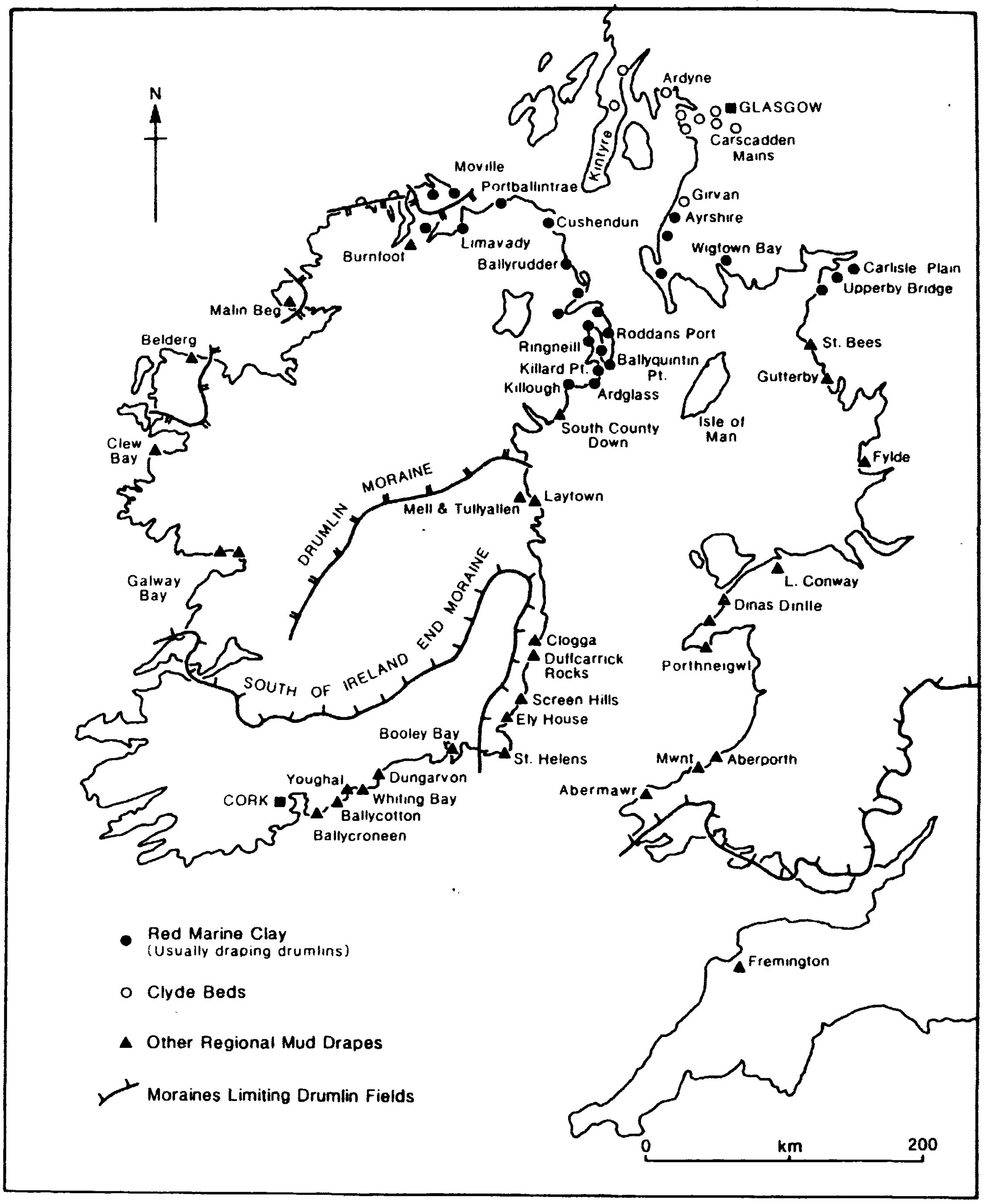

Figure 2.27 Coastal exposures of Late Devensian distal glaciomarine mud drape facies in the Irish Sea Basin and surrounding areas. (From Eyles and McCabe, 1989a).

presence of fine-grained distal glaciomarine facies which are widely exposed in coastal exposures above sea-level around the Irish Sea Basin. Commonly these deposits overlie ice-proximal morainal banks or delataic facies reflecting the position of the depositional site as it became more ice-distal. A number of workers have, nevertheless, classified the mud drape as an Irish Sea Till despite its consistent drape-like geometry; the presence of lamination and diffuse banding, intimate association with marine sediments and an extremely fine-grained texture (70 -90% clay) (Eyles and McCabe, 1989a).

Drumlins along the north-east coast of Ireland are reportedly draped by red marine clay up to 20 m OD (Stephens, 1963, 1968; Morrison and Stephens, 1965; Stephens and Synge, 1965), while a similar occurrence has been noted among drumlins from north-west England. In an account of the geology of northern England, Taylor *et al.* (1971) states:

> "In a final phase of the glaciation, a thin reddish till-like stony clay was deposited in low-lying parts of the Solway Basin and along the coast of north-west Cumberland. The clay swathes pre-existing drumlins and outwash deposits, the tops of which are undisturbed".

Eyles and McCabe (1989a) examined this drape in the northern Lake District, present up to 140 m OD, and found it to show a distal glaciomarine mud, which can be traced around the basin from North Devon and Southern Ireland to Scotland (Figure 2.27). In west Scotland, fine-grained, late glacial marine sequences are recorded at 41 m OD in the inner Clyde Valley (the Clyde Beds of Peacock *et al.*, 1977, 1978; Peacock, 1981). The widespread occurrence of mud drape facies has led Eyles and McCabe (1989a) to invoke the possibility of a correlation with the Errol Beds of the North Sea Basin which accumulated as the Late Devensian ice sheet retreated after 18 ka B.P. (Peacock, 1981).

2.7: Late Devensian and Holocene Relative Sea-Level History of the Irish Sea Basin

The Late Devensian deglaciation of Great Britain has produced a complex pattern of sea-level change around the British Isles (Lambeck, 1993a). This has been qualitatively understood in terms of two components: crustal rebound in response to the melting of the former local ice sheet, and the rise in sea-level produced by the melting of the more voluminous ice sheets over North America and northern Europe (Lambeck, 1993a). A considerable body of information is now available which quantifies the positions of past sea-levels both above and below present level. Evidence for past sea levels, expressed relative to the present level, around the shores of Great Britain come from a variety of different geomorphological and geological indicators. Among such forms of evidence are the raised beaches of eastern Scotland (Sissons, 1967, 1983), the platforms and raised beaches of western Scotland (Gray, 1974; Dawson, 1988), and the now submerged fresh-water peats of regions such as Morecambe Bay (Tooley, 1978), the Fenlands (Shennan, 1986), and the Thames Estuary (Devoy, 1982). These records clearly illustrate the interplay between local crustal rebound and global sea-level rise (Lambeck, 1993a). Significant is the spatial variability in sea-level response which has been recognised at these individual localities with areas further south, away from the main impact of glaciation, reflecting more of a response to global changes in the last ice sheets than to regional changes in the British ice sheet. A comprehensive summary of observational evidence for Lateglacial and Holocene postglacial sea-level change can be found in Lambeck (1993b). Figure 2.28 illustrates some of the observed sea-level curves (reduced to mean sea level) from principal observation sites for Scotland, England and Wales.

Attempts to explain the trends in relative sea-level during the last deglaciation in the British Isles, centred upon the Irish Sea Basin, have been made through the development of two contrasting models: (1) a numerical model (Lambeck, 1993a, b); and (2) a glacio-sedimentary model (Eyles and McCabe, 1989a; McCabe, 1997). Both models assess, to varying extents, the role played by a range of environmental factors that are seen to operate on different spatial and temporal scales, and differ in their approach in that they place certain emphasis on particular factors (Andrews, 1978, 1987). These include glacial eustasy, glacial isostasy, symmetry and area of ice sheets, timing and decay of ice sheets, location of ice-marginal positions during the decay cycle, crustal responses and unloading history (McCabe, 1997).

2.7.1: Numerical sea-level models

Numerical models are based on physical assumptions including the effective lithospheric thickness (rigidity), the effective mantle viscosity, ice thicknesses and earth response to loading and unloading (Lambeck, 1995). They predict low, eustatic sea-level trends in the Irish Sea Basin well into the deglacial cycle agreeing with the terrestrial model of deglaciation (Section 2.5.2). The downfall of numerical modelling is that the method relies heavily for model verification on matching sea-level predictions with mainly Holocene sea-level curves due to shortcomings in quantitative Late Devensian sea-level evidence. This is despite considerable spatial variability in qualitative observational records around the British Isles coastline (Lambeck, 1993a, b, 1995, 1996). Sea-level indicators are seen to provide an important data set for estimating the mantle response to changes in surface loading produced by the melting of the Late Devensian ice sheet over the British Isles, and by additional contribution to the rise in sea-level by the melting of the much more voluminous ice sheets of Fennoscandia and Laurentia (Lambeck, 1993b).

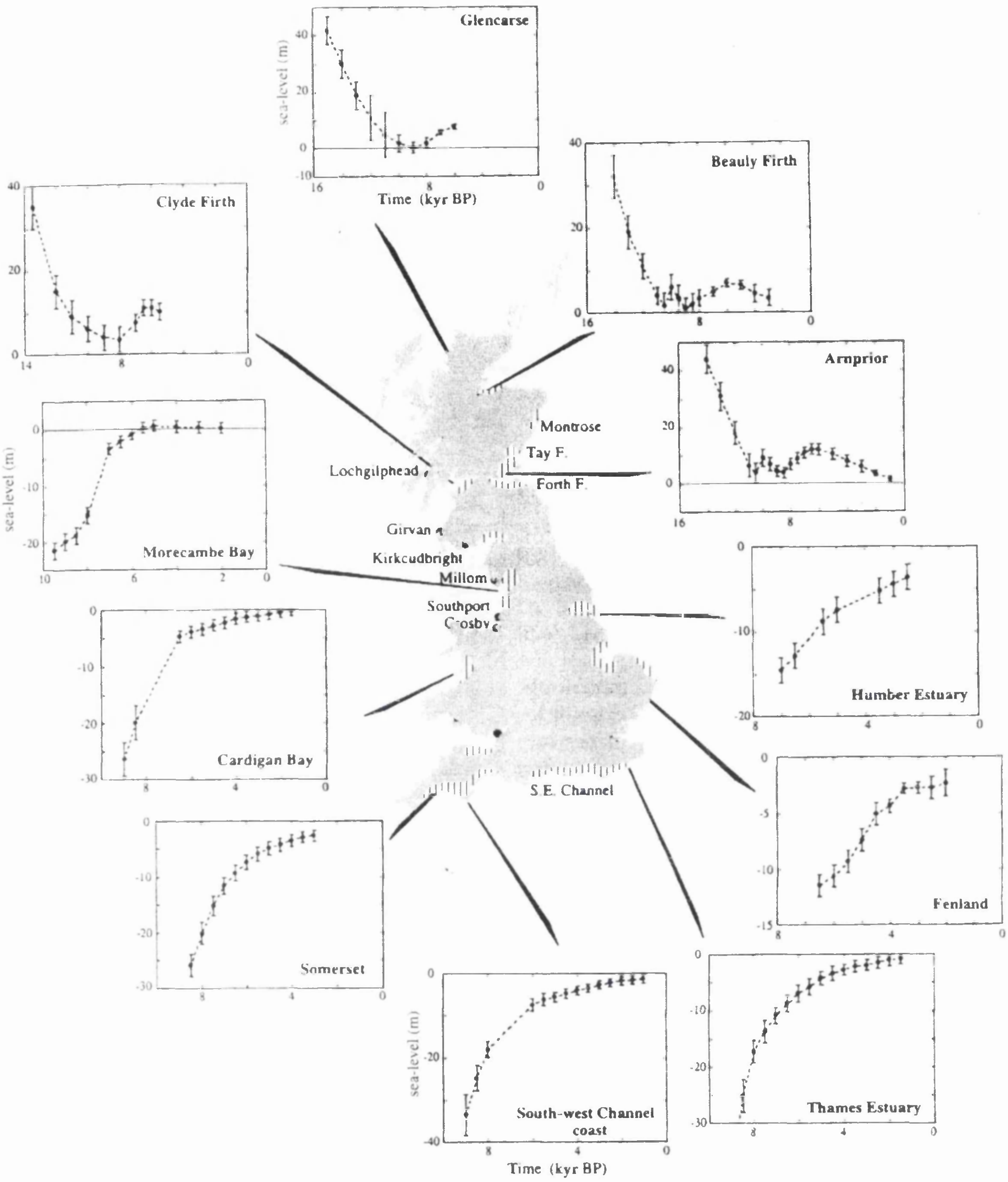

Figure 2.28 Postglacial sea-level curves from around the coast of Britain, demonstrating the influence of glacioisostasy in the north (emergence) and glacioeustasy in the south (submergence). (From Lambeck, 1995).

At best, the sea-level record for the Late Devensian and Holocene is based on data brought together from different localities, collected by different investigators and corresponds with different responses to changes in sea-level. The quantitative evidence used from south-west Scotland in the most recent numerical model (Lambeck, 1993b) has been based upon a review of the work by Jardine (1971, 1975) and Bishop and Coope (1977) which is discussed within Section 1.3.3. Lambeck (1993b) has used this observational data to compile two estimated sea-level curves for the east Solway Firth (Redkirk, Annan) and the west Solway Firth (Wigtown Bay and Luce Bay) (Figure 2.29).

The results from numerical modelling generally predict sea-level changes which mirror those for north-eastern and eastern Ireland, reflecting distance from the centre of rebound of the sites from north to south (Lambeck, 1996). Figure 2.30 illustrates the spatial pattern of sea-level change across the region for several epochs according to the model produced by Lambeck (1996). They are based on the same numerical model as used in earlier reconstructions (Lambeck, 1993a, b) except that a correction to the eustatic change in Late Holocene time has been applied. These results can be compared with Figure 2.31 taken from Lambeck (1995) which are based on the same earth model and an ice model identical in all respects except that the minimum ice thickness over Ireland is 500 m as compared with 600 m in Figure 2.30. Illustrated in Figure 2.30 are (i) the ice limits and the ice heights used in the model, (ii) the contours of sea-level change with respect to present sea-level, and (iii) the approximate positions of the shoreline. Differences between these results and those of earlier models mainly concerns the predicted locations of the palaeo-shorelines.

2.7.2: Sedimentary sea-level models

Glacio-sedimentary models apply a different approach to the reconstruction of Lateglacial sea-levels in that they are principally based on stratigraphic and dating evidence which emphasise the role of substantial glacioisostatic depression, high relative sea-levels during the deglacial cycle, and the part played by the Irish Sea Glacier as proposed by the glaciomarine model detailed in Section 2.5.3 (Eyles and McCabe, 1989a; McCabe, 1997). Eyles and McCabe (1989a) describe three phases of sea-level change within the Irish Sea Basin starting from deglaciation of the Late Devensian ice sheet from *c.* 17 ka B.P.:

Phase 1.Relative falls of sea-level from +90 m O.D. on the Cardigan Bay coast and +150 m O.D. and higher on north Irish Sea coasts.

Phase 2.Rising sea-levels starting in the south with the Phase 1 falls reaching relative levels of below -55 m O.D. on the north Celtic Sea coasts during the pre-Boreal (Stillman,

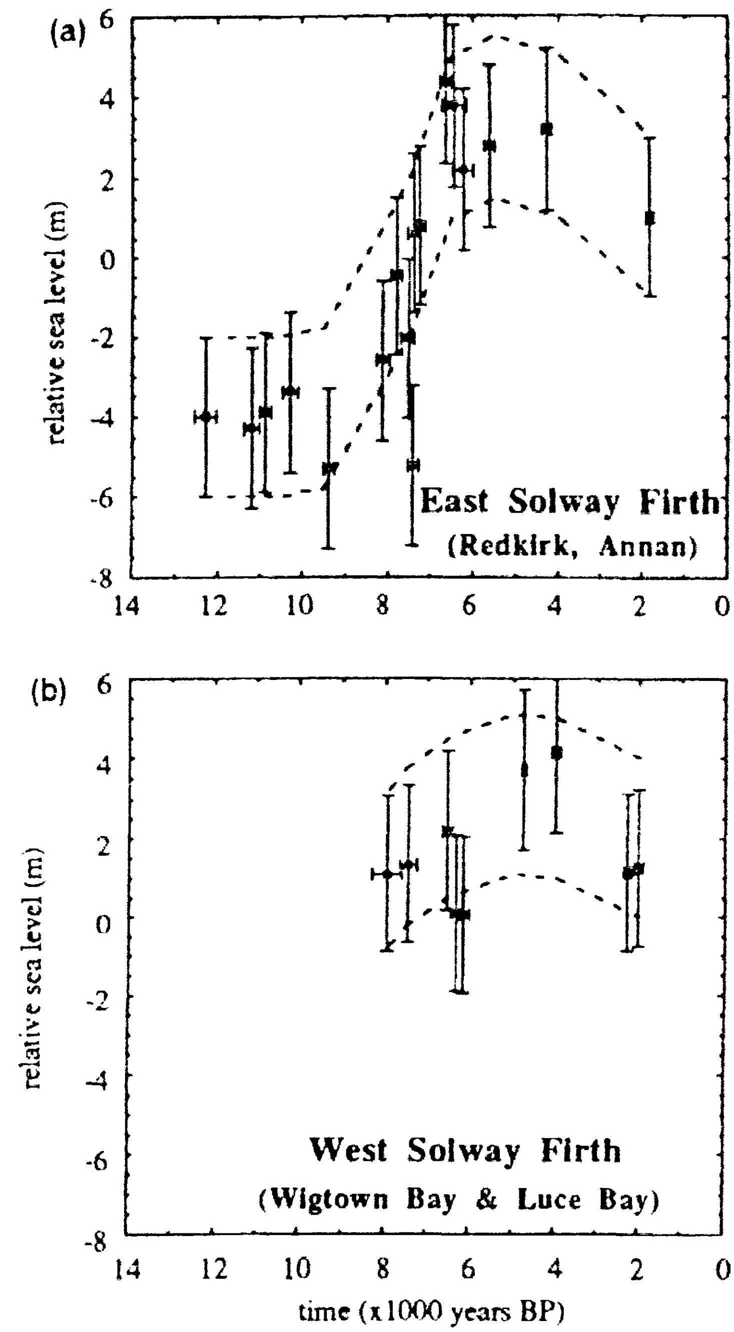

Figure 2.29 Age-height relations of sea-level indicators and estimated sea-level curves, all reduced to MSL, for south-west Scotland. (a) East Solway Firth near Annan and Redkirk point, (b) the west Solway Firth near Luce Bay and Wigtown. (From Lambeck, 1993b).

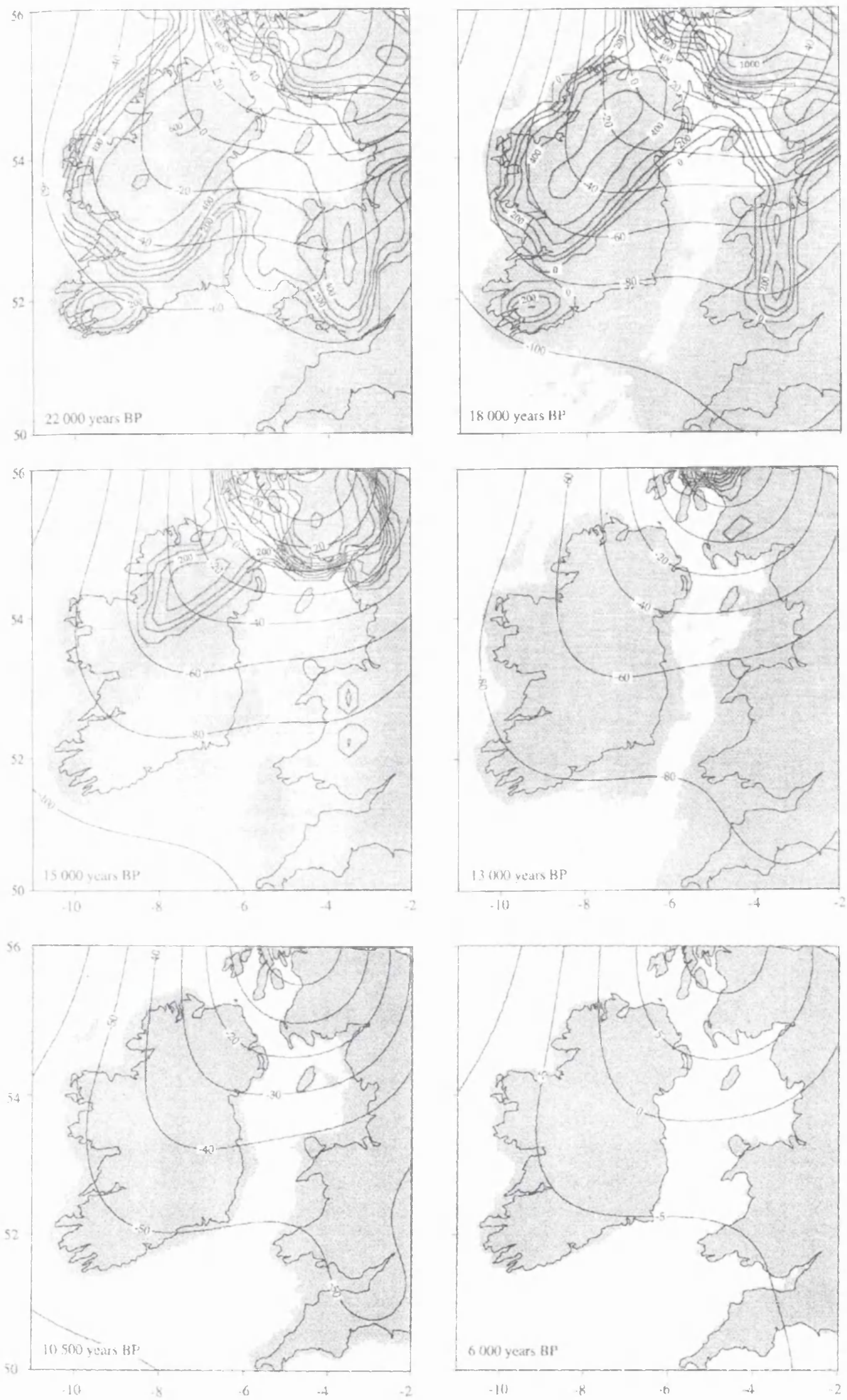

Figure 2.30 Results from Lambeck's (1996) model showing predicted isobases of sea-level relative to present levels (continuous black curves), the palaeo-shorelines (the limits of the shaded regions), and the ice limits and ice heights at selected epochs (continuous grey lines). The ice contour interval is 100 m. (From Lambeck, 1996).

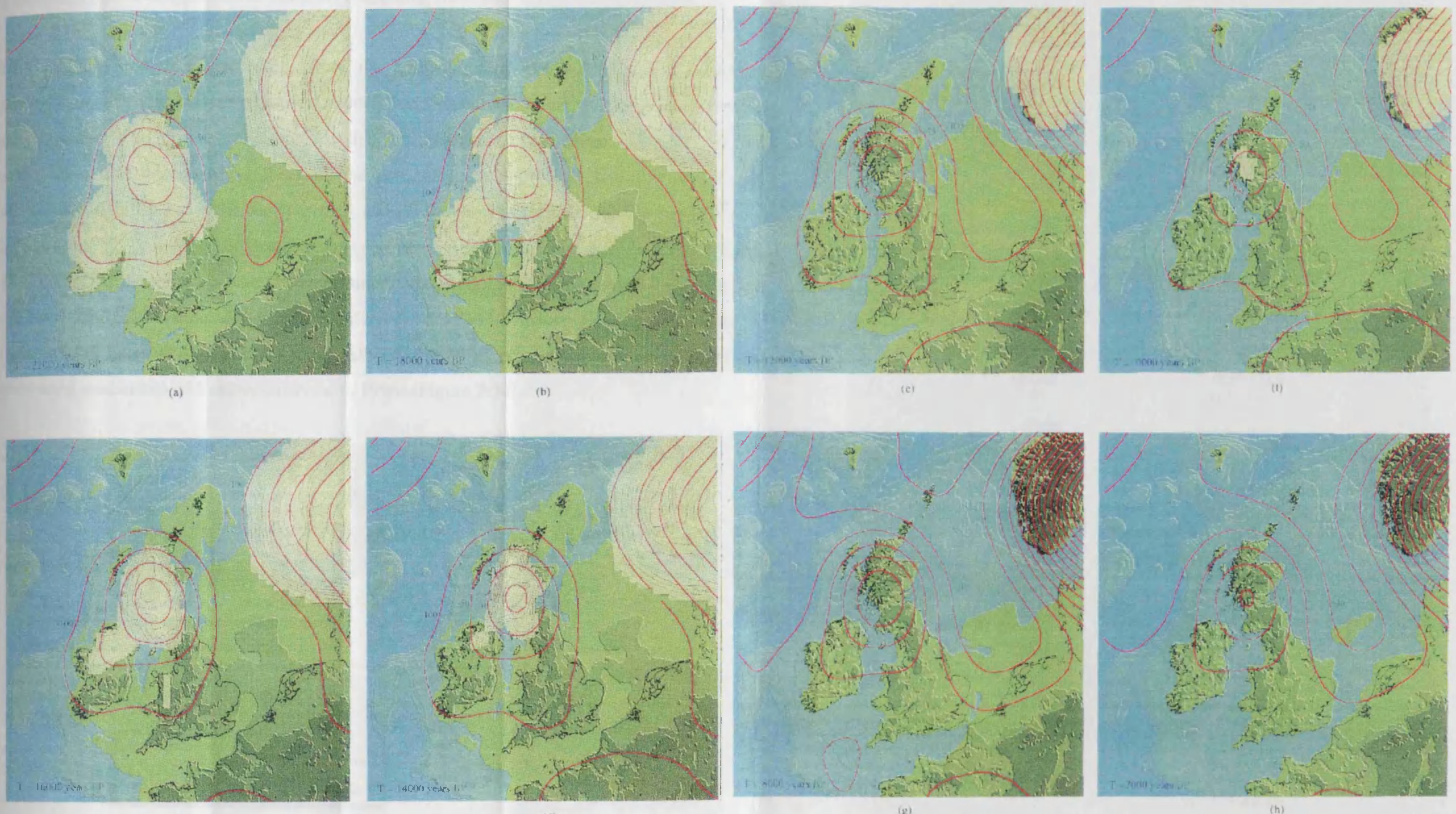

Figure 2.31 Isobase maps of predicted shorelines, shoreline locations and ice sheet limits from selected epochs, according to a model by Lambeck (1995). (a) 22 000 years B.P. corresponding to the adopted maximum glaciation over the British Isles. (b) 18 000 years B.P. corresponding to the time of the onset of deglaciation of the large ice sheets, (c) 16 000 years B.P., (d) 14 000 years B.P., (e) 12 000 years B.P., (f) 10 000 years B.P., (g) 8000 years B.P., (h) 7000 years B.P. The predicted maximum ice heights for the epochs are: 1500 m at the time of the glacial maximum at 22 000 years B.P., 1400 m at 18 000 years B.P., 1300 m at 16 000 years B.P., 1000 m at 14 000 years B.P. and 400 m at 10 000 years B.P. Palaeowater depths are also indicated with contours at 50, 100, 150 and 200 m. Isobase contour intervals are 50 m for (a) to (d), 25 m for (e) and (f) and 10 m for (g) and (h). (From Lambeck, 1995).

1968), and later northwards until the early Holocene lowstand north-east of the Isle of Man was, relatively, at least as low as -40 m O.D. (Pantin, 1977, 1978). The sea-level rise of Phase 2 continued until it levelled off with the attainment of the range of the present tides about 5 ka B.P.

Phase 3. A period of quasi-stable sea-level continuing from 5 ka B.P. to the present day.

McCabe (1997) claims that results from the application of glacio-sedimentary modelling strongly suggest that current numerical models are flawed and in general their application is deterministic. The principal reasons for this are thought to be because: (1) they do not incorporate time-dependent feedback mechanisms or process–process relationships between sea-level variability and ice-sheet dynamics; and (2) they fail to consider lithostratigraphic and biostratigraphic proxies of deglacial palaeo-sea levels (Eyles and McCabe, 1989a; McCabe, 1997). Clear differences between the deterministic trend computed from numerical models and the time-constrained water depth records inferred from field evidence along the east coast of Ireland have been identified by McCabe (1997) as used as primary evidence to dispute the computed predictions of Lambeck (1993a, b, 1996) (Figure 2.32).

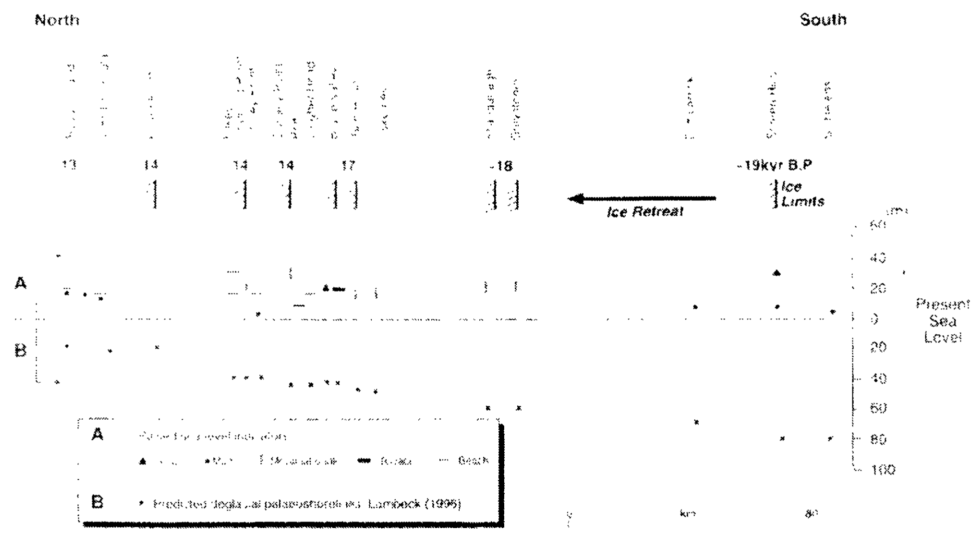

Figure 2.32 A comparison of water depths inferred from raised Lateglacial sea-level indicators along the coast of eastern Ireland with Lambeck's (1996) predicted shoreline levels. (From McCabe, 1997).

Evidence of sea-levels which are higher that those proposed by the numerical models have been reported from a number of localities around the seaboard of the Irish Sea Basin, and consequently have been used to challenge the validity of the predicted Lateglacial sea-level changes achieved from numerical modelling. This is despite the great debate that currently exists over the actual authenticity of the data used in the glacio-sedimentary interpretation. An example of this can be provided from the Mell Formation north of Dublin which occurs at 30-40 m O.D. and is reported to contain a marine fauna that is indicative of water depths of 90-110 m. Consequently, sea-levels at the time of sediment deposition have been proposed at heights of 120-150 m above the present level (e.g. Eyles and McCabe, 1989a; Warren, 1991). However, questions have been raised as to the age of these sediments and whether they are *in situ* or have been transported to their present position (e.g. Warren, 1991; Hoare, 1991).

Further misrepresentation of data has also been highlighted (e.g. Austin and McCarroll, 1992; Harris, 1991) with respect to elevated features from earlier sea-levels (e.g. sea caves, benches, wave-cut notches) which on a number of occasions have been applied by Eyles and McCabe (1989a) to infer Lateglacial sea-levels, despite the absence of quantitative data for the ages of formation of these particular features. Additionally questionable are the deposits used by Eyles and McCabe (1989a) along the coast of Wales indicating Lateglacial sea-levels up to 140 m above their present level near Banc-y-Warren, at 50 m near Cardigan, and near 80 m near Hendre and Brycir on the Llyn peninsula. These deposits have been interpreted as glaciolacustrine deltas and therefore cannot be used as evidence for high relative sea-level (Austin and McCarroll, 1992; Harris, 1991; McCarroll and Harris, 1992).

For further detailed accounts of the predicted relative sea-levels following the glacial maximum (*c.* 22, 000 yrs B.P.), based upon the glacio-sedimentary model, and specific site studies recording quantitative sea-level observations from around the Irish Sea Basin, the reader is referred to Eyles and McCabe (1989a), McCabe (1996), McCabe (1997), and subsequent references within.

CHAPTER THREE

Theory and methods

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Theory and methods

3.1: Introduction

The purpose of this chapter is twofold: to review the theory behind reconstructing palaeo-ice sheet dynamics and depositional settings; and to describe the multidisciplinary field and laboratory techniques incorporated into this investigation (Sections 3.2, 3.3). Part of this research is concerned with the possible influence of a marine environment on ice sheet recession (Section 2.5.3). Therefore, the latter part of this chapter (Section 3.4) addresses the methods behind assessing relative sea-level history within the north-east sector of the Irish Sea Basin during the Lateglacial period.

3.2: Reconstructing Regional Palaeo-Ice Sheet Dynamics

Three approaches to the reconstruction of former ice sheets have previously been identified (Andrews, 1982): (1) the use of geomorphological and geological evidence to piece together ice sheet configuration; (2) mathematical modelling; and (3) the use of shoreline evidence to provide a history of isostatic depression which can be used to infer the location of the main masses of ice. The basis for mathematical modelling of ice sheets is reviewed by Hindmarsh (1993), and the use of isostasy for reconstructing former ice sheets is described by Peltier (1982, 1994). This section discusses the techniques which can be applied in palaeo-ice sheet reconstruction, based upon geomorphological and geological evidence.

3.2.1: Geomorphological and geological record

The reconstruction of former ice sheets from geomorphological and geological evidence, on a regional and continental scale, has been a focus of research for over 100 years (Clark, 1997). The underlying theory behind this approach is that the extent and pattern of flow of an ice sheet is recorded as a geomorphological imprint on the landscape. By piecing together this evidence, which can often be fragmentary, a reconstruction of the geometry of former ice sheets and their evolution through time is possible. Changes of ice-sheet extent can be modelled utilising stacked glacial-nonglacial deposits to define the position of the ice sheet perimeter in space and time. Internal components of ice sheet geometry, the ice-flow patterns and ice divides, can be reconstructed from ice-flow indicators such as erratic dispersal, glacial striae and subglacial bedforms (Clark, 1997, 1999). Of these indicators, the latter features provide the most frequent and spatially extensive evidence of ice flow which previous workers have used to build or support reconstructions (e.g. Dyke and Prest, 1987; Boulton and Clark, 1990; Kleman *et al.*, 1997). This methodology, in conjunction with certain assumptions or preconceptions, normally allows the initial production of a limited reconstruction.

Inevitably, ice sheet reconstruction is somewhat of a slow technique in that much of the evidence arises from detailed field investigation and correlation between local areas. Hence, advances in past ice sheet reconstructions from various large-scale regions have tended to be incremental as the body of local evidence slowly increases. In a recent discussion on the application of this methodology, Clark (1997) identifies that the main barriers to arriving at a coherent synthesis of available geomorphological evidence are:

(1) *Fragmented nature of the evidence.* Commonly many spatially separate studies exist with few common links between them and many contradictions between areas remain unresolved.
(2) *Theory-laden evidence.* Evidence sometimes becomes tainted by interpretations which are based on old glaciological concepts that are now no longer considered valid. There might be a major change of view with regard to how to interpret field data, resulting in the same evidence being interpreted in a different manner. This may require previously documented sites to be re-interpreted in light of this new theory.
(3) *Poor dating control.* It is impossible to date evidence recording information from within ice margins, e.g. drumlins and striae. Dating techniques can only be applied in order to ascertain when an area became ice-free, thereby constraining ice margin positions. This poses limitations on the information that can be gathered relating to ice sheet build-up, configuration change and dynamics.
(4) *Synchronicity of evidence.* When reconstructing ice sheet geometry based on the nature, geographic distribution and pattern of evidence it is convenient to assume that all evidence has formed penecontemporaneously as this provides maximum information about the ice sheet at a snapshot of time. However, incorrect temporal grouping of evidence can lead to contrived and unrealistic reconstructions. This theme will be revisited in Section 3.2.1.2 when the notion of preservation of older landforms and deposits, and the ease to which they can occur, is discussed. It is therefore no longer acceptable to make simple assumptions about synchronicity of evidence.
(5) *Lack of consideration of glaciological plausibility.* Many geological reconstructions of ice sheet flow and geometry may in fact be glaciologically implausible.

The validation of ice sheet models based upon traditional methods of gathering geologically-based information from fieldwork, or through the more advanced use of aerial photographs (Section 3.2.2), is problematic since an overriding concern lies with the confined nature of collected evidence which quite often relates to only a small sector of the ice sheet under consideration. Additionally, there are frequently unresolved disparities between local areas (point 1 above). Further problems arise from the length of time that it takes to conduct detailed field mapping of drift bedforms. Wishart Mitchell, in a personal communication to Clark (1997), suggests that one day per 1-2 km^2 is about average. The application of this type of field investigation is therefore clearly restrictive and presents a number of limitations for regional-scale palaeo-ice sheet reconstruction. However, recent technological advances in the application of satellite imagery have occurred which can be favourably used to aid reconstructions of the evolution of former ice sheets from the fragmentary geomorphological and geological evidence that they leave behind. The use of a more advanced remote sensing technique provides information on a far wider spatial scale and promotes greater analysis and coherence of all geomorphological evidence. The application of this technique in the context of palaeoglaciological reconstruction is the subject of Section 3.2.3.

3.2.1.1: Palaeo-ice flow indicators

Drift bedforms of a linear nature (till lineations; Kleman and Borgström, 1996), parallel to the direction of former ice flow, are frequently displayed on previously glaciated terrain. There has been a consensus that such forms can only develop at the subglacial interface of wet-based ice masses under specific circumstances determined by conditions of materials, material properties and environmental constraints (Menzies, 1987). Under wet-based conditions the substratum is assumed to be continuously reshaped through basal sliding, with flow-aligned lineaments being produced and destroyed (Kleman and Borgström, 1996). Field observations (Goldthwait, 1960; Holdsworth and Bull, 1970) and theoretical work focusing on the basal strength at the ice/substratum interface have allowed glaciologists to conclude that basal sliding is uncommon under cold-based conditions (Boulton, 1972; Hughes, 1973; Paterson, 1981; Drewry, 1986).

The subglacial bedforms commonly used in reconstructing palaeo-ice sheet dynamics have been assigned by Kleman (1994) and Kleman and Borgström (1996) to what is classified as a "wet-based system" (Figure 3.1, Table 3.1). This morphologic system is separately defined from a "marginal meltwater system" (Figure 3.1, Table 3.1) which includes landforms such as eskers, channels and ice-dammed lake traces. Kleman (1994) also suggests that there is further justification for defining a third geomorphic system beneath ice sheets, a "dry-bed system" (Figure 3.1, Table 3.1), which can be seen to compliment the wet-based system and marginal meltwater systems. The "dry-bed system" incorporates the existence of relict surfaces and landforms of former glaciated areas.

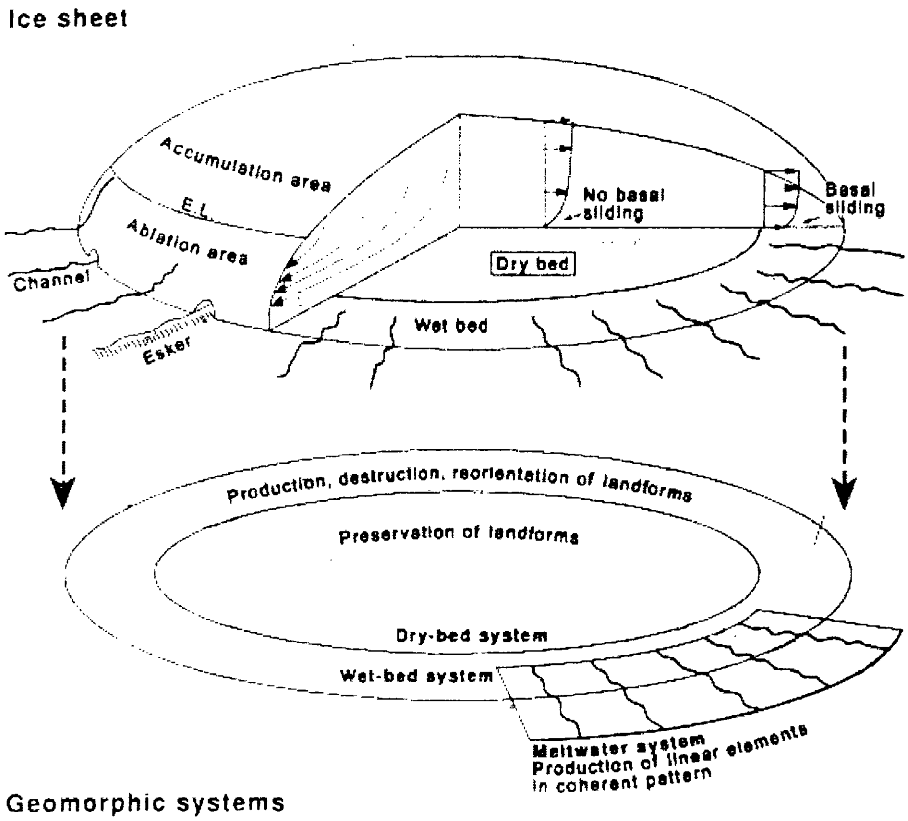


Figure 3.1 Geomorphic systems proposed by Kleman (1994) that result from processes at the ice sheet base and margin. (From Kleman, 1994).

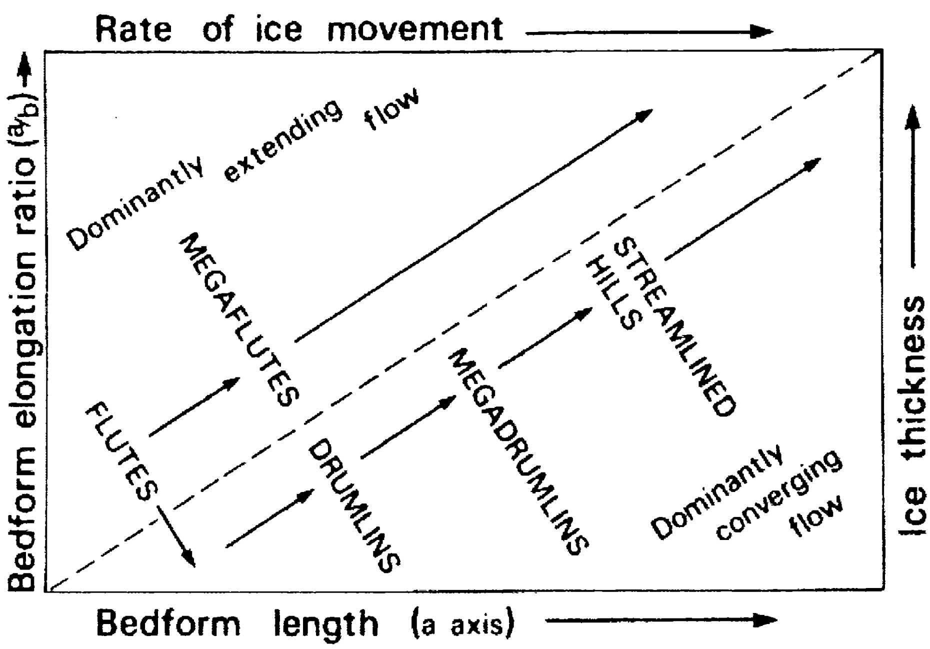


Figure 3.2 Glacier bedform assemblage in relation to ice thickness and rate of ice movement. (From Rose, 1987).

Table 3.1 Inherent properties of the geomorphic systems thought to exist at an ice sheet base and margin.

Morphologic system	*Inherent properties*
Dry-bed system	"No new landforms are created. The frozen bed conditions leads to a hiatus in landform development".
Wet-bed system	"Basal sliding occurs. Flow oriented lineations are continuously produced, destroyed and reoriented".
Marginal meltwater system	"A coherent system of linear features (eskers and/or ice-directed drainage channels) is produced".

Source: Kleman, 1994, p. 27.

Ice-moulded landforms, as distinct from ice-abraded features (e.g. striae, roche moutonnées), can be regarded as an assemblage of upstanding morphological forms that are related in terms of form and spatial frequency. They are traditionally categorised on the basis of their bedform length and elongation ratio and ice thickness and rate of movement (Rose, 1987; Figure 3.2). The different components of this landform assemblage are: flutes which are of the order of 10-10^2 m in length; drumlins, 10^2-10^3 m in length; megaflutes, 10^2-10^3 m in length; and megalineations, $8x10^3$-$7x10^4$ m in length (Clark, 1993). It is implied from this that each natural generic grouping reflects specific modes of operation of subglacial processes which in turn are determined by the natural scale of processes within the environment. Subglacial lineations tend to exist as collections of landforms rather than as isolated individuals. Of these lineations, drumlinised subglacial bedforms provide the most frequent and spatially extensive evidence of former ice flow in Carrick and Galloway.

An extensive and contrasting field of literature exists which describes the distribution, morphometry, sedimentology and glaciological significance of drumlins, as well as presenting the many competing theories invoked to account for their subglacial genesis (e.g. Smalley and Unwin, 1968; Gravenor, 1974; Menzies, 1979; Smalley, 1981; Boulton, 1982; Shaw, 1983; Dardis and McCabe, 1983; Menzies, 1984; Menzies and Rose, 1987; Shaw and Sharpe, 1987; Shaw *et al.*, 1989). Conclusive answers to questions relating to these attributes have not been forthcoming, and consequently different assumptions have been applied about the glaciodynamic context of bedform generation which has led to widely varying reconstructions of former ice sheets (Clark, 1999).

A second group of palaeo-ice flow indicators used in reconstruction are ice-abraded features such as striae, roche moutonnées, whalebacks (Flint, 1971) and rock drumlins (Linton, 1963). The morphology and distribution of ice-abraded features can reveal much about their mode of formation and supplements the information provided by ice-moulded landforms.

3.2.1.2: Cross-cutting relationships

Boulton and Clark (1990) have reported previously unrecognised patterns of crossing drift lineations from the North American continent, which they suggest represent temporal changes in ice sheet flow. Other studies in New York State, USA (Fairchild, 1929), in Scotland, northern England and Norway (Rose and Letzer, 1977; Riley, 1987; Rose, 1987) and on Victoria Island, Canadian arctic (Dyke and Morris, 1988) have recorded similar examples of cross-cutting topologies. It therefore appears to be clearly evident that the coexistence of subglacial bedforms of different orientation provides a geomorphological catalogue of different ice-flow events in these regions (Clark, 1993).

Within Canada, as many as four different orientations of bedform are understood to have been preserved (Clark, 1993). Clark (1993) suggests that if it is accepted that each is aligned parallel to a former direction of ice flow then the cross-cutting pattern must undoubtedly imply different directions of ice flow at different times. This development is understandable given that if a field of parallel drift lineations are no longer harmonious with the flow direction then they will serve as obstructing protuberances to planar ice flow, as well as acting as an available sediment supply (Clark, 1994). They are therefore highly susceptible to modification explaining why cross-cutting lineations are so widely reported.

The surprising factor which emerges is that many upstanding bedforms must actually survive overriding by ice from other directions and so are preserved as palimpsests beneath the more recent record of ice flow (Clark, 1994). Clark (1994) surmises that the assumption that a more recent, or dominant, ice flow direction will remould all of the underlying drift and obliterate previously produced orientations is completely unfounded. This is contrary to those thoughts of Prest (1984) and Boulton *et al.* (1985) who consider the ability of an ice sheet to reorganise sediment as high. A question of fundamental importance for palaeo-ice sheet reconstruction exists as to whether or not drift bedforms can survive prolonged ice-sheet coverage with no or minimal morphological disturbance, and if so, under what conditions (Kleman, 1994). Dyke and Morris (1988) claim, through spatial analysis of regional drumlin systems on Prince of Wales Island, arctic Canada, to have delineated areas that were unaffected by basal sliding. These older drumlin fields were interpreted on this occasion as having been preserved in frozen-bed zones (Dyke and Morris, 1988).

The cross-cutting topology of sediment ridges is currently thought to be a result of two types of relationship (Clark, 1993): (1) superimposition, or (2) pre-existing lineation deformation. Superimposed forms exist upon larger lineations of different orientation such as, for example, drumlins superimposed upon megadrumlins (Rose and Letzer, 1977) and flutes on megaflutes (Rose, 1987). For this feature to be observed, it is necessary for the superimposed forms to be of a smaller size than their parent landform (Clark, 1993). Deformation of pre-existing lineations involves forces that do more than simply mould the surface of the lineation into new shapes, as is the case in superimposition. Sediment is actually required to migrate in a downstream direction, causing breaching and attenuation of the parent landform. Clark (1993, 1994) proposes that this subglacial modification of pre-existing lineations into new ice-aligned landforms operates as a continuum of modification. The cross-cutting patterns represent an intermediate stage during the process of complete reorganisation of the bed. The progression is from no modification of the parent landform, to superimposition of smaller forms, to substantial breaching or deformation of the parent landform, to total reorganisation of sediment into an entirely new orientation (Figure 3.3).

The parameters that affect the continuum of lineation modification are unknown, but suggested controls are ice velocity, time elapsed, ice thickness or intricacies of sediment rheology (pore water pressures, frozen/unfrozen till) (Clark, 1993). It is suggested by Clark (1993) that if the main determinant of the extent of landform modification is the velocity of ice flow, then it would be expected that there would exist some form of spatial correlation between modes of modification and the velocity zones of former ice sheets. This is illustrated in Figure 3.4 where it is invoked that beneath the ice divide region there is no deformation of pre-existing bedforms (zero velocity), but as velocities progressively increase along the transect (or flowline) towards the ice margin this is reflected in the strength of subglacial deformation, and hence the degree of landform modification.

Clark (1993) raises an important question that if changes in ice flow direction were gradual, then why is it often the case that a small number of discrete orientations are observed rather than either a whole array of intermediate orientations, or just the final orientation after readjustment to the most recent ice flow direction? This point has also been highlighted by Kleman (1994) who similarly suggests that under continuously wet-based conditions and no-change conditions we should expect a much 'fuzzier' picture of lineations, with continuous transitions between sets, instead of the 'jumps' which have been reported on occasions. As stated above, continuous readjustment cannot be the case if cross cuts are observed. At present there are two main conceptual models by which cross-cutting lineaments can be interpreted in terms of ice flow evolution (Clark, 1997):

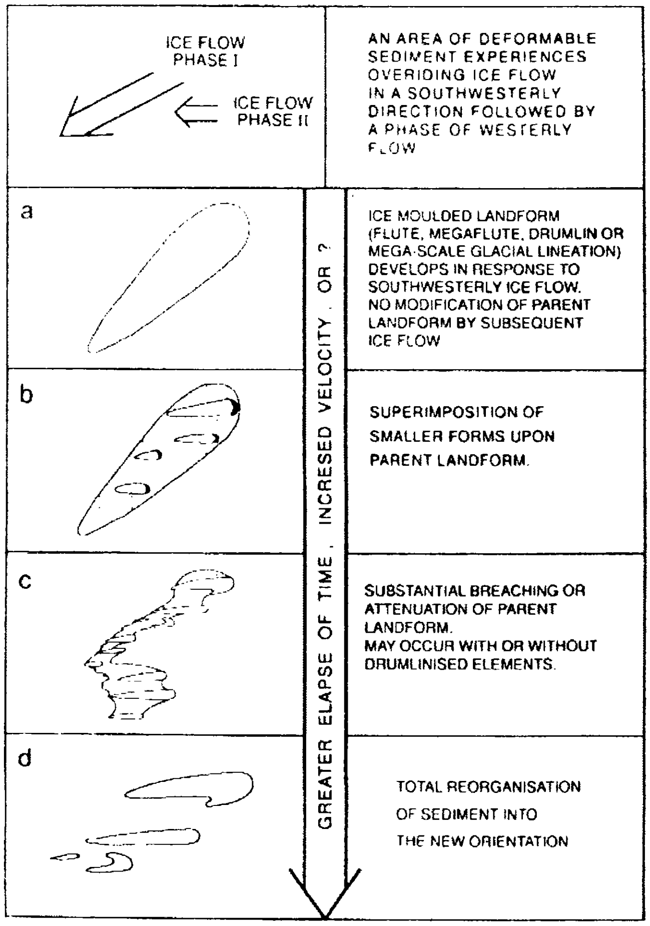


Figure 3.3 Theoretical continuum of subglacial modification of a lineation experiencing overriding ice flow from another direction. (From Clark, 1993).

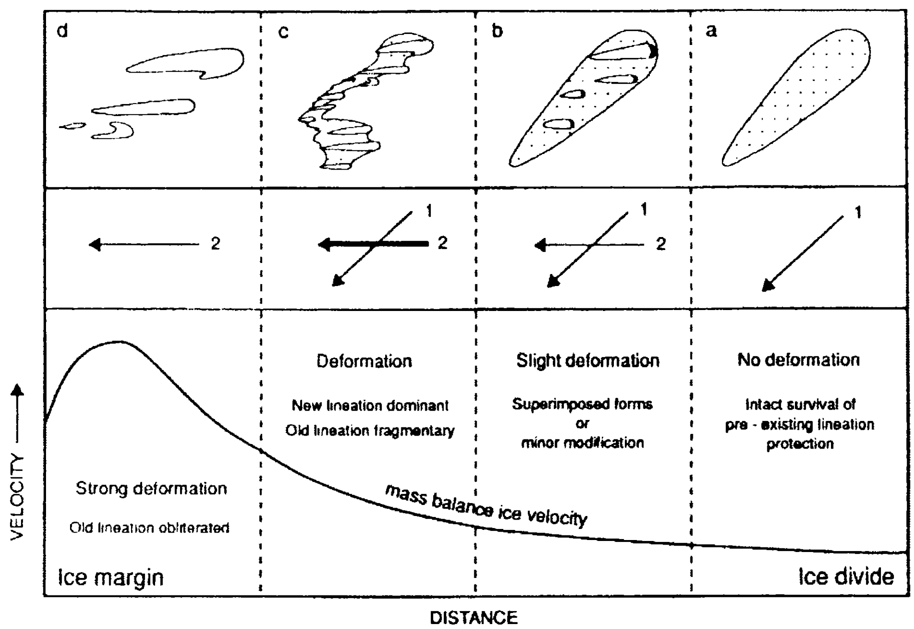


Figure 3.4 Indication of how the velocity of ice flow from the ice divide to ice margin may be the main determinant of the degree of subglacial modification. (From Clark, 1993).

1. *Ice divide migration.* The change in flow directions documented by cross-cutting lineations arise from changes in the location of ice-dispersal centres or ice divides. A series of rapid ice-divide shifts between positions of relative stability will induce the creation of a number of superimposed radial patterns (Figure 3.5). A similar pattern may arise also from a steadily migrating ice divide if some form of on/off mechanism of lineation generation operates under the emanating ice flowlines, allowing bedforms to be rapidly generated (Clark, 1993).

2. *Lobate margin retreat.* The systematic production of cross-cutting lineations arises from successive stages of margin retreat. Consideration of the process of lineation generation during ice margin retreat reveals that for any ice margin configuration, other than perfect straight margins which are rare, either discordant or systematic cross-cutting should result (Figure 3.6). Of crucial consideration in this model is the position within an ice sheet under which streamlined landforms are generated (Clark, 1994).

A further explanation is that some of the patterns may record ice flow configurations from more than one glaciation (Clark, 1997).

The geomorphological nature of cross-cutting relationships often permits the relative chronology of the different orientations to be ascertained, thus providing information on the changing dynamics of ice flow occurring at different times within the life of the ice sheet. This theme is visited in more detail throughout Section 3.2.3.2.

3.2.2: Application of aerial photography

The use of aerial photographs is a well-established technique in geomorphological research, providing information of great value to studies such as this. A number of former ice sheet beds have been investigated in detail utilising rapid mapping of aerial photographs. These have been shown to provide excellent information on pertinent features such as moraines, eskers, drumlins, flutes, crag and tails, meltwater channels, proglacial deltas, lake shorelines, marine limits and so on (e.g. Hare, 1959; Prest, 1983; Mollard and Janes, 1984). Sometimes, features show up clearly on the photograph which are difficult to appreciate on the ground, especially where relief changes are small (King, 1966). Furthermore, their main value is that they enable many morphological forms to be recognised which are not shown on topographic maps, even if they are on a large scale (King, 1966). The ease of recognition with which landform units and individual forms can be outlined is particularly useful, even if their genesis and age remains undetermined (Verstappen, 1972).

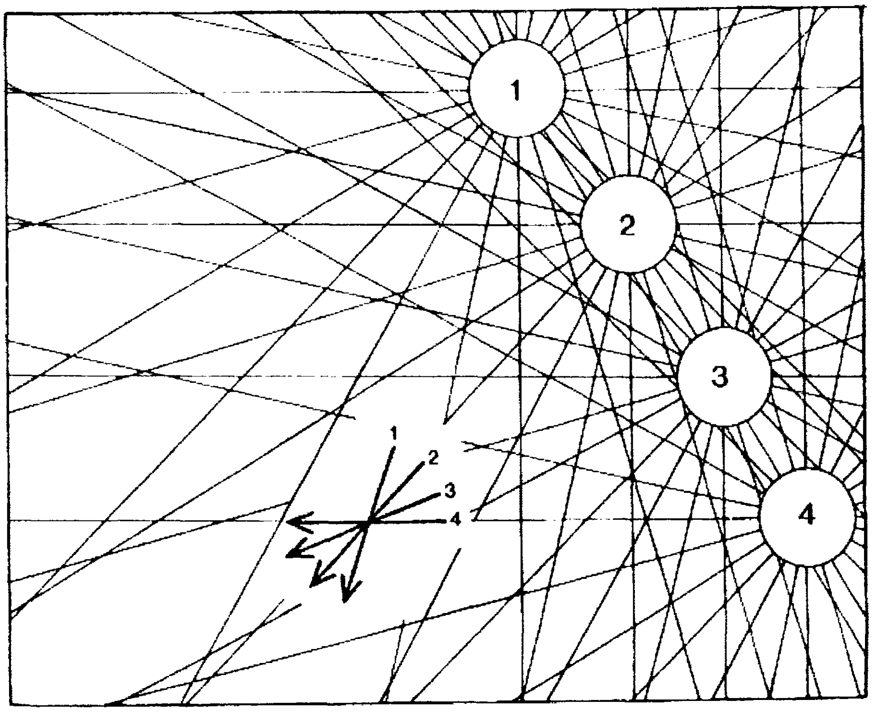


Figure 3.5 Rapid migration of an ice divide between relatively stable positions (1 to 4) creating a number of superimposed radial patterns. Close to the dispersal centres, flowlines emanating from different ice-dispersal centres will intersect at angle of approximately 90°. At greater distances from the ice-dispersal centres, the upstream angle between intersecting flowlines will be less. (From Clark, 1993).

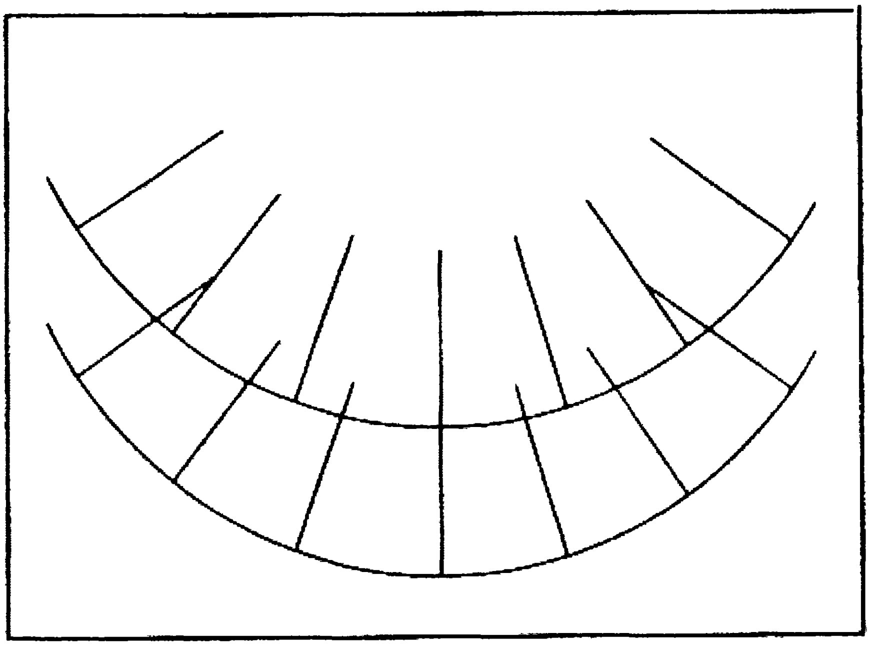


Figure 3.6 Probable lineation pattern from ice margin retreat displaying expected discordant or cross-cutting topology of lineations. (From Clark, 1994).

The degree of information obtained from aerial photographs is highly dependent on the following factors (Clark, 1997): (1) solar elevation at the time of photography; (2) vegetation cover; (3) photograph scale; and (4) the quality of the prints. Given that these conditions are satisfactory, the geomorphological interpretation of photographs becomes a fairly straight-forward type of investigation, particularly if the relief is observed in the form of a stereoscopic image. Geomorphological forms (e.g. drumlins, eskers, meltwater channels) can normally be readily identified on the aerial photographs on the strength of their photographic density/grey tone and/or their relief characteristics. They may also be recognised by their association with vegetation and often the nature of land use (Verstappen, 1972). A variety of photo-interpretation methods exist with the choice appearing to be determined by the level of detail required, nature of the terrain, the type of study, and the time and expense the researcher is willing to expend. Verstappen (1972) outlines the following routine as being commonplace in aerial photograph interpretation:

i. Rough scanning of the photographs after study of existing literature.
ii. Establishment of a mapping legend.
iii. Drawing of the major topographic features.
iv. Detailed mapping of geomorphological features.
v. Execution of measurements if required, e.g. height differences, slope angles.
vi. Field check / ground truthing.
vii. Final interpretation, merging field and photo observations.

3.2.2.1: Geomorphological mapping

The process of geomorphological mapping, as described recently by Hayden (2000), begins with the identification on aerial photographs of the fundamental land units that form the land surface. This is viewed as a critical process in allowing the nature and character of these units to be assessed in the wider context of a palaeo-environmental reconstruction. Miyogi *et al.* (1970) have described a geomorphic unit in general terms as 'an individual, genetically homogeneous landform produced by a definite constructional or destructional process'. The question of what conforms to the definition of a geomorphic unit has fuelled many debates (e.g. Wright, 1972; Speight, 1974; Gardiner, 1976) although it appears realistic to accept the argument that the identification of a geomorphic unit is significantly related to matters of regionalisation and scale (Hayden, 2000). Therefore, at different scales and covering regions of different size, different landform features can be identified as basic homogeneous units and so the choice of unit is dependent on the scale of investigation (Hayden, 2000).

For the purpose of this investigation, selective geomorphological mapping was conducted in order to:

1. Recognise the surface forms occurring in the mapped areas.
2. Provide a precise picture of the dynamics of relief for the purpose of:
    a. elucidating relief development.
    b. providing a way to evaluate factors and processes of morphogenesis and transformation.
3. Facilitate the search for associations between landforms with respect to their arrangement in space and mutual relations in the system.
4. Provide a means of comparison between contemporary and fossilised land surfaces.
5. Produce a cartographic representation of landform distribution and properties.

3.2.3: Application of Landsat satellite imagery

In recent times, the examination of glacial geomorphology portrayed on satellite images has presented the opportunity to elucidate features that were previously unperceived in the field or on larger scale aerial photographs. Using this technique, Boulton and Clark (1990) mapped continent-wide patterns of glacial lineations using Landsat satellite images over the whole area of mainland Canada north of 48°N which was formerly covered by the Late Wisconsinan Laurentide ice sheet. It was to their surprise that suites of superimposed lineations of varying directions and previously unsuspected large scale patterns of streamlining within drift were depicted on the satellite image. Overall, this presented a far more complex picture than that of a single lineation orientation earlier identified through means of ground surveys and conventional aerial photography (Boulton and Clark, 1990). Satellite imagery also permitted Boulton and Clark (1990) to observe palimpsests of lineations representing former ice movements beneath the last, often dominant lineation pattern. This is explained by the very large area that such single images cover which allow the continuity of earlier lineation patterns to be observed, in contrast to aerial photographs from which fragments of 'earlier' lineations are invariably difficult to re-compose visually into coherent lineation patterns (Boulton and Clark, 1990).

The recent advances in using satellite images to reconstruct former ice sheets from the fragmentary geomorphological and geological evidence that they leave behind permits the maximisation of field evidence and the conduction of widespread mapping in a systematic manner. This undoubtedly provides a more satisfactory interpretation of palaeo-ice flow dynamics at a scale relevant to ice sheet wide synthesis than would otherwise be obtained through traditional methods of field mapping and/or the use of aerial photographs (Clark, 1997). As most satellite imagery is digital it is also easier to work at a wide range of scales, up to the limit imposed by the spatial resolution, such as 1:45 000 for Landsat Thematic Mapper (TM) (Clark, 1997). The ability to work across a wide range of scales, for example from 1:45 000 to 1:1 000 000, provides greater opportunity to detect landforms or patterns that would otherwise have gone undetected (Clark, 1997). With aerial photographs it is difficult to work at smaller scales (e.g. 1:100 000) without access to specially prepared mosaics of photographs which are rare, and so mapping is normally restricted to a scale of around 1:20 000 (Clark, 1997).

In summary, the main benefits for using satellite imagery for identifying the expression of former ice flow in south-west Scotland are:

1. Many landforms can be detected more easily than by aerial photographs or ground surveys;
2. Widespread areal coverage permits the discovery of new landforms and patterns;
3. It is possible to work at a wide range of scales (e.g. 1:45 000 to 1:1 000 000);
4. Speed of mapping is far greater than can be achieved using aerial photographs or carrying out field mapping.

3.2.3.1: Geomorphological expression of ice flow on Landsat images

Landsat data sensed by the Thematic Mapper (TM) scanner produces individual images with pixel sizes approximately 30 m x 30 m and images typically covering an area 185 km x 185 km so that fifteen scenes are required to cover the extent of the British ice sheet (aerial photographs typically cover an area of 12 km x 12 km) (Clark, 1997). It is therefore the pixel size, and hence spatial resolution of such images, that limits the potential for identifying small ice-flow landforms (<80 m), while displaying medium-sized forms (80 m - 5 km) as a linear grain, and large features (>5 km) as distinguishable linear expressions (Clark, 1993). Flutes cannot be recognised, but a typical drumlinised terrain and megaflutes are readily visibly (Clark, 1993). On occasions, such forms may not necessarily be identifiable as a number of individual elements, but rather as a "co-linear texture or grain" (Clark, 1990, p. 2; Figure 3.7). It is therefore possible to use Landsat images to produce maps of former ice flow directions as depicted by the orientation of drumlins and megaflutes.

The co-linear texture or grain apparent on Landsat images is characterised by repeated parallel lineations arranged in a pattern similar to that of a field of drumlins but at a different scale (Clark, 1993). The majority of lineations combine to produce spatially coherent and distinctive

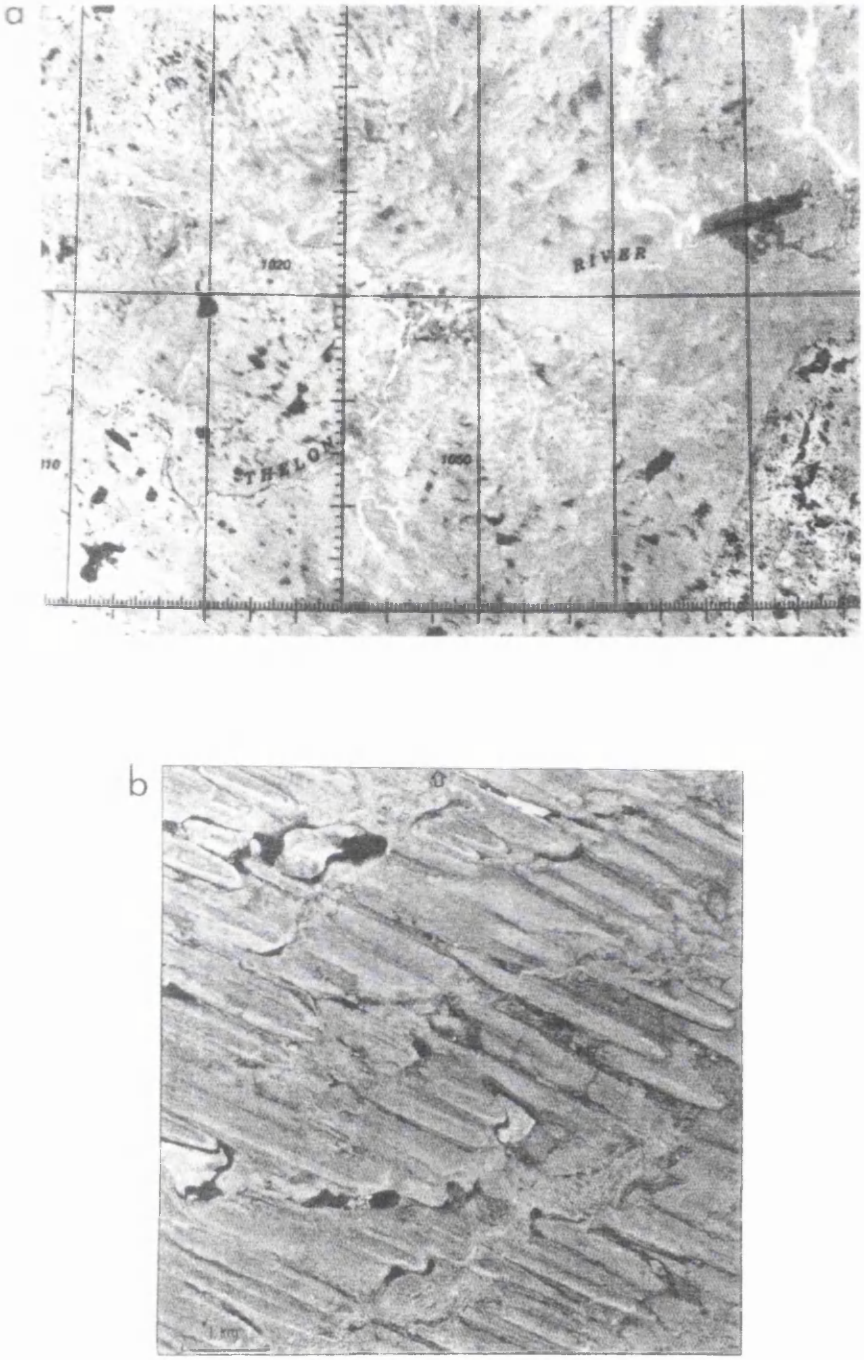


Figure 3.7 A grain of NW-SE orientation visible on a Landsat image (a) of Thelon River region, North West Territories, Canada; east-west dimension of image is 120 km. By comparison with an aerial photograph (b) of part of the same area, it is can be seen that the grain is composed of a multitude of drumlins; east-west dimension of the photograph is 7 km. (From Clark, 1993).

patterns often extending for many hundreds of kilometres which may be grouped into what Clark (1993, 1994, 1997) has called 'flow sets'. This grouping is usually performed by subjective visual pattern recognition, based on some combination of the following criteria (Clark, 1999):

1. *Parallel concordance.* Each linear bedform should have a similar orientation to its neighbours.
2. *Close proximity.* Each linear bedform should be in close proximity to its neighbours. The spacing will typically be of the order of up to two to three times the dimensions of the landform.
3. *Similar morphometry.* Neighbouring linear bedforms usually display similar morphometry.

Table 3.2 Components of ice-flow landform assemblage making up the glacial grain on Landsat images.

Landform component	*Strength of detection*	*Appearance on imagery*
Drumlins and megaflutes (10^2-10^3 m)	Size just above the resolving capabilities of the Landsat MSS sensor	Grain rather than as individual forms
Mega-scale glacial lineations ($8x10^3$-$7x10^4$ m)	An individual streamlined glacial lineation in drift which can be detected on a scale larger than would be readily observed on conventional aerial photographs	Strong glacial grain (Much greater in length than drumlins and megaflutes)
Fragmentary drift elements	Elements appear disordered at aerial photograph-scale but are seen to be organised into coherent sub-parallel linear patterns at the scale of Landsat images	Components of the pattern may display a simple uni-directional type, or be part of a more complex pattern representing multiple ice flow directions

Source: Clark, 1993, 1994.

Comparison between areas recording drumlins and megaflute orientations from aerial photographs, and mapped drumlinised terrain of the same site from Landsat images show high degrees of correlation between the two. No process is known, other than moving glacial ice, that could produce straight linear drift forms organised into such flow sets (Clark, 1993). Clark (1993) regards the glacial grain depicted on such satellite images as an ice-flow landform assemblage comprising three components (Table 3.2).

The visual identification of drumlins, megaflutes and other glacial landforms on satellite imagery is dependent upon their ability to provide an identifiable signal distinct from that of the background terrain (Clark, 1993). In moorland areas, where much of the land cover is saturated with groundwater, the topography of a field of drumlins is such that it may give rise to vegetation surface cover differences between the more boggy inter-drumlin zones and drier drumlin ridges (Clark, 1997). The advantage of using Landsat data is that it is known to be good at discriminating between different types of vegetation cover (Clark, 1997). Inevitably, many drumlin fields do not display any distinct changes in vegetation cover between drumlin ridges and inter-drumlin areas, and so are unlikely to be detected as they fail to meet the criteria of providing an anomalous signal (Clark, 1997).

As a result, other remote sensing techniques have been adopted to negate the non-guaranteed association between surface vegetation and drumlin position. The criteria which remote sensing must exploit are the changes and breaks in slope which allow the identification of drumlinoid forms in field work studies. A signal highly correlatable with slope angle and aspect can be produced from satellite imagery with a low angle illuminating source (Clark, 1997). Landsat data imaged while the sun is low in the sky provides such a result. Slaney (1981) outlines, however, that because the orbit of any satellite is fixed it is impossible for a user to acquire imagery of a particular area at a specified time, and hence at the appropriate solar elevation. Landsat for example will always acquire images of south-west Scotland at the same time of day, so the only way of obtaining low elevation data is by choosing the appropriate season of the year (Clark, 1997). For mid-latitude regions, sun elevations are lowest in the winter months.

3.2.3.2: Relative age determination of ice-flow sets

The relative ages of individual flow sets can be determined by using both individual conventional aerial photographs and high-resolution satellite images to reliably detect the reorientation of elements of pre-existing lineations into new lineation sets. It is invalid to assume that the determination of the relative age of two cross-cutting sets may be achieved by identifying the dominant or 'fresh-looking' lineation pattern as being the most recent ice-flow direction. Clark (1993) makes reference to a number of examples which falsify this assumption. It is only possible to reliably ascertain the correct sequence of ice-flow phases by analysing, in detail, the geomorphological relationship between the lineation sets. The two forms of relative age indicator, suggested by Clark (1993), which can be found through close examination are:

1. *Superimposition.* One set of streamlined lineations is visibly superimposed upon another, differently orientated, set.
2. *Pre-existing lineation deformation.* Recent phase deformation by ice movement can often alter the form or continuity of a pre-existing lineation, thus revealing the relative age (i.e. dissection of an older set by a younger set (Boulton and Clark, 1990)).

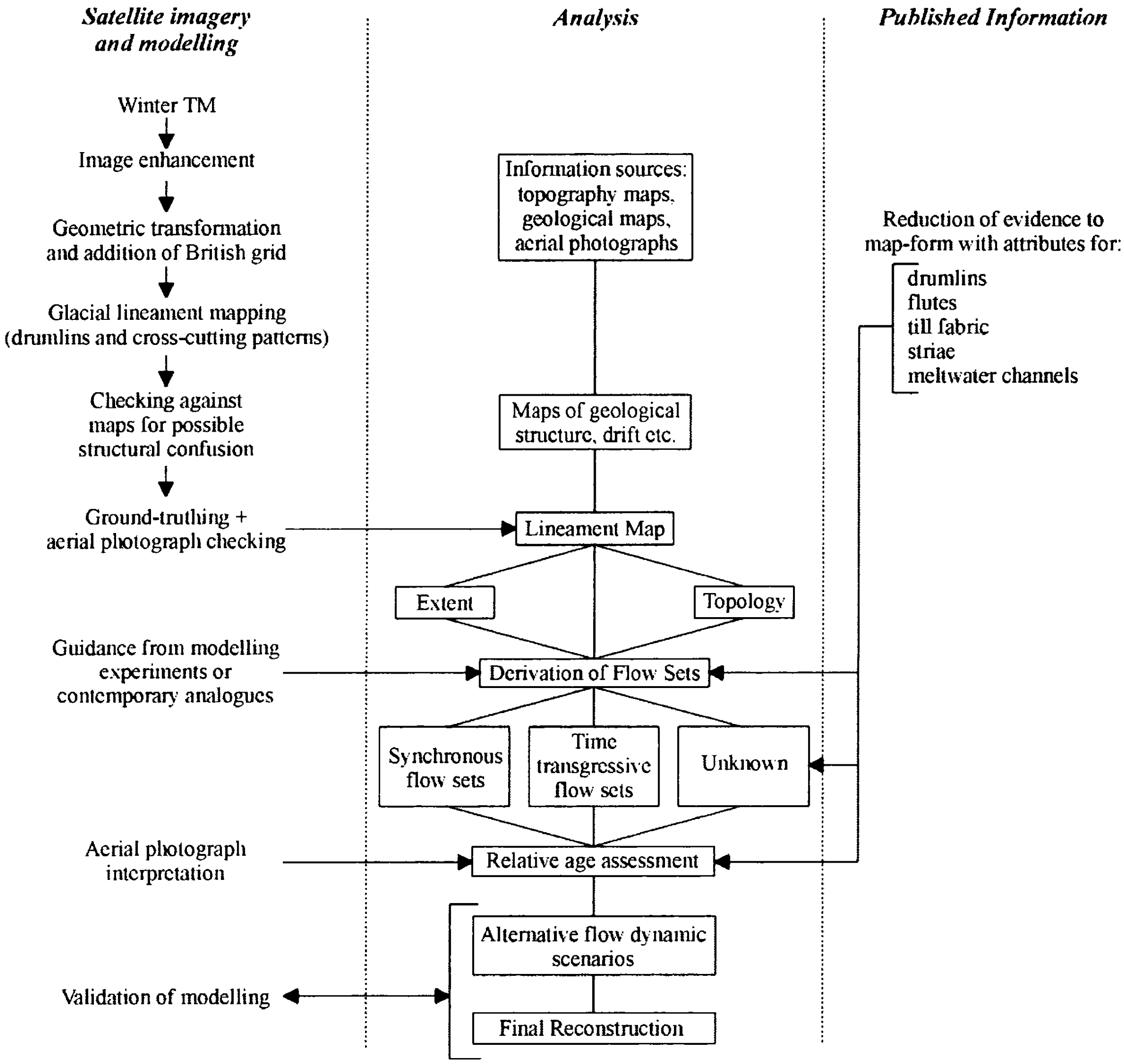


Figure 3.8 Summary of strategy used to reconstruct the palaeo-ice sheet dynamics of Carrick and Galloway, south-west Scotland (Modified from Clark, 1997).

3.2.3.3: Strategy for the reconstruction of flow geometries

This section, summarised in Figure 3.8, details how the technique of remote sensing was applied in the reconstruction of the palaeo-ice sheet dynamics of Carrick and Galloway regions, south-west Scotland. A number of constraints were placed on the extent to which the methods could be matched to one outlined by Clark (1997) in a review paper discussing the process of reconstructing former ice sheet dynamics using remote sensing and a geographic information system (G.I.S.). Therefore, the following strategy represents a modification of the extended procedure which would have otherwise been adopted.

The type of satellite imagery used to map the glacial geomorphology belongs to the Landsat series of satellites which have provided a continuous supply of images since 1972, resulting in complete coverage of previously glaciated terrain world-wide (Clark, 1997). The most widely used optical data source is that derived from the Multispectral Scanner (MSS) which allows the most detailed scale of mapping to go up to 1:120 000 (Clark, 1997). However, Landsat with the Thematic Mapper (TM) sensor was preferred on this occasion because of its ability to offer an image with a greatly improved resolution to that of the MSS (TM pixel size = 30 m^2; MSS pixel size = 80 m^2) (Clark, 1997). Such data have a much larger computational volume (270 Mb, rather than 32 Mb) which hinders its use, but its increased resolution allows mapping with much finer detail (up to 1:45 000) which makes it possible to identify individual drumlins (Clark, 1997). Other image types (SPOT, SEASAT, ERS, Radarsat) were thought to be unsuitable as they were either limited in the level of information that they provided (due to high solar elevation, limited coverage or terrain distortion) or they were difficult to analyse. The Landsat TM imagery used in this investigation (Figure 3.9) was supplied by the Department of Geography, University of Sheffield, where the initial stages of processing were conducted.

The most effective approach in processing the optical data was found to be the technique outlined by Clark (1997, p. 1078-1087). Firstly, a subset of the image was chosen for which a fairly good knowledge of the landform configuration was known and a series of tests were carried out using image enhancement techniques until the most appropriate was found, which was then applied to the whole image. A decision had to be made at an early stage as to whether to proceed with a single waveband displayed in monochrome or a multi-band image as a colour composite. It is well known that the human eye can perceive linear features when they are present in the form of a monochrome image (Drury, 1983). However, the majority of drumlins within Carrick and Galloway maintain a different vegetation cover from the inter-drumlin areas. This was found to be clearly displayed on a colour composite image which was selected

Figure 3.9 Colour composite Winter Landsat (TM) image of south-west Scotland prior to geometric correction.

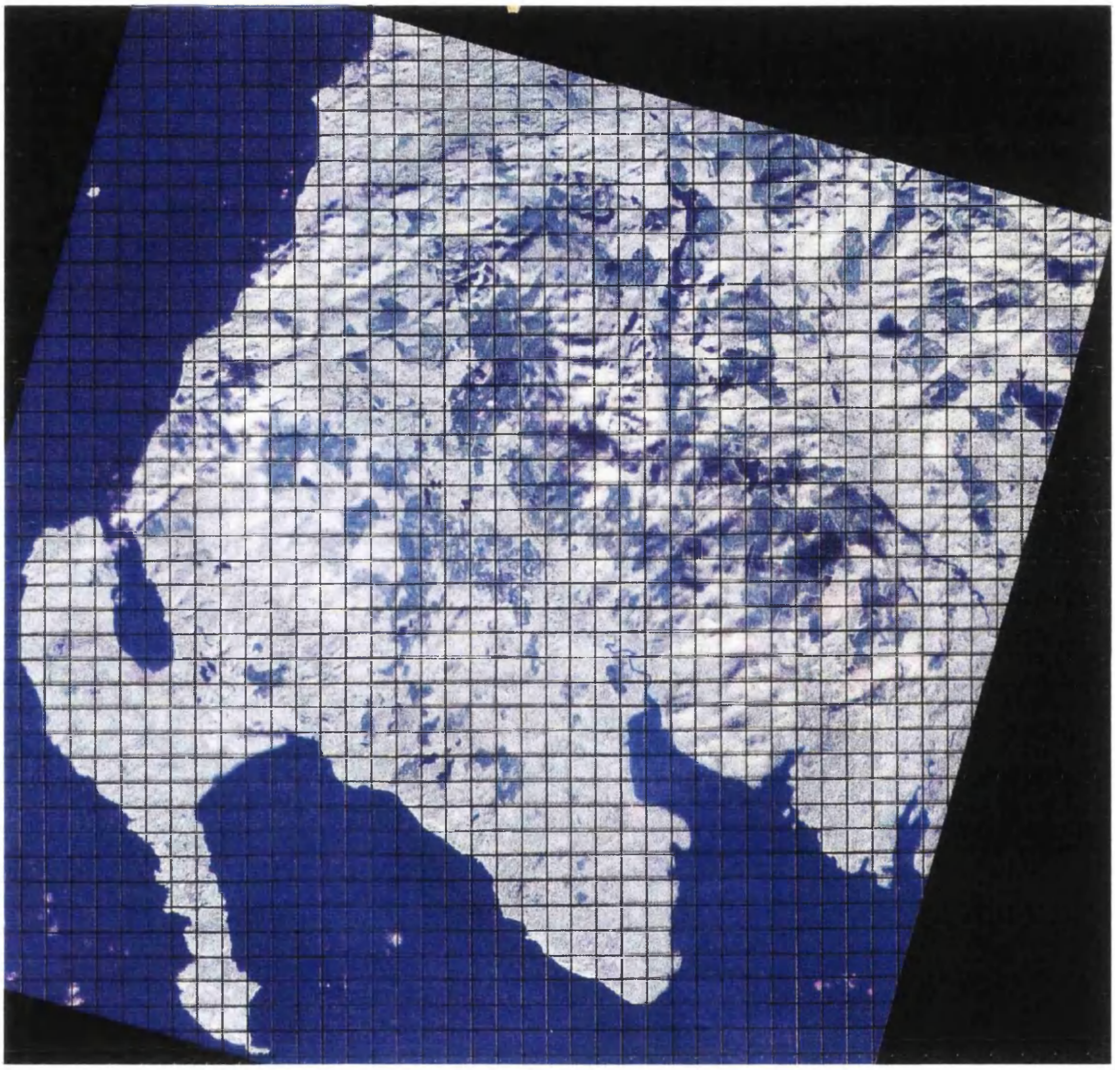


Figure 3.10 Geometrically corrected colour composite Winter Landsat (TM) image with superimposed 1:50 000 British grid (each square = 1 km^2).

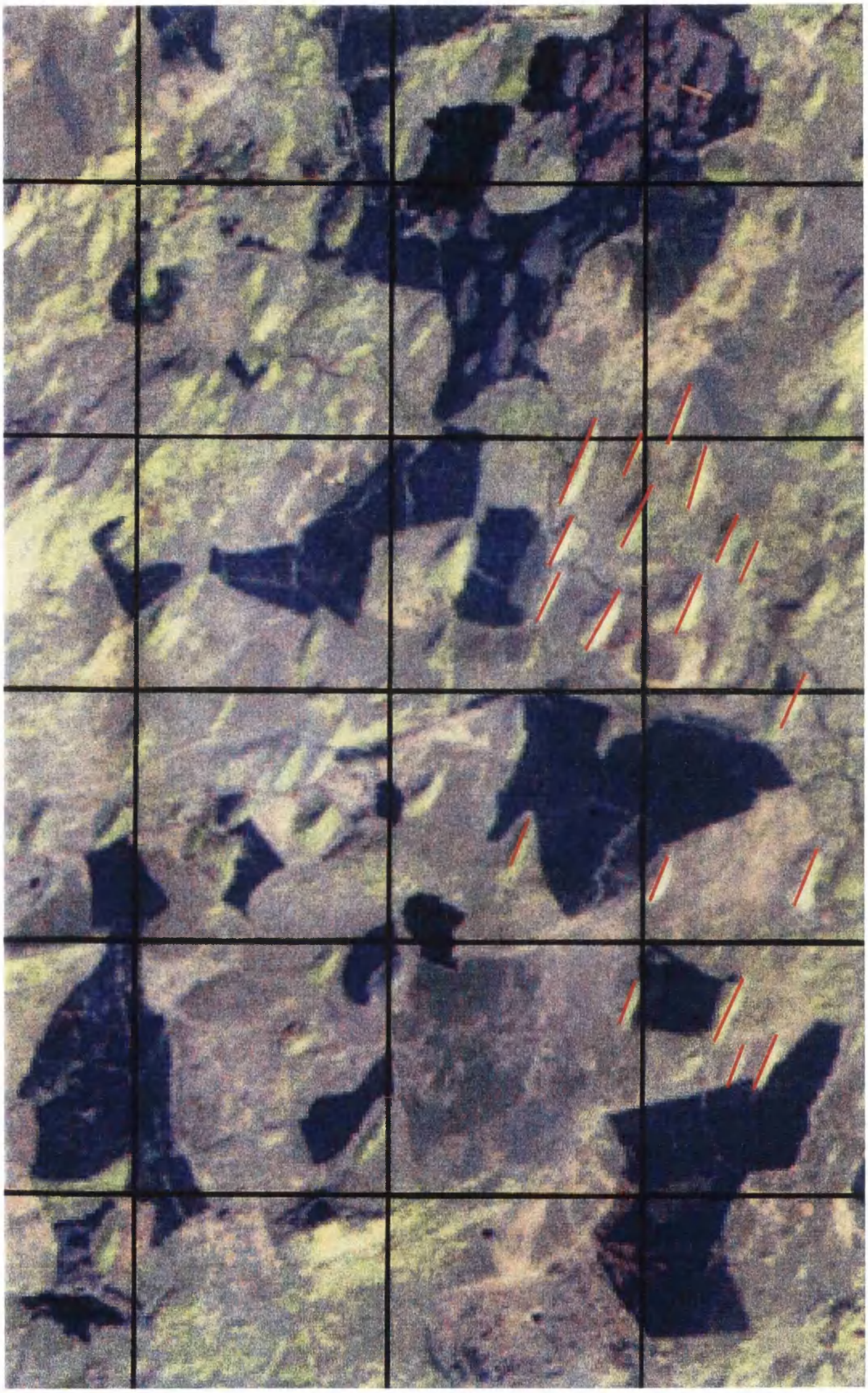


Figure 3.11 Example of one of the sixty-six panels making up the satellite image of south-west Scotland with manual photo-interpretation of some of the ice-flow drift lineations (red solid lines). Superimposed grid is the 1:50 000 British grid (each square = 1 km^2).

after subtle variations in the hue had been made to intensify the visual separability between the vegetation cover types.

Satellite data are not usually supplied in a geometry that accords with common map projections (Clark, 1997). There were found to be a number of sources of geometric distortion and map projection changes which required a certain degree of correction. To achieve this it was necessary to locate tie points, features that can be recognised on the image and a map and for which map co-ordinates are known, which were then used in a geometric transformation. The 1:50 000 British national grid was superimposed over the satellite image with the view that this would later allow any features of interest to be easily located on either a published map or on the ground (Figure 3.10, 3.11).

From digital satellite data the mapping technique of glacial landforms can take one of two routes (Clark, 1997). One method requires the acquisition of photographic prints of the satellite image printed out on a high quality printer which can then be treated in the same way as a conventional aerial photograph. Analysis (photo-interpretation) must therefore be by manual means, tracing features of interest from these prints onto an overlay (Figure 3.11). The disadvantage of this method is the under-utilisation of the spatial and spectral resolution of the data which is displayed on the high quality display of a computer monitor (Clark, 1997). This is particularly the case if the printer is not of sufficiently high quality. Consequently, a better approach is via on-screen digitising where subtle changes in tone or fine structure are more easily detected (Clark, 1997). This also has the added advantage of permitting mapping at a multitude of scales, by zooming in and out as required, and also allows the interpretations to be captured digitally, allowing the information to be used in a GIS. if desired. Unfortunately, the latter option was not attainable due to a number of constraints, therefore the image was processed using the initial method, divided up and printed onto sixty-six panels for photo-interpretation at the Department of Geography and Topographic Science, University of Glasgow (Figure 3.11).

The actual process of photo-interpretation (similar in theory to the process discussed in Section 3.2.2) is one which requires great care in order to recognise the lineaments pertinent to the reconstruction of ice flow direction (Clark, 1997). For example, in areas of thin drift cover with a strong structural control on the underlying bedrock, differentiation between glacial lineaments and bedrock lineaments can sometimes be problematic. In addition, drift can sometimes be draped over bedrock lineaments which can lead to erroneous interpretations of

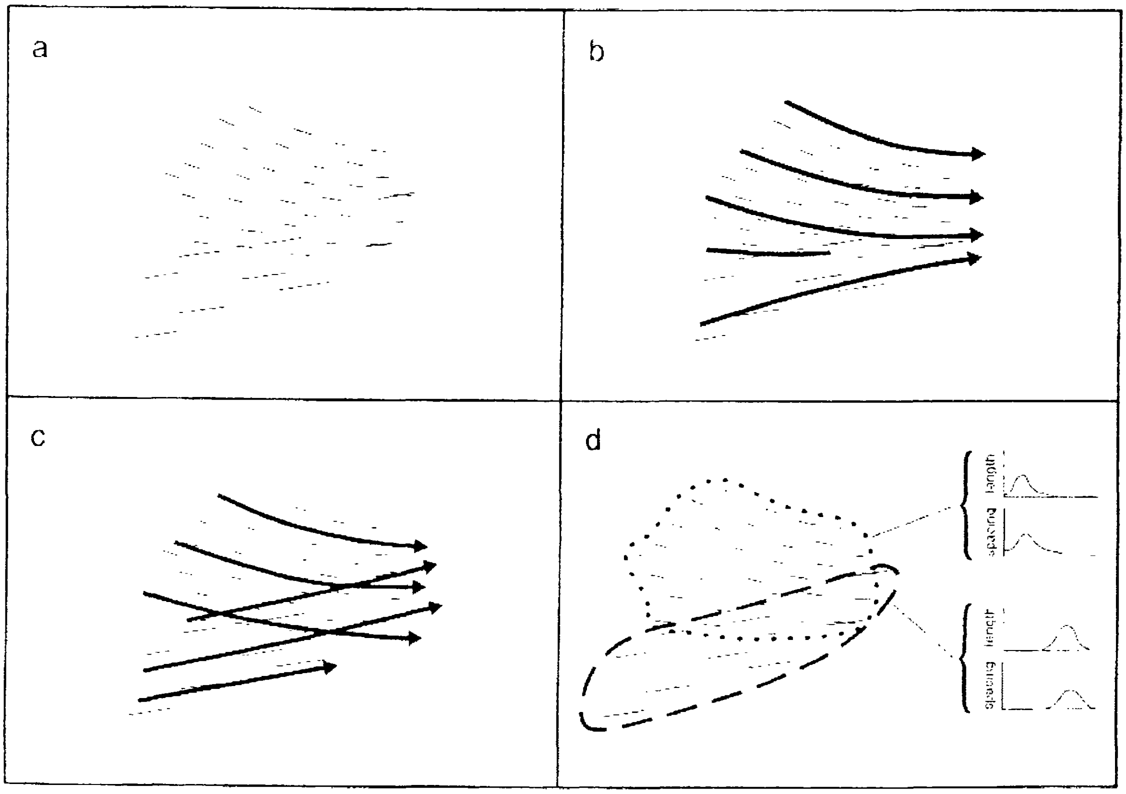


Figure 3.12 The grouping of lineaments (a) into individual flow-sets. It is easy to make the mistake of grouping evidence into a single flow event (b), when it may in fact represent information of different ages cross-cutting one another, and hence representing two flow events (c). The spatial pattern and morphometry can sometimes assist in discriminating flow events (d). (From Clark, 1997).

ice flow orientation. To aid in the recognition of drift lineations, maps of surficial drift cover and structural geology were used to interpret areas of possible confusion.

Following the mapping of glacial lineaments from the remote sensing imagery, the next stage in reconstructing the ice flow pattern was to group the lineament patterns into coherent flow-sets, in a similar way to that outlined in Figure 3.12. Supplementary measures such as aerial photograph checking and ground-truthing were critical in accurately achieving this. Clark (1997) views this as a crucial stage in the procedure in that these form the basic units in any further reconstruction. Clark (1997) further states that the main criteria to be considered during assessment is the degree to which the lineaments can be grouped into sub-parallel sets whose 'topology and extent is glaciologically plausible'. Once the flow sets have been established they must be assigned as representing a time transgressive or synchronous flow event which again requires consideration of field evidence (e.g. striae, meltwater channels). Relative age information can be sought from aerial photographs if the examination of the flow-set map reveals superimposed flow locations (Clark, 1993) in order that the flow-sets can be sorted in terms of age. A final reconstruction can then be made, following careful consideration of other ice sheet models, alternative scenarios, and the validity of any assumptions applied.

3.3: Reconstructing Palaeo-Depositional Settings

Assessing the properties and history of depositional modes in former glaciated environments from sediment sequences is understandably a troublesome task particularly when consideration is given to their variety and complexity, and the way that they characterise the constantly changing nature of their processes of deposition (Benn and Evans, 1998). Throughout a large-scale palaeo-depositional environment it is normal for a wide spectrum of sediments to be represented, reflecting former activity within the glacial system such as: (1) the shifting positions of active depositional centres; (2) deformation, reworking and resedimentation of debris by active and stagnant ice; (3) redistribution by flowing meltwater, gravity and aeolian processes; and (4) cycles or reworking during periglacial and paraglacial activity (Benn and Evans, 1998).

Understanding the final depositional position of a complex range of glacigenic sediments and the landforms that they underlie has provided glacial geologists with problems in how to accurately interpret them. Many solutions have been proposed to the problems associated with their correct description and classification, as described in a detailed review by Dreimanis (1989). In recent years, workers have started to appreciate that glacial depositional systems are ordered on a number of different levels, either as a series of steps in time or a range of scales in space. The temporal dimension is important for understanding the origin of sediment deposits, because different attributes of a sediment's characteristics develop at different points of the erosional, transportation and depositional history of the material as it migrates through the glacial system (Benn and Evans, 1998). The sequence of stages can be used as a basis for rudimentary sediment classification (Boulton and Deynoux, 1981; Dreimanis, 1982, 1989).

An appreciation of the spatial dimension is also necessary in order to establish the position of a deposit with respect to not only its adjacent sediments, but also the wider environment, and the landscape as a whole (Benn and Evans, 1998). In modern environments, there are commonly a range of active processes which are responsible for glacigenic sediment being laid down as part of an assemblage, rather than as an isolated deposit. This is such that when viewed on a regional scale, as is being done in this investigation, the character and distribution of the glacial deposits should provide an intuitive insight into the shifting patterns and processes of the last glaciation. The environmental context of a sediment can therefore be defined at different levels of a spatial hierarchy, beginning with the immediate locality and panning out to wider and wider horizons. This hierarchical approach to sedimentology is seen as a powerful means of describing how sediment landforms and landscapes fit together, and determining how organisation in the landscape reflects the organisation of depositional processes (Benn and Evans, 1998).

A greater understanding of temporal and spatial dimensions within the context of sedimentology has encouraged the development of an effective systematic approach to the classification of glacial sediments. This form of methodological approach is presently viewed as a reliable method for palaeo-environmental reconstruction and is described in Section 3.3.1. Section 3.3.2 introduces the formal description scheme which has been used to define and model glacigenic sediment units and assemblages.

3.3.1: Sediment classification

Increased knowledge of glacial depositional settings, in particular those experienced in contemporary environments, has resulted in rapidly evolving classifications of glacial sediments (Eyles *et al.*, 1983). These are based both on actual field observations of depositional processes and *a priori* theoretical considerations of likely depositional processes (Dreimanis, 1976, 1979; Stankowski, 1980). In the past, there has been a tendency for workers to employ traditional criteria emphasising laboratory-derived characteristics (Willman and Frye, 1970; Karrow, 1976) which realistically are not inherently related to the depositional environment and fail to account for the variability of sediment units. Eyles *et al.* (1983) describe this as a

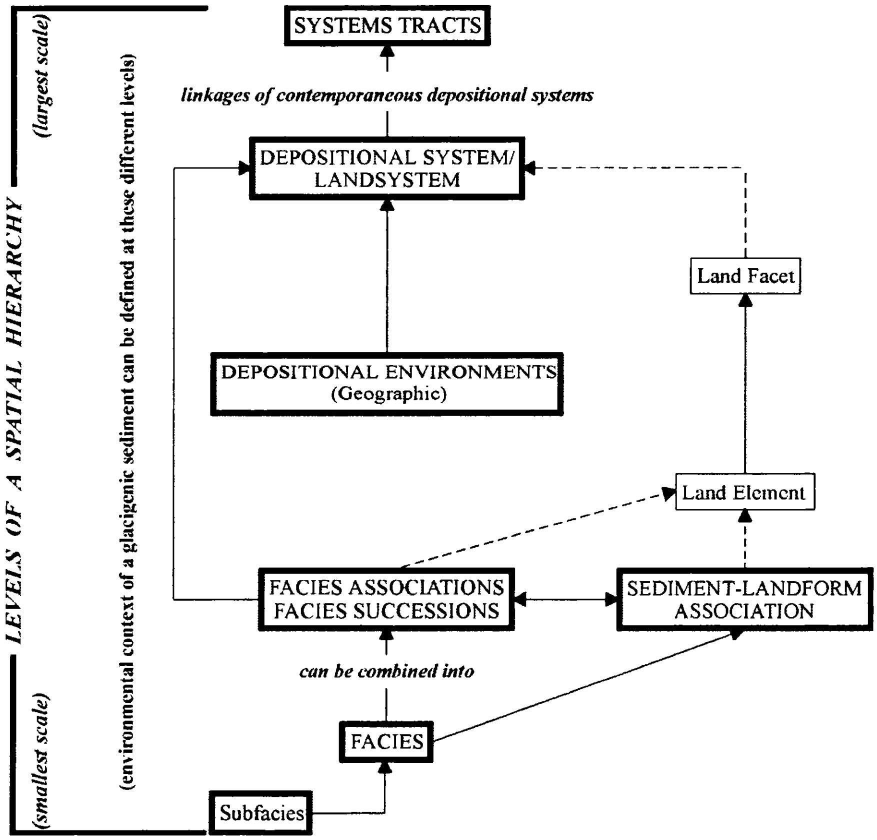


Figure 3.13 Hierarchical sediment classification. Glacial sediments and associated landforms at four levels of organisation: (1) facies; (2) sediment-landform associations; (3) depositional systems, or landsystems; and (4) glacial systems tracts. (Modified from Walker, 1992).

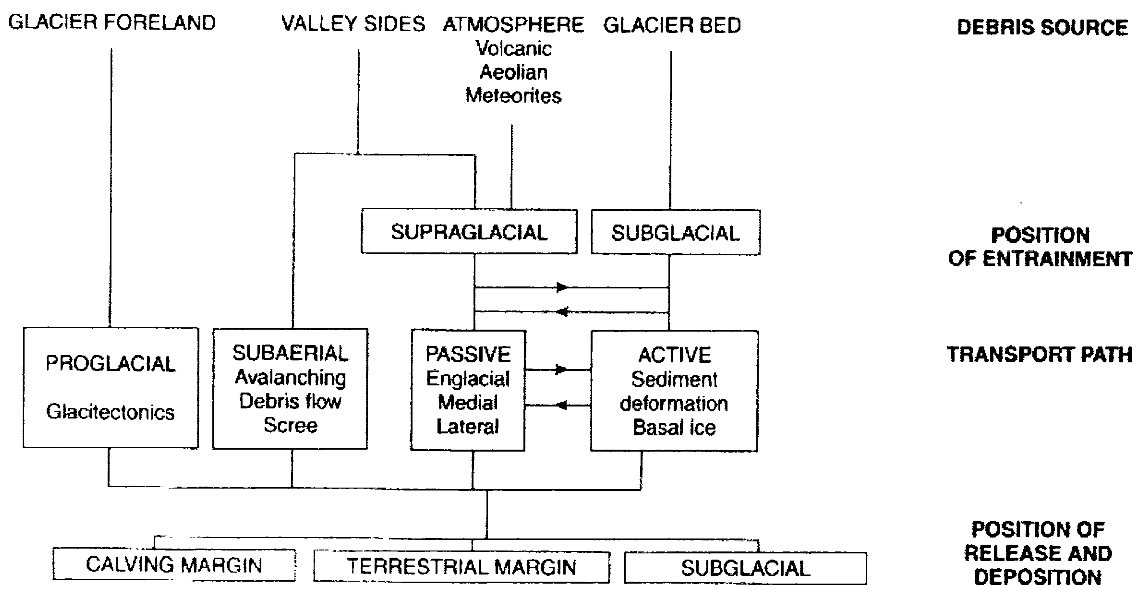


Figure 3.14 Debris cascade system. (From Benn and Evans, 1998).

'procedural' approach to the classification of glacigenic sediments which overall reflects traditional interest in stratigraphic correlation rather than the modern interest in reconstructing depositional environments. The following sections (3.3.1.1 - 3.3.1.5) introduce the background and criteria for modern-day classification schemes and examines the hierarchical approach used in the classification of glacial sediments in this investigation (Figure 3.13).

Table 3.3 Determinants on the physical properties and characteristics of debris within a glacial system.

Stage	*Event*	*Position within glacial system*	*Influence on physical properties of deposits*
(1) **Debris source**	Primary input of material into the system	• subglacial (e.g. plucked and abraded bedrock, or overridden sediment) • extraglacial (e.g. rock walls and pre-existing deposits)	• controls the lithology of the particles in a sediment • particle morphology • grain size distribution
(2) Transport path	Carrying of debris through the system along one or more routes	• active (subglacial) • passive (englacial, medial, lateral) • subaerial or subaqueous • proglacial	• particle morphology • grain size distribution • particle durability
(3) **Depositional processes**	Mechanism that lays down the final deposit (glacial, fluvial, aeolian)	• subglacial • terrestrial margin • calving margin	• geometry and extent of beds • sedimentary structures (e.g. lamination, cross-bedding and grading) • particle morphology • fabric • geotechnical properties (e.g. porosity, shear strength and permeability)

Source: Church and Gilbert, 1975; Boulton, 1978; Slatt and Eyles, 1981; Dowdeswell *et al*., 1985; Humlum, 1985; Gomez *et al*., 1988; Werrity, 1992; Benn and Ballantyne, 1994; Benn and Evans, 1998.

3.3.1.1: Debris cascade system

Glacigenic deposits can be viewed as the products of a sequence of processes extending throughout the whole of the glacial system. Chorley *et al.* (1984) suggest that the outcome of a series of steps taken by debris, that is their physical properties and characteristics, can be accounted for within what is termed the debris cascade system (Figure 3.14). This system is broken down into three stages: (1) debris source; (2) transport path; and (3) depositional processes, each taking on a unique role in determining the development of sediment properties (Table 3.3). Benn and Evans (1998) view the debris cascade system as a background for the genetic classification of sediments found in glacial environments. It can be observed in Table 3.3 that the main criterion for the classification is depositional process, but within the earlier stages of the debris cascade system there is clearly a certain degree of influence placed upon the developing properties of deposits. Examples of properties acquired during each of the three stages experienced by the material are summarised in Table 3.3.

In the past, many classification schemes have been proposed for glacigenic deposits in which deposits are named according to process *and* position of deposition (Benn and Evans, 1998). However, it is important to realise that the environment of deposition has little effect on the sediment characteristics. The result has been a profusion of named types of 'till' and related deposits (see Dreimanis, 1989; Brodzikowski and van Loon, 1991). Confusion ultimately reigns when the environment of deposition is uncertain, and is therefore normally inferred from careful analysis of the surrounding sediments and landforms, though even then it may be ambiguous (Benn and Evans, 1998). The alternative approach used in this investigation is to classify sediments by the processes of formation, as revealed through their physical properties (Lawson, 1981a, b, 1989; Eyles *et al.*, 1983). The approach to defining the position and environment of deposition of a sediment are discussed below.

3.3.1.2: Facies

At the smallest scale of organisation within a depositional environment is the individual sedimentary deposit or *facies* (from the Latin word meaning 'aspect' or 'appearance of' something) (Benn and Evans, 1998; Figure 3.13). Reading (1986, p. 4) describes a facies as a "body of sediment with specified characteristics" which may be subdivided into sub-facies or combined into a facies association or sediment-landform association; the next level in the hierarchy (Section 3.3.1.3). In this context, it should be possible to distinguish a facies from neighbouring sediments because of its deposition by a single process or groups of processes acting in close association (Walker, 1992; Benn and Evans, 1998).

The term *lithofacies* is applied when reference is made to only the objective, physical characteristic of the deposit, with no reference to depositional processes. This is distinct from *genetic facies* which are named sedimentary units that imply a specific mode of formation, and provide varying amounts of environmental information (Benn and Evans, 1998).

3.3.1.3: Facies associations and sediment-landform associations

The next level in the hierarchy of sediment classification is the *sediment-landform association* or *facies association* (Figure 3.13). Facies associations can be defined as an assemblage of facies that occur together and are considered to be genetically or environmentally related (Reading, 1986; Walker, 1992). They consist of sediments that were deposited adjacent to one another or were laid down in an unbroken vertical succession. A conformable vertical transition between two facies reflects a lateral shift in depositional processes (or laterally adjacent environments), such as where glacier retreat causes ice-marginal deposits to be superimposed on subglacial sediments (Walther, 1894; Blatt *et al.*, 1980; Reading, 1986; Benn and Evans, 1998). The boundaries of facies associations are marked by erosional surfaces or non-deposition or bounding discontinuities, which represent breaks in deposition of varying duration. Sediment-landform associations are facies associations that have geomorphic expression in a landform which is genetically related to the facies from which it is composed. An example of a sediment-landform association is a streamlined ridge of deformed sediments aligned parallel to glacier flow, characteristic of a fluted till surface (Benn and Evans, 1998).

The importance of facies associations and sediment-landform associations are that they provide additional evidence which makes environmental interpretation easier than treating each individual facies in isolation (Reading, 1986). They therefore place facies within an environmental context, indicating the local stratigraphic, structural and genetic relationship between closely associated facies (Benn and Evans, 1998). Moreover, they can provide preliminary implications for the nature of particular depositional environments or sub-environments, in a way that lithofacies or genetic facies infrequently can (Benn and Evans, 1998).

3.3.1.4: Landsystems

Facies associations and sediment-landform associations can in turn be grouped together into *depositional systems* or *landsystems* (Figure 3.13). A landsystem comprises a large-scale grouping of facies which have been deposited in an overall environment, such as the assemblage of sediments and landforms deposited in the proglacial region of a retreating glacier (Benn and Evans, 1998). An individual landsystem therefore has a topography, subsurface stratigraphy and sediments which are characteristic of it. This form of terrain evaluation constitutes an holistic approach, wherein both the geomorphology and underlying materials that characterise a particular environment are genetically related to the process involved in their development (Benn and Evans, 1998).

Type examples of glacial depositional systems and landsystems were introduced by Boulton and Paul (1976) and developed further by Boulton and Eyles (1979) and Eyles (1983c). Three glacial landsystems have been defined by Eyles (1983a, c): (1) the *subglacial landsystem*, formed at the bed of the glacier; (2) the *supraglacial landsystem*, which forms as a drape over the former glacier bed following ice retreat; and (3) the *glaciated valley landsystem*, associated with mountainous areas. The categorising of these depositional settings is, however, not inclusive as it omits landsystems formed in proglacial and glacimarine environments.

Each landsystem has been identified by Eyles (1983c) as the largest unit in an ascending hierarchy of landscape classes (Figure 3.13). This hierarchy consists of (1) a *land element*, which is uniform in form and material and can be mapped at a large scale, e.g. a drumlin or kame; (2) a *land facet*, which is one or more land elements grouped as an homogeneous landscape, and suited to mapping at scales of 1:50 000 to 1:100 000, e.g. a drumlin field or outwash plain; and (3) a *landsystem*, which is a repeated pattern of associated land facets which are suited to mapping at scales of 1:250 000 to 1;1 000 000. A glacial example of land elements, land facets and landsystems might be drumlins, drumlin fields, and the whole assemblage of forms representing the former glacier bed (Benn and Evans, 1998).

Eyles and McCabe (1989a) have identified two major depositional systems in the Irish Sea Basin: (1) a subglacial depositional system; and (2) a glaciomarine depositional system (Section 2.6.2; Figure 2.21). The study of depositional systems at this scale is important because the impact of Irish Sea glaciation is most apparent.

3.3.1.5: Sequence stratigraphy and systems tracts

A new and powerful approach to the study of glacial depositional systems has arisen from the concept of *sequence stratigraphy*. From a geological perspective, sequence stratigraphy may be defined as "the study of rock [sediment] relationships within a chronostratigraphic framework of repetitive, genetically related strata bounded by surfaces of erosion or non-deposition, or their correlative conformities" (Wagoner *et al.*, 1988, p. 39). Within such sequences, strata may be subdivided into *systems tracts*, defined as linkages of contemporaneous depositional systems (Wagoner *et al.*, 1988; Benn and Evans, 1998). Therefore, sequence stratigraphy provides a method of analysing the complex relationship between facies in a depositional setting, and assists in interpreting patterns of environmental changes recorded in the geological record.

In recent times, some principles of sequence stratigraphy have been applied by workers to glacial depositional systems (e.g. Boulton, 1990; Eyles and Eyles, 1992; Martini and Brookfield, 1995). It has been found that the most appropriate glacial systems tract is the glacier advance and retreat cycle (Benn and Evans, 1998). This is because within the maximum glacial limit, depositional systems deposited during advance and retreat phases are normally separated by a subglacial erosion surface or a wedge of subglacially deposited or deformed sediment (Berthelsen, 1978; Boulton, 1996). Beyond the ice limit, distal-proximal-distal successions can often be found recording the migration of depositional systems in response to glacier advance and retreat (Boulton, 1990; Eyles and Eyles, 1992).

3.3.2: Sediment description

> "A description of the basic properties of geomorphological materials is frequently the most important starting point for an explanation of a geomorphological process" (Whalley, 1990, p. 112).

From Section 3.3.1, it is clear that sedimentary facies are not entirely unambiguous, for they are ultimately determined by the controlling depositional process(es) acting in an environment. Although a depositional basin might be regarded as an entity, it has also been highlighted that a wide array of sediment types may be deposited within it, through transformation of dominant processes in time and/or space, and through physiographic differentiation. The description of sedimentary deposits therefore requires a methodology which can be applied to a wide range of deposits. Inadequate description and labelling of glacigenic material can often restrict the outcome of any detailed environmental interpretation (Eyles *et al.*, 1983). In glacial sedimentological studies, a reliable method has been developed which is based upon the systematic documentation and representation of three significant parameters unique to an individual deposit: (a) its properties, allowing details of its erosional, transport and depositional history to be inferred; (b) its geometry; and (c) its position with respect to neighbouring sediments and the land surface (Benn and Evans, 1998). A similar approach, detailed in Sections 3.3.2.1 - 3.3.2.3, is applied in this investigation.

3.3.2.1: Sediment properties

Sediment properties provide the foundation for defining and describing a sedimentary unit. The choice of properties to be described are normally determined by the overall object of the investigation, because different properties convey different types and levels of information (Benn and Evans, 1998). Properties described in this investigation are: (1) sediment texture (dominant grain size and grain size distribution; Appendix 1); (2) particle shape; (3) particle roundness; (4) mineralogy; (5) fabric; and (6) palaeocurrents. A description of the techniques for measuring these properties can be found in Bradley (1985), Goudie (1990) and Gale and Hoare (1991).

3.3.2.2: Lithofacies coding

The method used in this investigation to record lithofacies properties in the field is the application of shorthand or *lithofacies codes*, a technique that is increasingly employed by sedimentologists. Miall (1977, 1978) first introduced this scheme for fluvial deposits which

Code	Description
Diamictons	*Very poorly sorted admixture of wide grain size range.*
Dmm	Matrix supported, massive.
Dcm	Clast supported, massive.
Dcs	Clast supported, stratified.
Dms	Matrix supported, stratified.
Dmg	Matrix supported, graded.
Dml	Matrix supported, laminated.
- - - (c)	Evidence of current reworking.
- - - (r)	Evidence of resedimentation.
- - - (s)	Sheared.
- - - (p)	Includes clast pavement(s).
Boulders	*Particles of >256.0 mm.*
BGms	Matrix supported, massive.
BGmg	Matrix supported, graded.
BGcm	Clast supported, massive.
BGcg	Clast supported, graded.
BGfo	Deltaic foresets.
Gravels	*Particles of >8.0 - 256.0 mm.*
Gms	Matrix supported, massive.
Gm	Clast supported, massive.
Gsi	Matrix supported, imbricated.
Gmi	Clast supported, massive (imbricated).
Gfo	Deltaic foresets
Gh	Horizontally bedded.
Gt	Trough cross-bedded.
Gp	Planar cross-bedded.
Guc	Upward-coarsening.
Guf	Upward-fining.
Go	Openwork gravels.
Gd	Deformed bedding.
Glg	Palimpsest (marine) or bedload lag.
Granules	*Particles of 2.0 - 8.0 mm.*
GRcl	Massive with clay laminae.
GRch	Massive and infilling channels.
GRd	Deformed bedding.
GRh	Horizontally bedded.
GRm	Massive and homogeneous.
GRmb	Massive and pseudo-bedded.
GRmc	Massive with isolated clasts.
GRmi	Massive with isolated, imbricated clasts.
GRmp	Massive with pebble stringers.
GRo	Open-work structure.
GRruc	Repeating upward-coarsening cycles.
GRruf	Repeating upward-fining cycles.
GRt	Trough cross-bedded.
GRuc	Upward coarsening.
GRuf	Upward fining.
GRp	Cross-bedded.
GRfo	Deltaic foresets.

Code	Description
Sands	*Particles of 0.063 - 2.0 mm.*
St	Medium to very coarse and trough cross-bedded.
Sp	Medium to very coarse and planar cross-bedded.
Sr	Very fine to coarse with ripple forms.
Scr	Climbing ripples.
Ssr	Starved ripples.
Sh	Very fine to very coarse and horizontally/plane bedded or low angle cross-laminated.
Sl	Fine, horizontal and draped laminations.
Sfo	Deltaic foresets.
Sfl	Flasar bedded.
Se	Erosional scours with intraclasts and crudely cross-bedded.
Ss	Fine to coarse with broad shallow scours and cross-stratification.
Sm	Massive.
Sc	Steeply dipping planar cross-bedding (non-deltaic foresets).
Sd	Deformed bedding.
Suc	Upward-coarsening.
Suf	Upward-fining.
Srg	Graded cross-laminations.
SB	Bouma sequence.
Scps	Cyclopsams.
Ssa	Aeolian version of Ss.
Sha	Aeolian version of Sh.
Spa	Aeolian version of Sp.
- - - (d)	Dropstones.
- - - (w)	Soft sediment deformation or water escape structures.
Silts and Clays	*Particles of <0.063 mm.*
Fl	Fine lamination often with minor fine sand and very small ripples.
Flv	Fine lamination with rhythmites or varves.
Fm	Massive.
Frg	Graded and climbing ripple cross-laminations.
Fcpl	Cyclopels.
Fp	Intraclast or lens.
- - - (d)	Dropstones.
- - - (w)	Soft sediment deformation or water escape structures.

Figure 3.15 Lithofacies coding scheme. (Modified from Miall, 1978; Eyles *et al.*, 1983; and Benn and Evans, 1998).

showed that the majority (including glaciofluvial outwash deposits) could be satisfactorily described using a set of twenty lithofacies types. Each type is assigned code letters for the convenience in logging. The codes are in two parts; the first is a capital letter G, S or F which represent gravel, sand or fines (silts and clays). The second part of the code consists of one or two letters, designed as mnemonics, to describe the most characteristic internal feature of the lithofacies.

This scheme has since been extended by Eyles *et al.* (1983) to include the code designator D for diamictons, thereby creating a more comprehensive lithofacies code which can be used in the description of glacigenic deposits (Figure 3.15). The letters m and c are used to identify either a matrix-support or clast-support respectively. Eyles *et al.* (1983) stress that these terms are not intended to provide any specific value of matrix content within the diamict or the size range of matrix material. Hyphens have also been introduced in the diamicton coding to indicate that several possible combinations of internal characteristics can be displayed by diamicts (Eyles *et al.*, 1983).

The third letter of the diamict code refers to internal structures and is also designed to separate massive unstructured facies (m) from stratified (s) and graded units (g). The last letters of the code are there to stress certain aspects of the diamict which may prove to be of value in palaeo-environmental reconstruction (Eyles *et al.*, 1983). The last letters are set in parentheses to emphasise that their identification is dependent on an interpretation, and are thus an optional source of information to that represented in the main part of the objective code. In addition to the lithofacies codes, the type of contact between lithofacies is portrayed symbolically on field logs (Figure 3.16; Section 3.3.2.3).

3.3.2.3: Vertical lithofacies profiles and two-dimensional logs

The lithofacies coding described in the previous section is of particular importance when recording the geometry and position of lithofacies in the graphical form of vertical profiles and/or two-dimensional logs. These methods of lithofacies documentation have been incorporated into this investigation because of their ability to represent distinctive associations of lithofacies which characterise depositional environments. Vertical profiles are measured vertical sections on which bed thickness and lithology are represented, alongside other relevant data such as structural information and the nature of contacts between lithofacies (e.g. gradational, erosional, deformed). The modal grain size of each lithofacies can be illustrated on the vertical profile using the width of the bed as a means of representation (Figure 3.17).

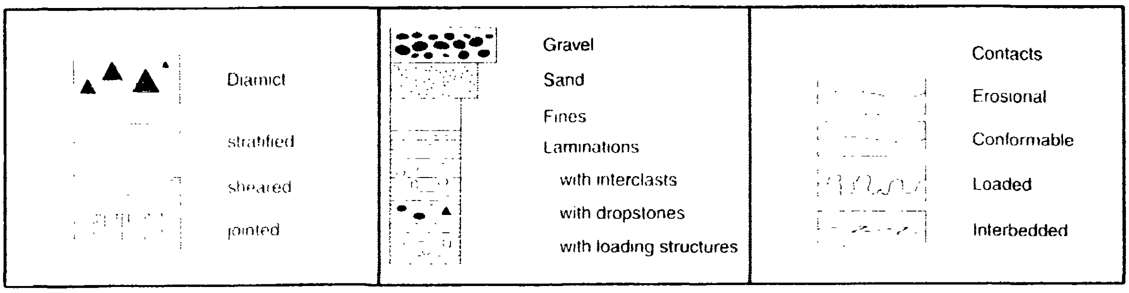


Figure 3.16 Lithofacies symbols and types of contact represented on field logs. (Modified from Benn and Evans, 1998).

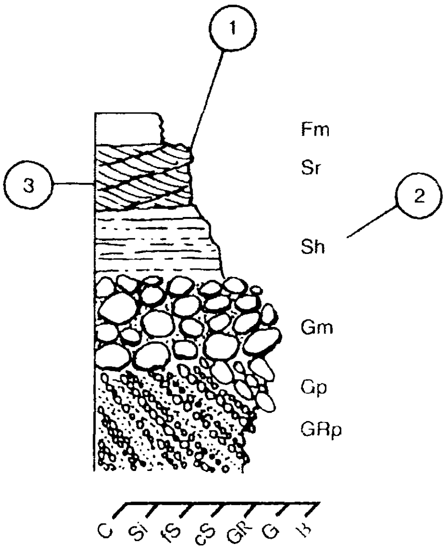


Figure 3.17 Conventions used in vertical lithofacies profiles. (1) Log width indicates modal particle size, as depicted on the lower scale, where C = clay, Si = silt, fS = fine sand, cS = coarse sand, GR = granules and G = gravel; B = boulders; (2) facies codes; (3) graphic symbols indicate sediment structures. (Modified from Benn and Evans, 1998).

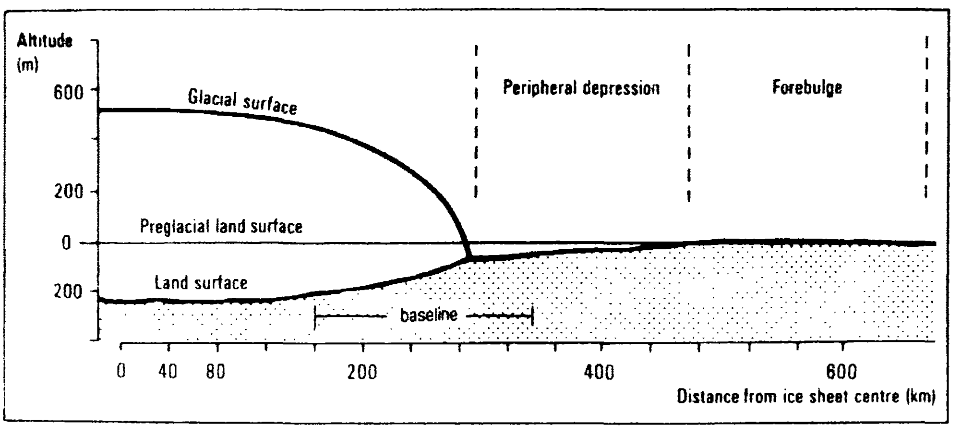


Figure 3.18 The principle of glacioisostasy, showing the depression of the crust below a 700m thick ice sheet. (From Benn and Evans, 1988).

The practice of two-dimensional logging is also carried out in order to allow information to be recorded on such aspects as the large-scale geometry of facies and facies associations, and the extent of bounding surfaces. Such information cannot be gained from vertical profile models.

3.4: Reconstructing Sea-Level History

This investigation is concerned, in part, with reconstructing the sea-level history of the north-east sector of the Irish Sea Basin by tracing and dating, where possible, former marine levels. These 'strandlines' are represented by erosional platforms and delta terraces which occur at various elevations up to the maximum height of the marine inundation (marine limit). In reviewing the previous field investigations in south-west Scotland (Sections 1.3, 1.4) the complex relationship between the glaciated coastline and the history of relative sea-level was introduced. It can be concurred, however, that where the former ice margins contacted the sea, marine transgression behind the margins could only have occurred following glacial retreat. Hence, the initial entry of marine waters directly records the Lateglacial history of that site. Moreover, an assessment of postglacial transgressions and the character of subsequent land emergence is thought to provide a large amount of information about the style of glacial recession and the recovery of the crust from its glacioisostatic depression.

3.4.1: Strandlines

On a geological timescale, the formation of individual strandlines represents a short period of time, assuming normal postglacial emergence. In the eastern Canadian arctic, Andrews (1970) noted that on average their formation takes from 50 to 300 years. Therefore, on a time scale of several thousands of years, strandlines can in theory provide good chronological markers.

Strandline development occurs when land and sea are stationary relative to one another. However, once isostatic uplift is initiated it is normally a continuous motion, such that unless eustatic sea-level rise is rapid, other sources of distinct strandline formation must be sought. Several workers have suggested that strandlines are often associated with periods of glacial stillstands (Løken, 1962; Sissons, 1963, 1967; Andrews, 1970), though this is generally no longer accepted. If the strandline marks the marine limit and terminates at an ice margin (i.e. an ice-contact delta) then this corresponds with initial emergence, and hence, deglaciation of that particular site (Andrews, 1968). The elevation of the marine limit will also give some guidance as to the magnitude of the former ice load.

3.4.1.1: Marine limit

The marine limit marks the "highest level reached by the sea on glacioisostatically depressed coasts" (Andrews, 1975, p. 181). Two types of marine limit can be defined: the *regional* marine limit, which represents the highest level attained by the sea over a large region; and the *local* marine limit, which is the maximum sea-level reached in local areas. The local marine limit may be well below the level of the regional marine limit in areas where ice has been retained during regional isostatic uplift and relative sea-level fall (Benn and Evans, 1998).

A range of landforms and sediments have been used to identify the marine limit in the field. These are: (1) the highest glaciomarine delta; (2) the highest occurrence of marine shells; and (3) the uppermost wave-cut erosional platform or notch. Subsequent lower marine levels have also be identified using a number of the above criteria.

3.4.2: Peripheral depression

Since the density of ice is approximately one-third that of the crust, the depression created beneath an ice sheet is approximately 0.3 times the ice thickness (Figure 3.18). This depression will decrease towards the ice sheet margins, where the ice is thinner, but even at the edge of the ice sheet the depression will be approximately one-twelfth of the thickness at the centre. As a result of the crust's rigidity, beyond the margin of the ice sheet the crust is deformed creating a 'peripheral depression'. Eventually the peripheral depression gives way to the 'forebulge' created by upward displacement of the crust away from the area of glacioisostatic depression. Consequently the isostatic effects of glaciation are felt and recorded outside the area actually covered by the ice. The main impact of this is that the sea can flood coastal sites in unglaciated areas. Thus, the marine limit may not necessarily, in places, record deglaciation of the immediate area but the occurrence of a 'full glacial sea' (England, 1983).

3.4.3: Relative sea-level curve

The primary method of depicting sea-level histories for particular localities is to plot a relative sea-level curve, which is a graph showing sea-level relative to today against age. Inside the Devensian ice limit a number of workers have shown that glacioisostatic uplift has a predictable logarithmic response to unloading (cf. Andrews, 1968; Walcott, 1970). Initial emergence in these cases is seen on 'emergence' curves (sea-level curves) to be very rapid (3-10 myr^{-1}), however the rate quickly decelerates towards the present (Andrews, 1968, 1970; Blake, 1975). Postglacial uplift represents an unvaried response through time, therefore the construction of sea-level curves has been used to provide ages for intervening emergent shorelines where datable material is lacking. Andrews (1968) has shown that 'predicted' sea-level curves can be derived from just a single sea-level point given its age and elevation. Such curves could, in turn, be used to estimate the age of other undated sea-level indicators in the region.

3.4.3.1: Relative sea-level curve construction

In the construction of any relative sea-level curve, there are three variables which need to be acquired: (1) proper identification of the sea-level indicators; (2) accurate measurement of their elevation; and (3) correctly dating them.

1. *Identification of sea-level indicators.* Identification of former sea-levels based on landforms and sediments, such as those mentioned in Section 3.4.1.1, is a problem which is regularly encountered. The selected elevation at a particular site can be somewhat subjective. For example, the altitude of a shingle ridge may be several metres above the sea-level, if it were formed during an extreme storm event. If unaccounted for, this measurement could give erroneously high estimates of the former marine limit. Often, the outer lip of a raised delta is used as the former sea level. However, this approach can also lead to erroneous estimates of sea-level because the lip of the delta may have been subject to backwasting during postglacial time. Furthermore, kame terraces may also be misinterpreted as marine delta terraces (Andrews, 1970; Bednarski, 1984).

   A more accurate method of pinpointing former sea-levels can be obtained from the distribution of marine sediments. An ideal case would be to obtain a datable sample from Gilbert-type foreset beds within a delta terrace, which could be traced upwards to contemporaneous topset beds. This point of intersection is commonly interpreted as the mean elevation of local sea-level at that time (Bednarski, 1984). However, the stratigraphy of the delta is not always straightforward because well developed topset, foreset and bottomset sequences may be absent or unexposed making sea-level interpretation difficult. Moreover on many coasts, marine sediment containing datable material may be absent, therefore sea-level reconstruction must rely upon morphological evidence. Whatever evidence is used, the establishment of sea-level chronologies depends on the presence of dateable materials (Benn and Evans, 1998).

2. *Measurement of sea-level indicators.* Once a sea level indicator has been identified it is necessary to measure its elevation accurately relative to present sea level. For the purpose of taking multiple readings an altimeter is generally used, however, because the accuracy of the altimeter is pressure and temperature dependent, it is important to recheck sites several

times. In order to increase surveying accuracy it is often necessary to level the elevation of a reference point at the beginning and end of altimeter surveys in order to provide an accurate datum (Evans, 1988). Also, sea level readings can be obtained before and after each sea-level indicator has been levelled, if logistically possible, allowing further calibration of the altimeter. Generally mean high tide is used as the common datum from which to measure present sea-level (Bednarski, 1984).For the most confident results a surveying level can be used which yields an accuracy of approximately 1% error on transects of greater than one kilometre (Bednarski, 1984).

3. *Dating of sea-level indicators*. In order to construct a sea-level curve strandlines, once they have been discovered and measured, finally have to be accurately dated. The main problem which exists here is that it is often not possible to obtain uncontaminated datable material in the field. The most likely method of dating a raised marine feature is through the discovery of shells in their original life position (i.e. *in situ*) within marine sediment. These shells may be radiocarbon dated and their age then related to a specific sea-level. However, it is obvious that most shells live under water and therefore relate to a sea-level which is higher than the sample collection site (Bednarski, 1984). Therefore, it is critical that their stratigraphic position (e.g. foreset or bottomset beds in a delta) be ascertained in order to determine the height of the sea-level at the time that they died. Shell samples in stratigraphically isolated sediments (e.g. silt terraces) relate to a sea-level at some unknown elevation above their collection height (Evans, 1988). These can therefore be used only as minimum dates of local deglaciation.

The samples submitted for radiocarbon dating in this investigation were prepared at the Scottish Universities Research and Reactor Centre (SURRC) radiocarbon dating laboratory, East Kilbride, Scotland, and were analysed at the National Science Foundation-Arizona Accelerator Mass Spectrometry (AMS) Laboratory, Tucson, Arizona, U.S.A.

3.5: Synopsis

This chapter has outlined the methods employed in studying the palaeo-ice sheet dynamics and depositional settings of the last ice sheet over the regions of Carrick and Galloway, south-west Scotland. It has furthermore addressed the method behind assessing sea-level chronology within the north-east sector of the Irish Sea Basin during the Lateglacial period. A summary of the main points are:

- The reconstruction of former ice sheets from geomorphological and geological evidence is based on the fact that the extent and pattern of flow of an ice sheet is recorded as a geomorphological imprint on the landscape.
- Subglacial landforms, in particular drumlins, provide the most frequent and spatially extensive evidence of ice flow in Carrick and Galloway.
- In the study of former ice sheet beds, valuable information can be gained from aerial photograph interpretation, which can then be cartographically reproduced as a geomorphological map for analytical purposes.
- Recent technical advances in the application of satellite imagery have occurred which can be favourably used to aid reconstruction of the evolution of former ice sheets from the fragmentary evidence that they leave behind.
- A winter TM Landsat image was used to interpret former ice flow directions in south-west Scotland as depicted by the orientation of drumlins and striae.
- The coexistence of subglacial bedforms of different orientation (cross-cutting flow sets) has previously been noted on satellite imagery from other regions, providing a geomorphological catalogue of different ice-flow events.
- Cross-cutting topology of sediment ridges is the result of two types of relationship: (1) superimposition, or (2) pre-existing lineation deformation.
- Two conceptual models have been developed to interpret cross-cutting lineations: (1) ice divide migration, and/or (2) lobate margin retreat.
- Relative ages of individual flow sets can be determined by using remote sensing to ascertain the correct sequence of ice-flow phases by analysing the geomorphological relationship between the lineation sets.
- Glacial depositional systems are ordered on a number of different levels: (1) temporal, and (2) spatial. This recognition has allowed the development of a powerful, systematic landsystems approach to sediment classification and palaeo-environmental reconstruction.
- The reconstruction of sea-level history is based upon the tracing and dating, where possible, of former marine levels, and the construction of a relative sea-level curve, which in turn will provide ages for intervening emergent shorelines where datable material is lacking.

CHAPTER FOUR

Ice-flow directional record and glacial stratigraphy

CHAPTER FOUR

Ice-flow directional record and glacial stratigraphy

4.1: Introduction

The aim of this chapter is to document the nature and extent of the ice-flow directional record, and to examine the glacial geomorphology and stratigraphy at specific sites throughout south-west Scotland. Although the palaeo-ice sheet dynamics and depositional settings appear to have been treated independently in this chapter, it is recognised that they should compliment each other, and therefore a combined assessment and evaluation of the facts that they both provide is reserved for Chapter 5.

Section 4.2 establishes the relative regional chronology of discrete flow sets from mapped ice-flow drift lineations (Section 4.2.1), and reconstructs a relative age sequence of these flow stages (Section 4.2.2). A summary of the principal flow stages and the nature of the ice sheet, whose behaviour they depict, is then provided (Section 4.2.3). Section 4.2.4 assesses the general character of the geomorphological evidence used by Charlesworth (1926b) to establish a readvance episode within the Loch Ryan Basin, and provides a reappraisal of the diagnostic criteria used to promote this model. Section 4.3 presents the field observations on the sedimentology and stratigraphy of sites within Carrick and Galloway and interprets the type of depositional environment responsible for the development of each sediment-landform assemblage (Section 4.3.2). A reconciliation between the glacial stratigraphy recorded in Section 4.3 and the geomorphology reported in Chapter 1 is provided in Chapter 5.

4.2: Regional Palaeo-Ice Sheet Dynamics

A Landsat satellite (Thematic Mapper) image taken during the winter season has been used as a means of identifying large scale, region-wide, patterns of glacial lineations. Using satellite imagery the author has mapped ice-flow drift lineations covering the majority of the area demarcated in Figure 1.3, which for the most part forms the regions of Carrick and Galloway (Figure 4.1). The satellite image clearly shows an extensive coverage of drift bedforms and suites of superimposed lineations of varying directions which deviate significantly from the single lineation (assumed to be the most recent) which previous field investigations by Geikie (1901) and Charlesworth (1926a) have identified. This in turn contradicts their long-standing glacial model which demonstrates a radial, unopposed dispersion pattern of lineations from an upland ice-dispersal centre (Figure 1.6, 2.13).

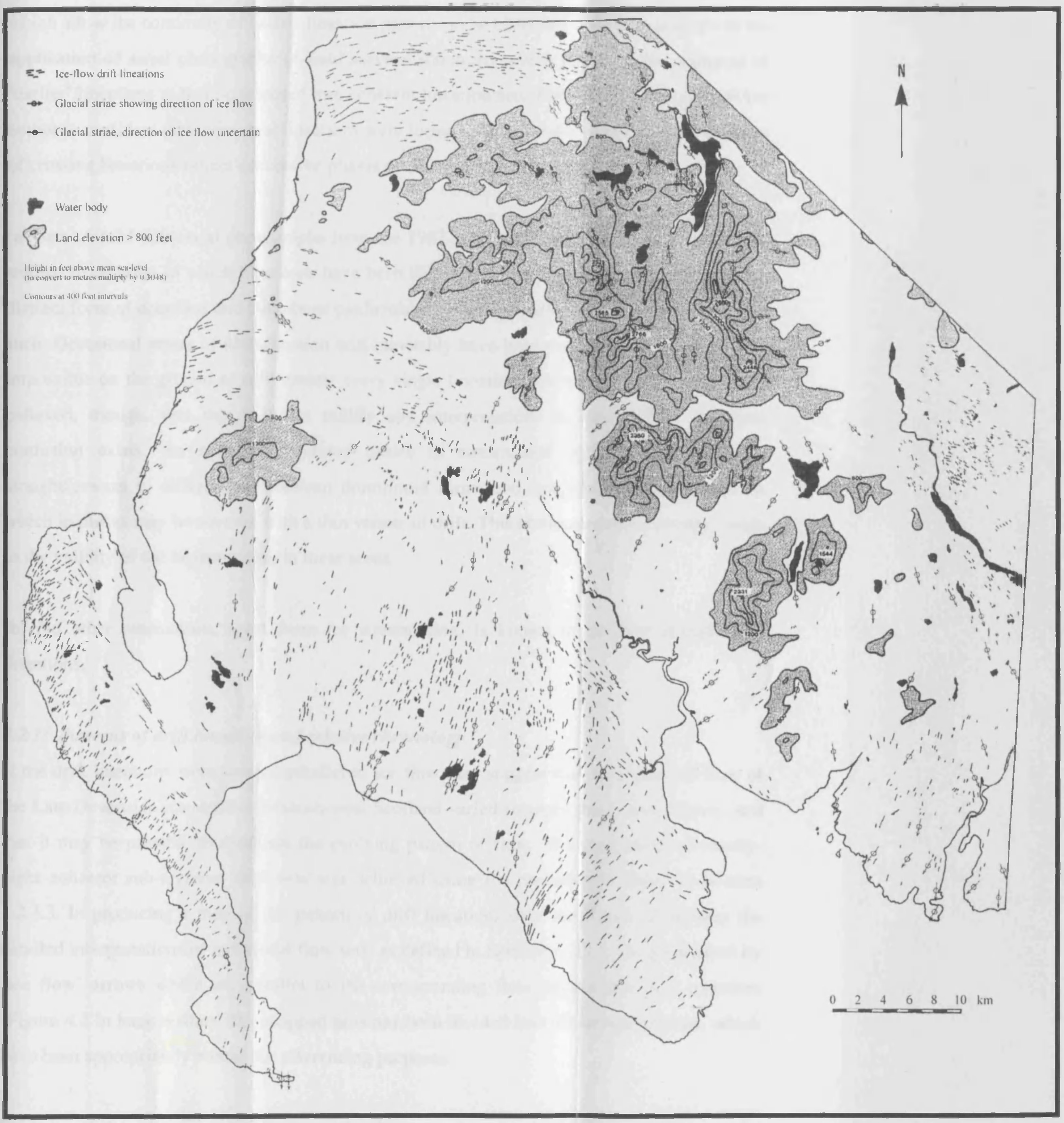


Figure 4.1 Pattern of ice-flow drift lineations observed on a LANDSAT image of south-west Scotland.

The satellite imagery has therefore permitted the observation of palimpsests of drift lineations representing former ice movements beneath the last lineation pattern. As highlighted in Section 3.2.3, this is the consequence of the very large area that such single satellite images cover which allow the continuity of earlier lineation patterns to be observed. This is in contrast to the application of aerial photographs or field surveys which invariably fail to allow fragments of 'earlier' lineations to be re-composed into coherent lineation sets. For the following reasons the author is confident that individual lineations were formed parallel to ice sheet flow and that sets of crossing lineations reflect successive phases of changing glacier flow:

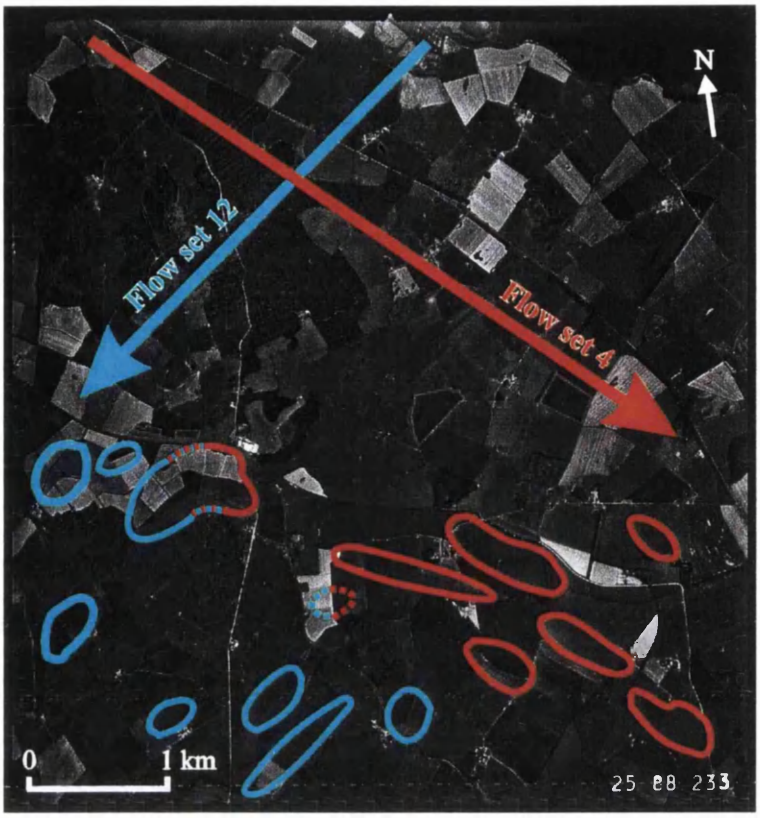
(a) Sets of 1:24 000 aerial photographs from the 1987-1989 "All Scotland Survey" have been studied in all areas in which lineations have been observed on the satellite image. All have the distinct form of drumlins and have been confirmed through extensive ground-truthing as being such. Occasional errors of identification will inevitably have been made in some areas as it is impossible on the ground to substantiate every single lineation over such a large region. It is believed, though, that this will not nullify any interpretation. In areas where structural confusion exists, the smooth, vegetated nature of streamlined drift surfaces makes it straightforward to differentiate between drumlinoid forms and irregular underlying bedrock which in places may be covered with a thin veneer of drift. This obvious contrast strongly lends to the validity of the interpretation in these areas.

(b) No other mechanism, apart from ice streamlining, is known to produce straight drift lineations.

4.2.1: Patterns of drift lineation and relative chronology

If the drift lineations were formed parallel to ice flow they suggest that the pattern of flow of the Late Devensian ice sheet over south-west Scotland varied strongly through its history, and that it may be possible to establish the evolving pattern of flow. The recognition of twenty-eight coherent sub-regional flow sets was achieved using the procedure outlined in Section 3.2.3.3. In producing a map of the pattern of drift lineations over the whole of the area the detailed interpretations of individual flow sets, as defined in Section 3.2.3.1, are generalised by 'ice flow' arrows which are parallel to the corresponding flow pattern that they represent (Figure 4.2 in back wallet). The mapped area has been divided into eleven sub-regions, which have been appropriately named for referencing purposes.

Evidence of the relative ages of lineations has been sought using 1:24 000 conventional aerial photographs. Figure 4.3 shows an example of where relative ages of lineations can be a)



b)

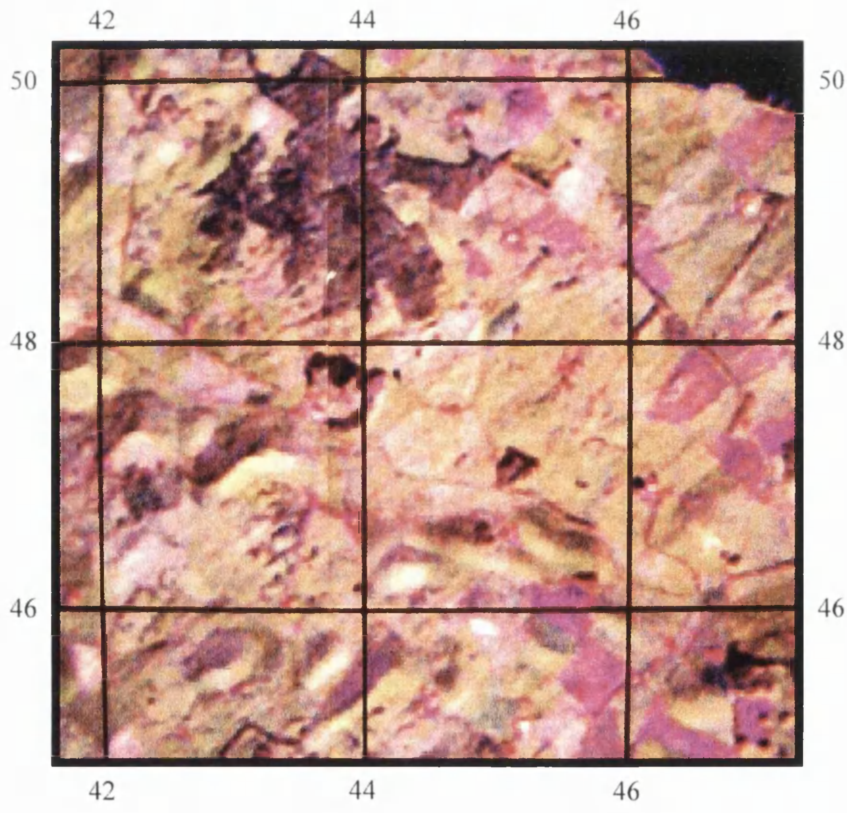


Figure 4.3 Example of where the relative age of cross-cutting lineations can be determined through examination of the reorientation of drift elements on an aerial photograph (a) and a satellite image (b). Area situated in the Machars, with eastings and northings taken off sheet 83, O.S. 1:50 000 (2^{nd} series).

established from high-resolution satellite images by the clear reorientation of elements of pre-existing lineations into new lineation sets. Where individuals, which make up a set, occur in relatively dense fields the set can be traced with confidence over considerable distances. Extensive discrete flow sets are particularly valuable in helping to determine the relative ages of less extensive sets. Figure 4.4 shows an area in which four discrete flow sets are illustrated. Using this technique the relative ages of discrete flow sets has been established using two levels of organisation in order to systemise the procedure: (1) a sub-regional chronology whereby the relative age of flow sets within a defined area are established, and (2) a regional chronology, whereby relative ages are established between sub-regional flow events (Table 4.1). Relatively restricted flow sets, where no evidence of relative age can be found, are excluded as residuals from the analysis (Flow sets 8, 9 and 20). The relative chronology in Table 4.1 indicates which of the sets could be synchronous and therefore possibly compatible with a single flow pattern (e.g. 16, 10 and 13), and which are demonstrably asynchronous (e.g. 16 and 17) and therefore of different age.

Table 4.1 Correlation of Late Devensian flow sets. Relative ages based on cross-cutting flow sets and spatial concentration. Vertical lines separate flow sets whose relative ages have not been established. Those in the same vertical column, or those separated by horizontal lines are older above, younger below. Figure 4.2 (in back wallet) shows the location of the flow sets.

	The Rhins	The Machars	Glenluce	Ballantrae	Girvan	Ken valley	Cree valley	Loch Ryan	Fleet valley	Doon valley
Older	16	10	13							
	17			19	23					
	18		14	21	24					
		11	15	22	25					
		12			26					
					27					
					28					
			Topographically constrained flow sets			1	3			
Younger						2	4	5	6	7

In the next stage of synthesis it is necessary to consider if those flow sets which Table 4.1 would permit to be synchronous are glaciologically plausible, which during the consideration of two discrete flow sets requires an assessment of the following two factors to be made: (1) their geographical relationship with respect to ice flow proximity; and (2) the likelihood of a common ice dispersal area. It is suggested that set 10, which Table 4.1 would permit to be synchronous with sets 16 and 13, is glaciologically incompatible as its flow geometry does not accord with the logical positioning of the source area at this time. On the other hand, not only would Table 4.1 permit 11 and 15 to be synchronous, but they are also glaciologically compatible because of their close proximity to one another and their parallel flow alignment. The view taken by Boulton and Clark (1990) is adopted here in that sets which can be correlated in time and which are glaciologically compatible in space should be assumed to be synchronous until further evidence indicates otherwise. This approach allows the tentative construction of a relative age sequence shown in Table 4.2 and an inferred sequence of flow stages (A-G) shown in Figure 4.5.

Table 4.2 Relative ages of Late Devensian flow stages. Flow stages based on glaciologically compatible flow sets. **Note:** The use of a dashed line is the recognised method of linking individual flow sets and does not indicate an inclusive range.

Flow stages	Flow sets
A	16-13
pre-B	14
B	11-15
post-B	12*
C	17-19-23
D-1	21-22
D-2	25-26-27-28
E	18
F	1-3
G	2-4-5-6-7

*infers the earliest possible stage of development

4.2.2: Age of the ice-flow stages

The evidence presented in Section 4.2.1 gives no indication of the age of the flow stages. Stage C clearly correlates with the maximum and stages F and G with the decay of the Late Devensian ice sheet throughout Carrick and Galloway. It is not clear if flow sets 13 and 16 of stage A both formed during the Late Devensian period or whether they are palimpsests of earlier Quaternary glaciations over south-west Scotland. The view presented in Section 2.2.1 is that throughout the northern Irish Sea Basin and peripheral onshore regions subglacial erosion resulting from ice movement during the Late Devensian has left the area devoid of any pre-existing deposits (Garrard, 1977; Kerr, 1982a). It might therefore be assumed, on the premise that no evidence of pre-Late Devensian deposits in Carrick and Galloway have been discovered, that the present drift lineations developed during the Late Devensian stage. Unfortunately, there are no stratigraphic sections within this region that would help to resolve this question. Sections along the western shoreline of Loch Ryan reveal a till which contains orientation fabrics with a mean lineation azimuth of 358.5°. Of the satellite-image lineations present in that area (Figure 4.4), the best correlation would seem to be of the till with flow set 5 of stage G.

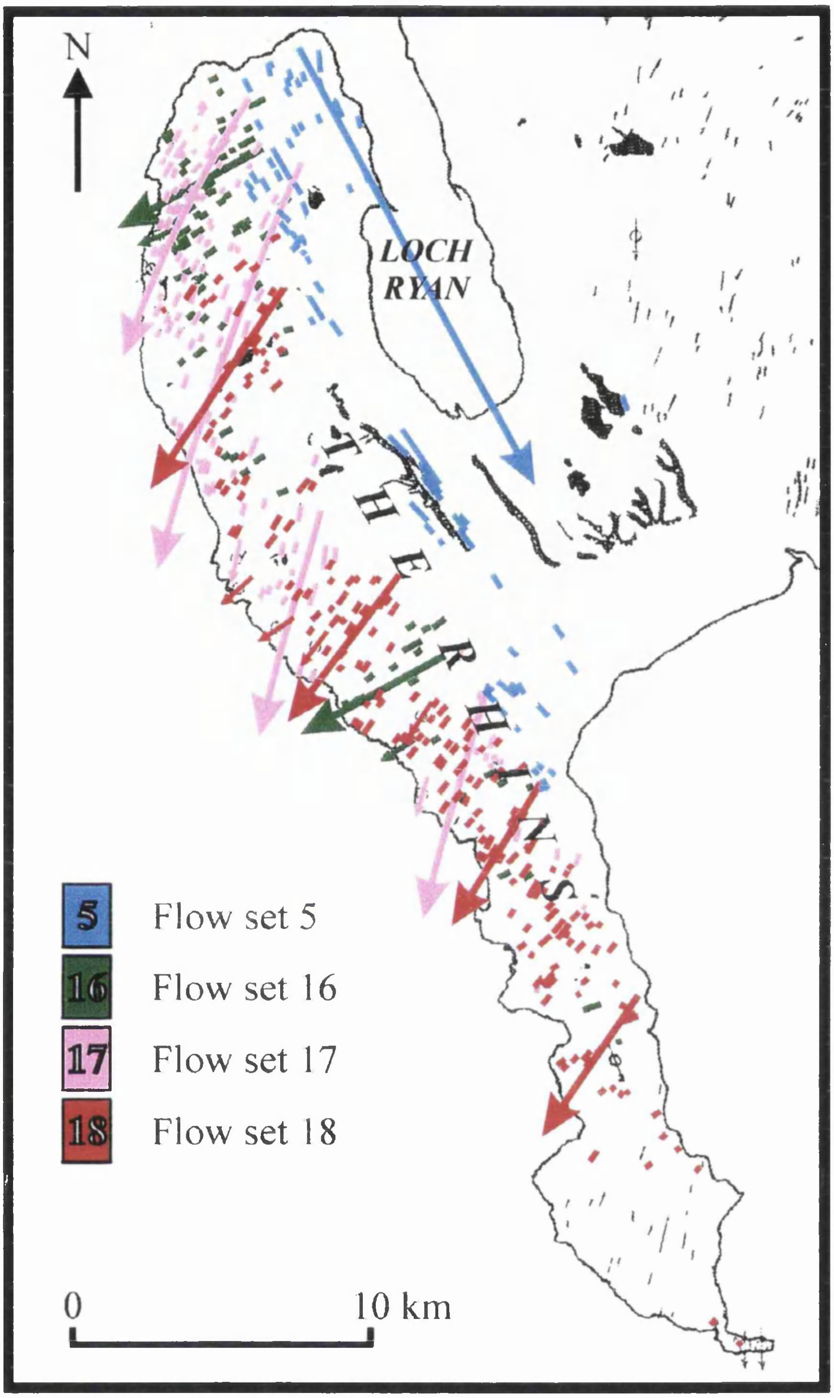


Figure 4.4 The Rhins of Galloway exhibiting four discrete flow sets.

4.2.3: Evolution of the ice sheet through the last glacial cycle

Below are summarised the principal flow stages and the nature of the ice sheet whose behaviour they reflect and the extent to which they are consistent or inconsistent with earlier findings where possible. There are two major sources of uncertainty in the foregoing account of the evolving Southern Uplands ice sheet through the last glacial cycle. The first is the assignment of individual lineations to generate flow sets. These allocations are made through the application of subjective visual pattern recognition, based on the criteria outlined in Section 3.2.3.1: (1) parallel concordance; (2) close proximity; and (3) similar morphometry. The second uncertainty is the correlation of discrete flow sets to produce flow stages. These correlations are merely compatible in terms of distribution and relative age. They do not impose absolute constraints on reconstruction (Boulton and Clark, 1990). Nevertheless, it is believed that this synthesis represents a major step forward compared with the restricted interpretations of earlier workers, particularly in view of the new insights that the recognition of cross-cutting drift lineations represents.

4.2.3.1: Flow stage A (sets 16-13)

This stage (Figure 4.5a) is represented by limited, though consistently orientated, lineations (NE-SW) which are most concentrated within two low-lying undulating areas between the south-east margin of the upland mass of Milljoan Hill (403 m) and Beneraird (439 m) in Ballantrae district, and in the southern periphery of Glenluce district (constituting flow set 13). Further to the south-west, evidence of this stage is similarly presented by two prominent lineation clusters of consonant orientation covering the north-west and west-central Rhins peninsula (flow set 16). Although there is no direct evidence of age, there are two scenarios which could explain the scarcity of lineations representing flow sets 13 and 16:

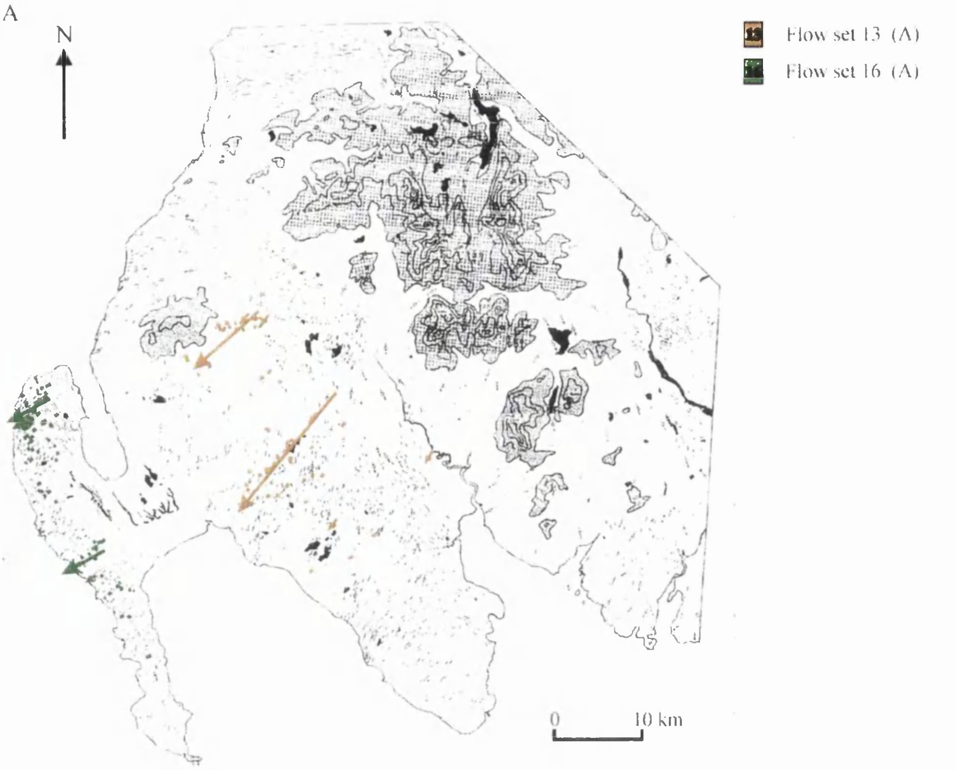
i) Stage A reflects earliest Devensian ice sheet expansion from centres of initiation in the upland massif to the north-east (e.g. Merrick, Lamachan Hill).
ii) Stage A relates to initial south-west expansion of Southern Upland ice prior to the onset of Late Devensian maximum conditions.

Both interpretations would be compatible with an ice divide ridge trending NW-SE, but whose precise location is indeterminate.

4.2.3.2: Pre-B flow stage (set 14)

There is no direct evidence for the age of this stage but the north-south aligned drift lineations appear to overprint flow set 13 of stage A and therefore this stage is suggested to post-date it

a) Flow stage A



b) Flow stage pre-B

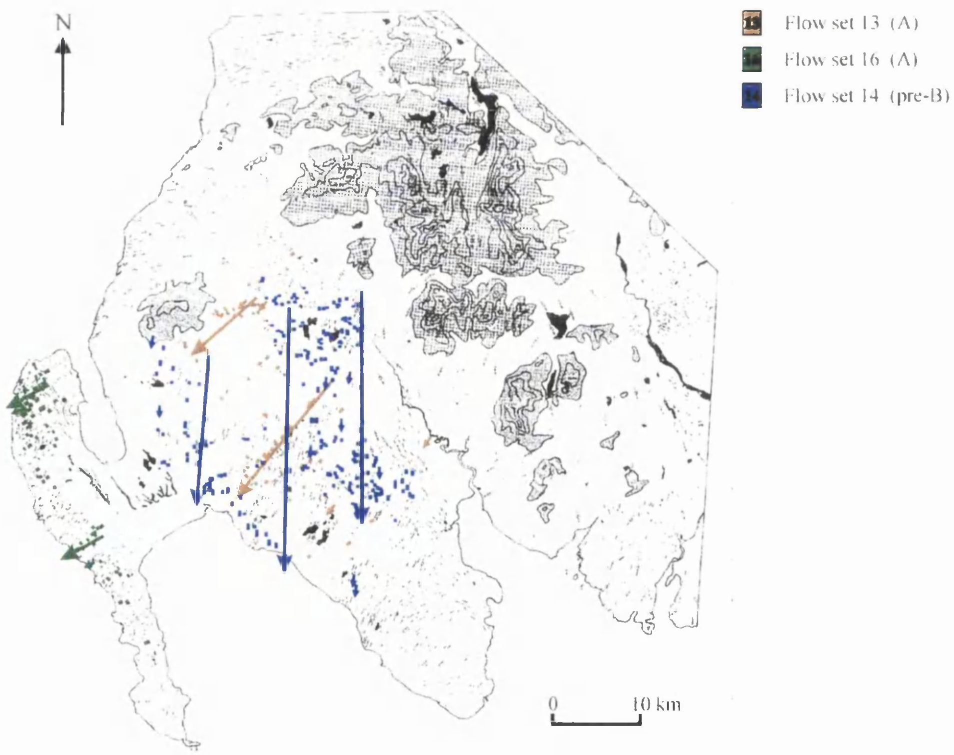
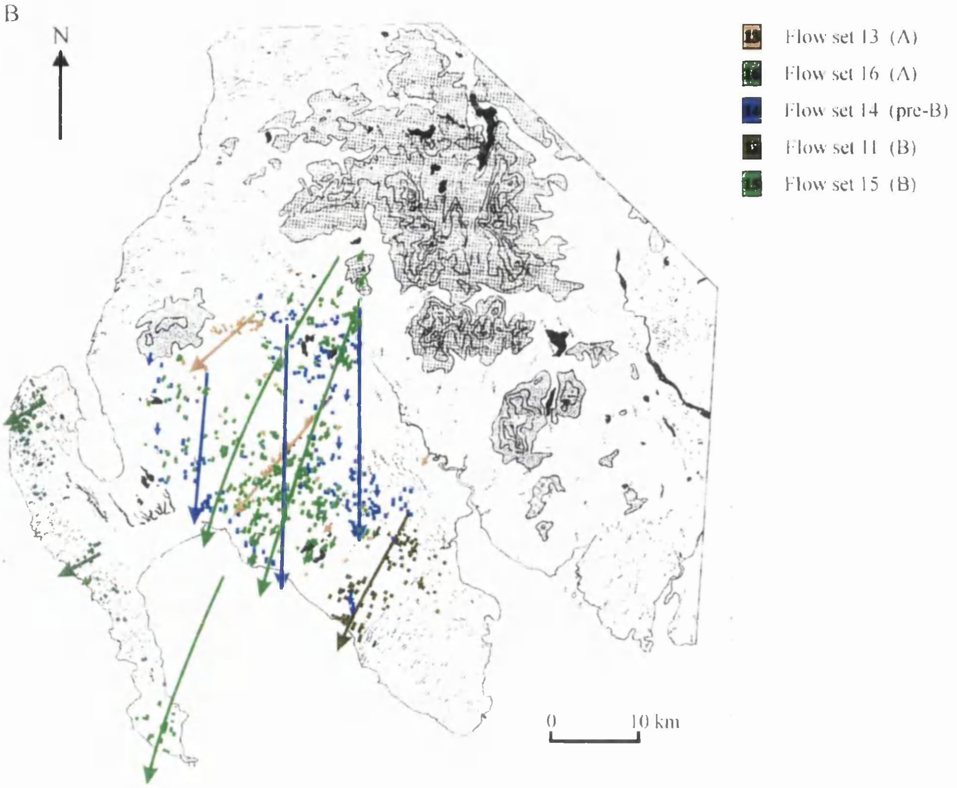


Figure 4.5 Late Devensian flow stages based on glaciologically compatible flow sets. Flow stage A (a) and pre-B (b).

c) Flow stage B



d) Flow stage post-B

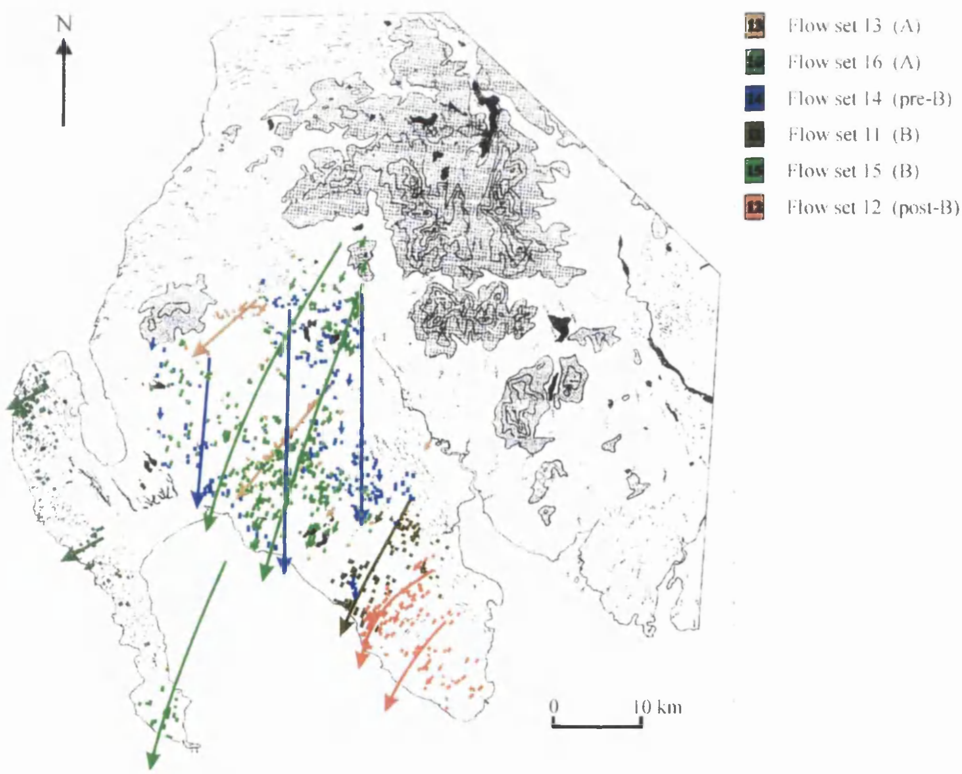
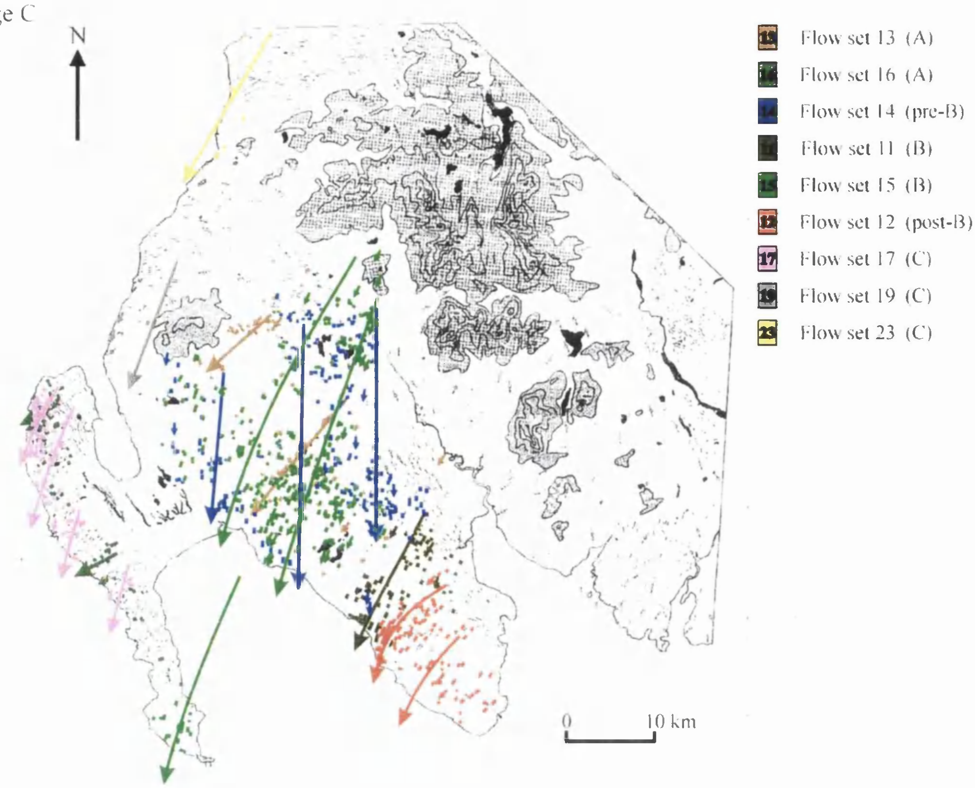


Figure 4.5 (cont) Flow stage B (c) and post-B (d).

e) Flow stage C



f) Flow stage D-1

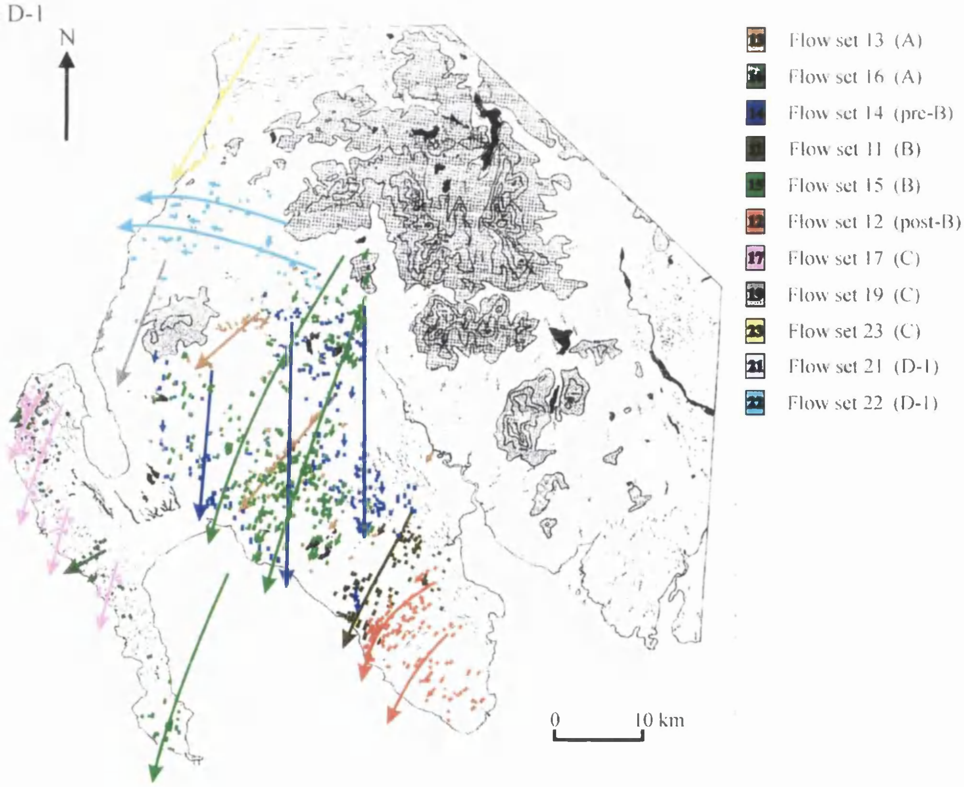
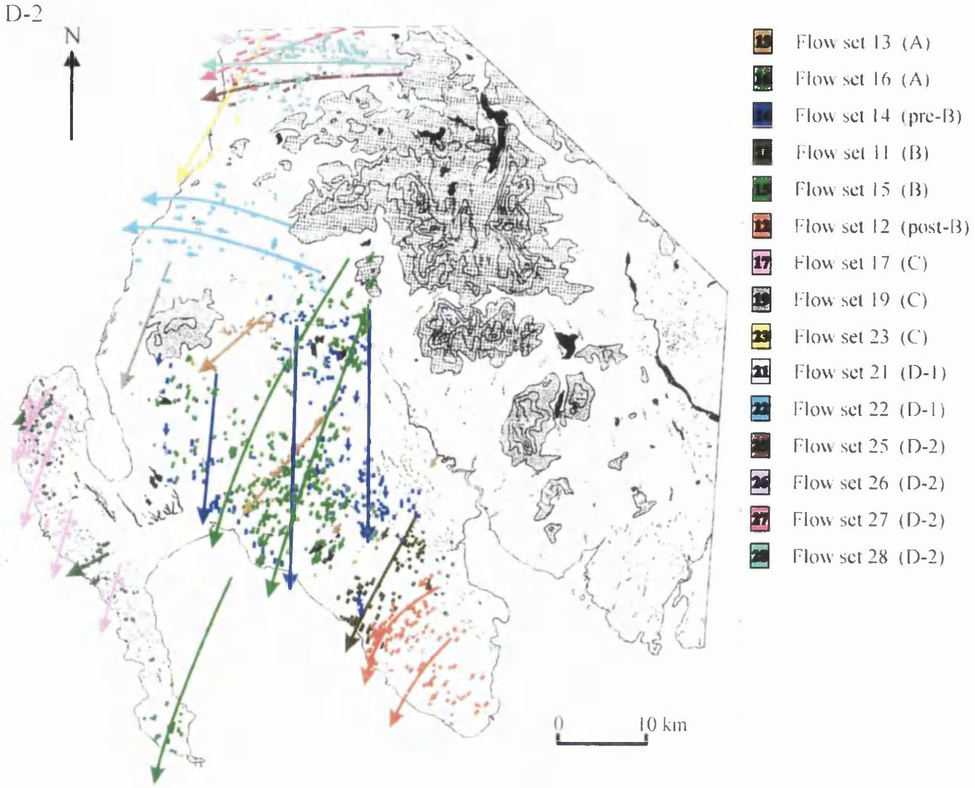


Figure 4.5 (cont) Flow stage C (e) and D-1 (f).

g) Flow stage D-2



h) Flow stage E

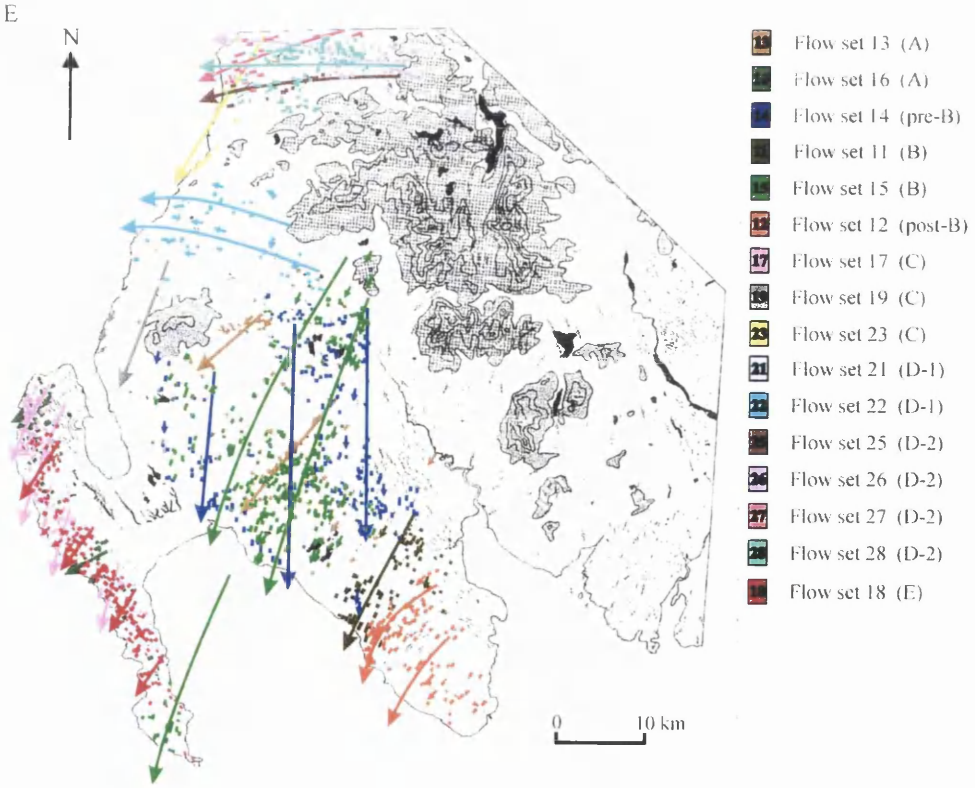


Figure 4.5 (cont) Flow stage D-2 (g) and E (h).

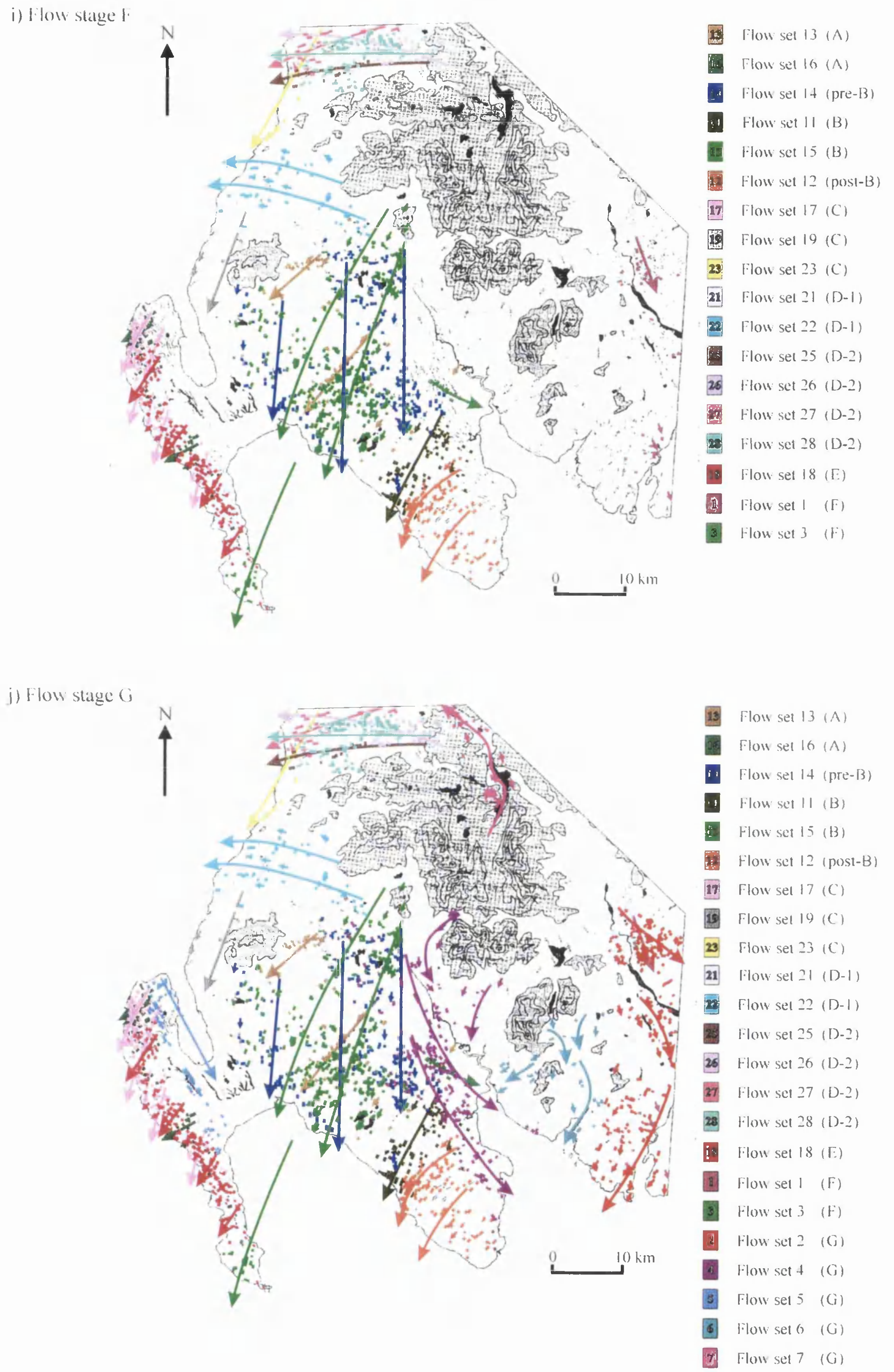


Figure 4.5 (cont) Flow stage F (i) and G (j).

(Figure 4.5b). The spatial coverage and concentration of the lineations which developed during this stage are also characteristically higher than those represented in stage A. This is suggested to be a reflection of the extent to which flow set 13 has experienced subglacial modification. It is thought that pre-B flow stage reflects an earlier flow to stage B (Section 4.2.3.3) in that it appears to display a similar dominant flow over Glenluce district but cannot be directly related to it. Two scenarios can explain the alignment of drift lineations and generalised flow direction from this stage:

i) Stage Pre-B marks the expansion of Southern Upland ice into the lowland area of Glenluce facilitated by a very dramatic shift of the flow divide to an east-west alignment.
ii) Stage Pre-B reflects the early dominance of more powerful Highland ice streaming from the Firth of Clyde in a southerly direction over the districts of Girvan and Ballantrae in part.

Acceptance of the second scenario lends support to the early inference made by Charlesworth (1926a) in which dominant Highland ice is thought to have had a strong impact on ice issuing from ice-dispersal centres within the Southern Uplands (Section 2.4.2). Among the evidence used to support this conjecture, Charlesworth (1926a, p.3) makes reference to striae which were found to "run parallel with the coast on the flanks of Pinbane Hill, at Downan Point, and south of the Stinchar fault". Charlesworth (1926a, p. 3) concludes that the confluence of Highland ice with local ice flowing in a "general south-westerly course over the Machars of Wigtownshire from Glen Trool and the Merrick Hills" was over the area of Luce Bay.

4.2.3.3: Flow stage B (sets 11-15)

Set 15 appears to reflect a dominant south-west flow from the upland area centred around Merrick (843 m) though is predominantly represented by north-north-east to south-south-west trending lineations throughout the lower elevated Glenluce district (Figure 4.5c). It is believed that flow set 15 can be linked further with similarly aligned lineations situated on the extreme southern part of the Rhins peninsula.

It is postulated that set 11 can be equated with flow set 15 if consideration is given to both the similar alignment of bedforms and their spatial distribution and concentration, particularly near to the coastal margin of Luce Bay. An equivalent lineation pattern most probably existed in the north-east sector of the Machars towards Cree valley prior to modification by subsequent flow events. The generalised flow direction of set 11 is compatible with a source area centred close to Lamachan Hill (717 m) and/or Cairnsmore of Fleet (711 m).

Stage B can be explained if there is again a shift in the position of the main ice-divide to one aligned from west-north-west to east-south-east. This might be equated to the earlier phase of ice flow forming stage A, perhaps suggesting an ice sheet whose centre of mass and flow radiation were relatively fixed but which experienced an episode of relative stability or decay prior to major expansion of ice as represented by stage B. The reorientation of the principal flow direction from south-west, near to the source area, to one more reflected by a south-south-west direction is thought to be strongly linked to the interaction of a more dominant ice stream flow to the west. This inference lends credence to the interpretation of a Pre-B stage of prominent Highland ice flowing south over the lower lying areas peripheral to the coastline, prior to the development of what appears to have been a more dominant ice flow of Southern Upland ice. This leads to two possible circumstances which might have arisen:

i) An easing off of Highland ice to the west allowing the overprinting of flow set 14 (pre-B stage) in Glenluce district by developing Southern Upland ice (sets 11-15).
ii) A coalescing and manipulation of the Highland ice by Southern Upland ice resulting in the divergence of the former ice stream into an orientation dictated by the local ice stream (i.e. south-south-west).

4.2.3.4: Post-B flow stage (set 12)

Although there is no direct evidence of its age, flow set 12 clearly demonstrates on satellite imagery an overprinting and realignment of drift lineations from previous stage B, predominantly in the area to the south-west of the Machars district (Figure 4.5d). It is undetermined as to when this flow phase was initiated after stage B, therefore it can only be confidently established that set 12 reflects a Post-B flow stage.

It is postulated that the flow convergence represented between set 11 (stage B) and set 12 (stage post-B) represents a meeting point of local Southern Upland ice (set 11) and Lake District ice (set 12) (Figure 4.6). Charlesworth (1926a) makes reference to a congestion of ice in the Solway Firth and signifies the intensifying impact of Lake District ice at this time as being recognisable by the increasingly western component of the drift lineations over this area. The sequence which is envisaged is that synchronous Lake District ice, thought to have been flowing in a west to north-west direction, coalesced and was subsequently reoriented by south-south-west flowing Southern Upland ice. There is no obvious north-east to south-west aligned topographic barrier in the area of suggested coalescence which might alternatively explain the abrupt change in generalised flow direction of set 12. Furthermore, it is inconceivable that flow set 12 reflects a single pattern of flow from a local source area as there is no obvious ice-dispersal centre, with perhaps the exception of the isolated upland mass of Criffel (569 m), which would serve as a viable dispersal centre for a dominant westerly flowing ice stream.

The extent to which Lake District ice might have encroached onto Galloway region is indeterminate. The author believes that evidence of its existence around Wigtown Bay is largely depleted due to total reorganisation of sediment by later stages of ice flow, in particular by late-stage topographically constrained ice present within the Cree (flow set 3-4), Fleet (flow set 6) and Ken valleys (flow set 2) (Figure 4.6).

4.2.3.5: Flow stage C (sets 17-19-23)

This stage is believed to reflect the southerly penetration of an extensive ice sheet originating from a source area outwith the regions of Carrick and Galloway to the north. The pattern represented by the lineation sets in the districts of Girvan and Ballantrae alone (flow set 23 and 19 respectively) is strikingly compatible with the orientation of striations measured by the Geological Survey in 1869 (NNE-SSW). These were later used as the main evidence by Charlesworth (1926a) to support the theory of a south-south-westerly flowing Highland ice sheet along the western seaboard of south-west Scotland.

The three lineation sets of equivalent alignment, which have been used to define flow stage C (sets 17-19-23), would therefore appear to be consistent with these 'early striations' (Figure 4.5e). The generalised orientation of the lineations is also compatible with evidence documented by Charlesworth (1926a) of erratic dispersal trains directed from the north, along the west seaboard, and over the Rhins of Galloway to the south-west (Figure 4.7). On the western shoreline of Loch Ryan, near McCulloch's Point [NX 048624], the author has recorded several erratics of paisanite; a medium-grained 'microgranite' specific to the island of Ailsa Craig (340 m) which lies sixteen kilometres off the coast of south Ayrshire. No other physical process can be thought of which might account for the transport and deposition of these erratics, other than through entrainment and subsequent deposition by Highland ice.

Throughout the districts of Girvan and Ballantrae, the few isolated bedforms relating to the inferred Highland ice sheet are confined to a narrow zone less than five kilometres from the coastline, but running parallel to it for approximately thirty-eight kilometres (Figure 4.8). It is thought that other lineations from stage C in Girvan and Ballantrae districts have been superseded by lineations distinguishing later flow sets, and originating from ice of more local provenance (flow stages D-1 and D-2) (Figure 4.8, Section 4.2.3.6). Despite the evidence for interaction between competing regional and local ice streams, there appears to exist two

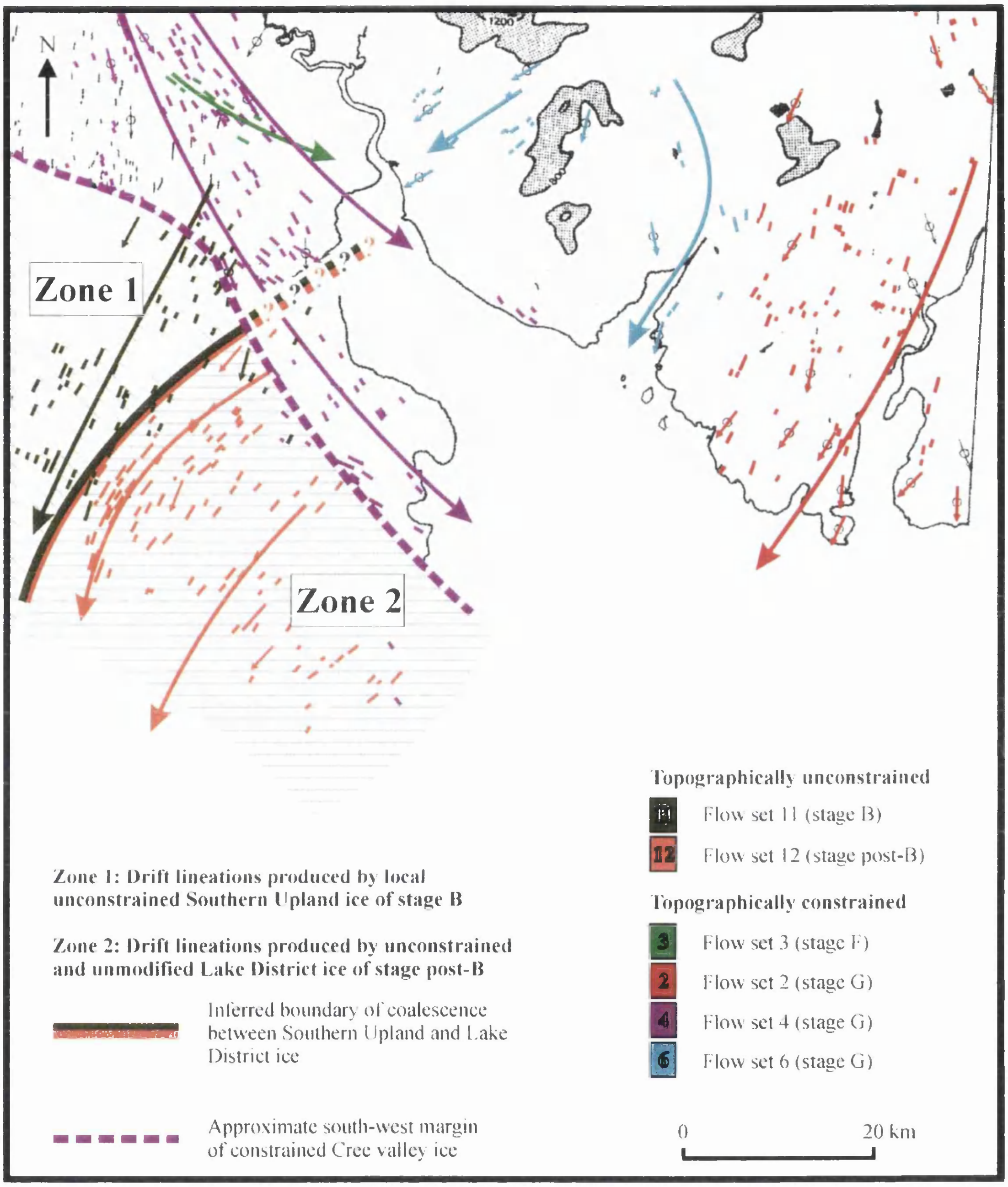


Figure 4.6 The Machars of Wigtownshire illustrating the inferred meeting point between local Southern Upland ice (set 11) and Lake District ice (set 12).

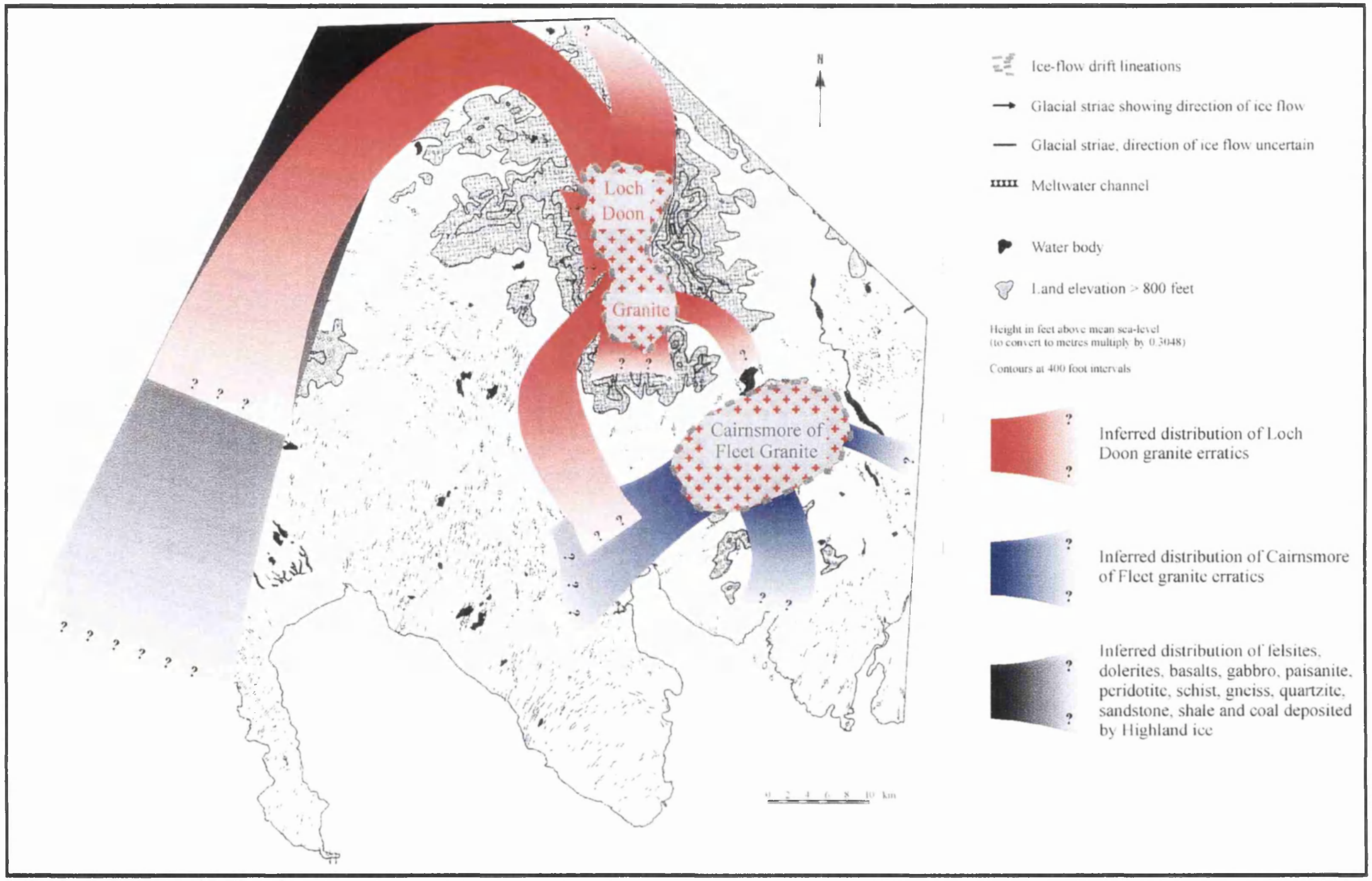


Figure 4.7 Summary of the location of the main erratic dispersal traces proposed by Charlesworth (1926a) as evidence to support the early reconstruction of glaciation in south-west Scotland.

distinct areas along the western seaboard which are devoid of bedforms representing either of the two flow patterns. The most plausible reason for this focuses upon the presence in these specific areas of a parameter, or series of parameters, which has inhibited bedform development. The author suggests that bedrock topographic obstructions, although regarded as one of the more peripheral parameters determining bedform development (Menzies, 1987), have been the inhibiting factor for the reasons outlined below.

Figure 4.8 shows a schematic diagram depicting the extent to which the Highland ice sheet is believed to have overridden the contemporary western coastline of south-west Scotland, based upon evidence of the distribution of lineations reflecting flow sets 17-19-23 of stage C. In areas displaying a scarcity of drift bedforms of Highland ice descent, and virtually no lineations reflecting Southern Upland ice, it is visible that the upland topography has imposed a topographic constraint on the flow of the lateral margin of the Highland ice sheet. In south Girvan, the predominant upland masses of Hadyard Hill (323 m), Saugh Hill (296 m), Troweir Hill (296 m) and Grey Hill (297 m) must have played a critical role in restraining the Highland ice to what appears to have been a thin coastal strip as earlier insinuated. To the south of Ballantrae district, the large upland area centred around Beneraird (439 m) incorporating Carlock Hill (319 m) and Milljoan Hill (403 m) also probably dictated, to a large extent, the coarse taken by the Highland ice stream.

It is strongly suggested that the positioning of these two distinct upland zones additionally provides a valid explanation for the existence of two coastal stretches where it is perceived that intensive overprinting of Highland ice bedforms by Southern Upland ice took place. In Figure 4.8, flow stages D-1 (Girvan district) and D-2 (Ballantrae district), which are discussed in Section 4.2.3.6, have been generalised into two zones on the basis of the distribution of lineations distinguishing each respective flow stage. It is apparent that the direction of flow of stages D-1 and D-2 have been partially dictated by the positioning of the upland areas along the western seaboard which have acted as a 'screen' to preserve lineations of Highland ice origin. Accordingly, this is strongly reflected in the location of cross-cutting drift lineations of Highland and Southern Upland ice flow. However, the development of these cross-cutting areas may only be permitted if the Highland ice sheet is to have retreated back to the north or north-west (or at least away from the seaboard) from the position portrayed in Figure 4.8. This is because of the inability of the smaller localised ice streams, at this time, to have been able to deviate the more powerful Highland ice sheet, and subsequently to override the drift bedforms which it had developed to the west of the inferred lateral position. The cleaving of the confluent Highland and Southern Upland ice masses is suggested by Charlesworth (1926a) to have

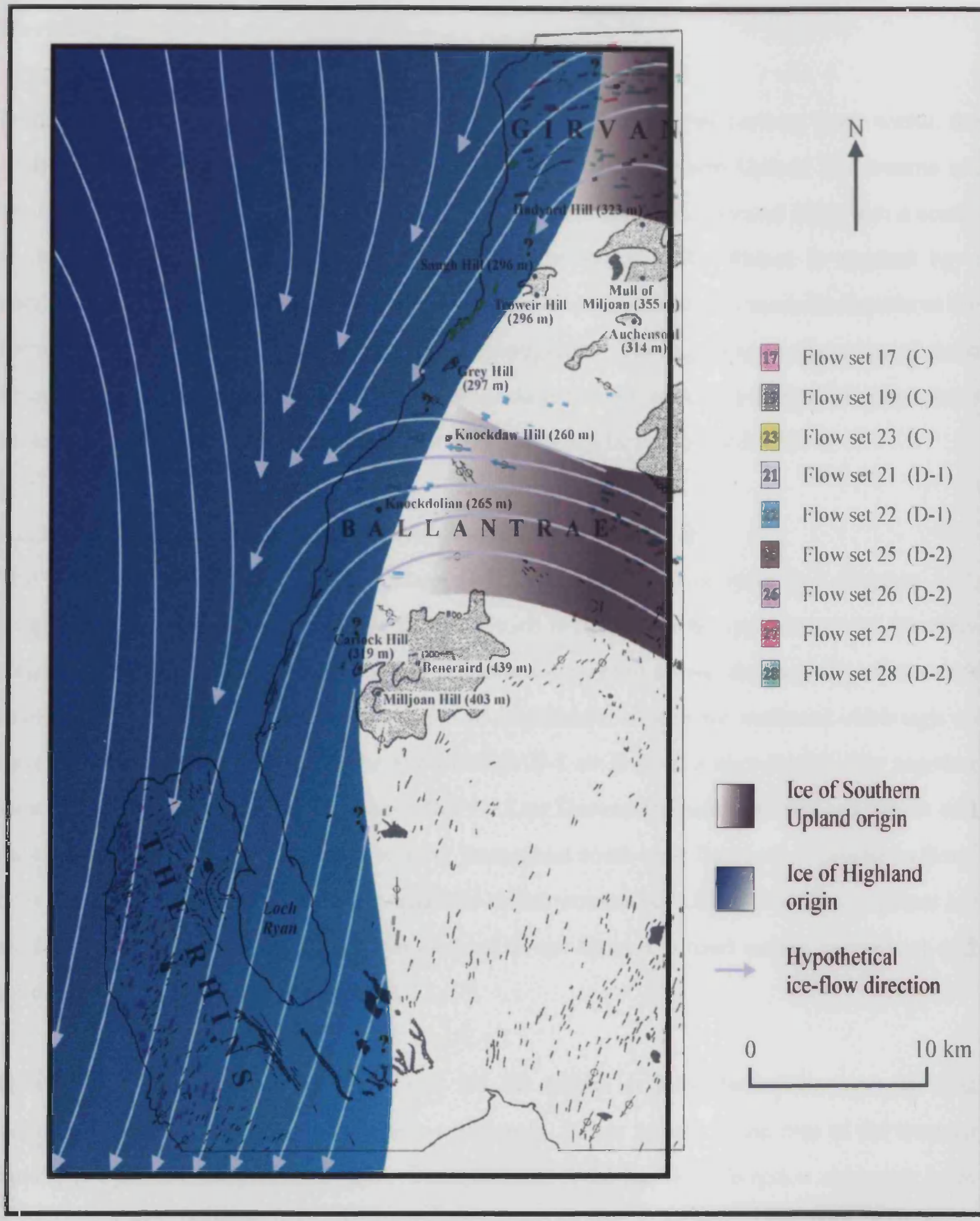
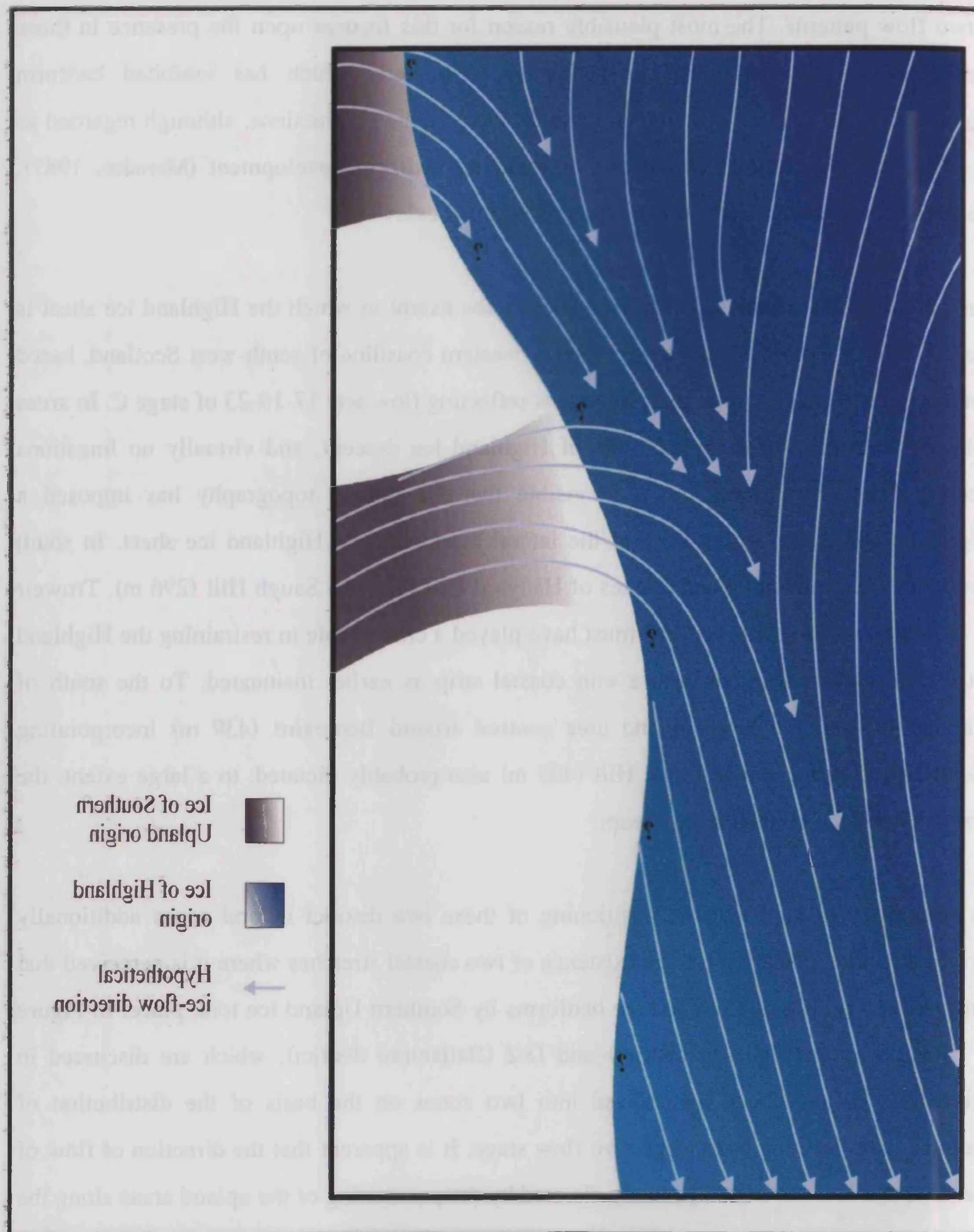


Figure 4.8 Schematic diagram representing the inferred position of the Highland and Southern Upland ice streams during the Late Devensian maximum (as inferred by the overlay), and the impact that these have had on distinctive flow sets originating from both sources.


Ice of Southern Upland origin
Ice of Highland origin
Hypothetical ice-flow direction

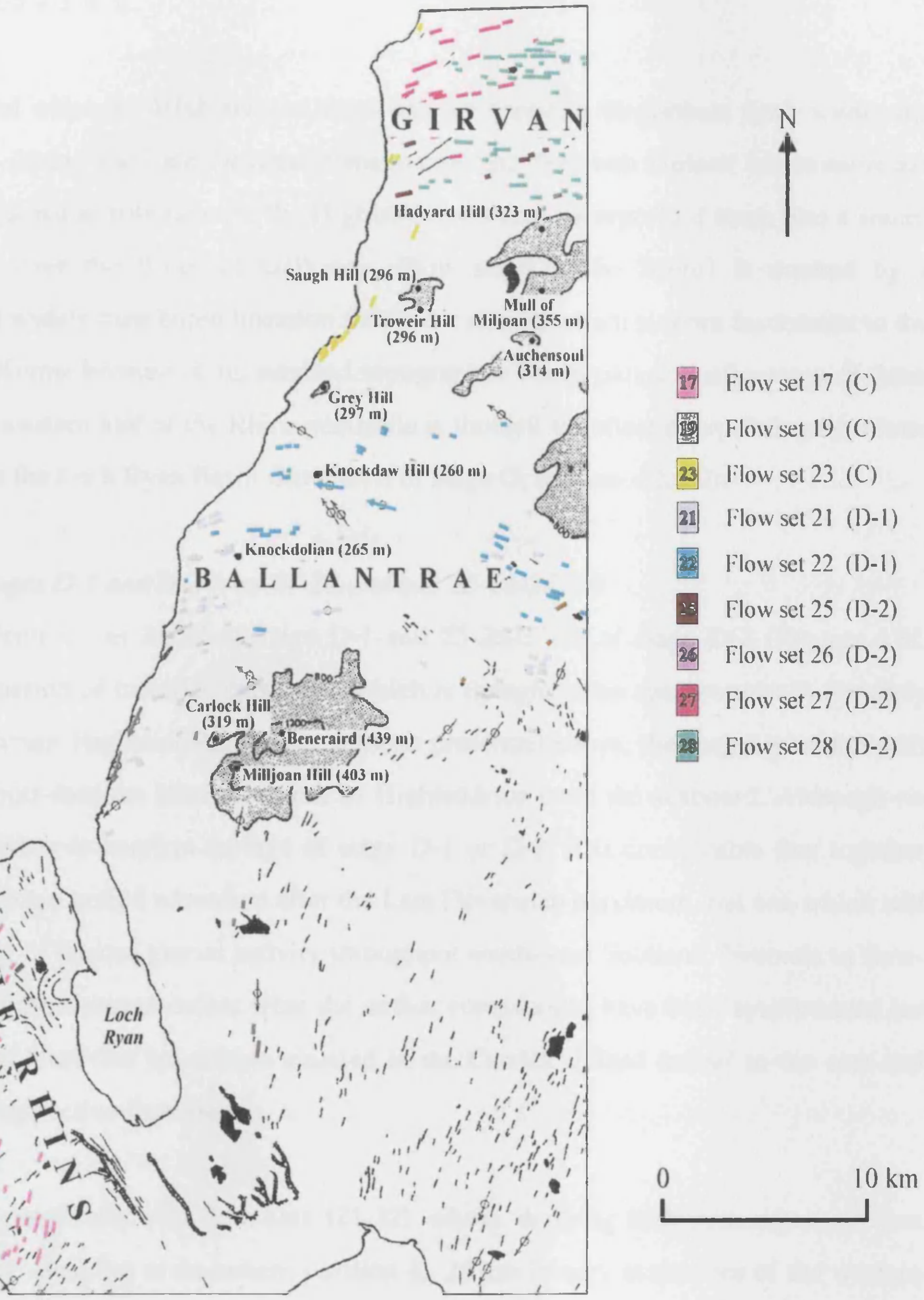


Figure 4.8 Schematic diagram representing the inferred position of the Highland and Southern Upland ice streams during the Late Devensian maximum (as inferred by the overlay), and the impact that these have had on distinctive flow sets originating from both sources.

occurred at an indeterminable position during a stage prior to a readvance of the Highland ice sheet (Sections 2.5.4.3, 4.3).

During the period when the Highland ice sheet had advanced to its furthest limit within the Irish Sea Basin, during the Late Devensian maximum, the Southern Upland ice streams are assumed to have acted as tributaries to the Highland ice, which incorporated them into a south-south-west flow over the Rhins of Galloway. Flow set 17 (The Rhins) is marked by a concentrated and widely distributed lineation field over an area which is more favourable to the formation of bedforms because of its subdued topography. The apparent confinement of these lineations to the western half of the Rhins peninsula is thought to reflect overprinting by a later flow stage within the Loch Ryan Basin (flow set 5 of stage G; Section 4.2.3.9).

4.2.3.6: Flow stages D-1 and D-2 (sets 21-22 and sets 25-26-27-28)

The flow sets identified as 21-22 of stage D-1 and 25-26-27-28 of stage D-2 (Figures 4.5f, 4.5g) indicate a period of local ice expansion which is thought to be synchronous in the early stages with maximum Highland ice, but for reasons presented above, the majority of the drift lineations must post-date the later recession of Highland ice from the seaboard. Although no direct evidence exists to confirm the age of stage D-1 or D-2, it is conceivable that together they occurred during a period sometime after the Late Devensian maximum, but one which still maintained a level of intense glacial activity throughout south-west Scotland. Normals to flow-lines generalising these stages define what the author considers to have been synchronous ice expansion from at least two ice centres situated in the Carrick upland massif to the east and south-east of the respective flow stages.

Flow stage D-1 constitutes two flow sets (21-22) whose defining bedforms cover an area extending from the coastline to an eastern position (*c.* 20 km inland) at the foot of the western Southern Upland massif (Figure 4.5f). Both sets mirror an ice flow direction changing from west-north-west, through west, to west-south-west at the coastline. This flow pattern is represented particularly by the configuration of bedforms constituting flow set 21, whose generalised flow orientation appears to have been controlled, in part, by the valley of the Stinchar which conforms to an approximate north-east to south-west alignment. This would therefore account for the deviation of the western section of the ice-flow direction, in connection with the situation of this valley axis. The overall flow geometry of stage D-1 probably relates to ice expansion from a source area centred over Merrick and Lamachan Hill, but whose precise location is indeterminate.

Flow stage D-2 is represented by four distinct flow sets (25-26-27-28) whose lineations are concentrated throughout the low-lying area bisected by the south-west flowing Water of Girvan (Figure 4.5g). A notable characteristic of the flow sets identified in this area is the apparent degree of cross-cutting which the generalised ice-flow arrows in Figure 4.5g seem to suggest. Three scenarios can provide a plausible explanation for this pattern:

i) An expansion of Charlesworth's (1926a) early hypothesis whereby ice streams flowing northwards (e.g. along the Doon valley) have been initially diverted by dominant Highland ice situated within the Firth of Clyde through west into south-west (e.g. flow set 27), and later overprinted themselves over flow sets which maintained their ice-flow direction to the west from the source area (e.g. flow sets 26-28). Charlesworth based his reconstruction predominantly on the distribution of 'Loch Doon' granite erratics (Figure 4.7).
ii) Production of cross-cutting flow sets arising from the successive stages of ice margin retreat towards the upland ice-dispersal centre to the east (Clark, 1994; Section 3.2.1.2).
iii) A reorientation of the principal flow direction over the district by migration of the main ice-dispersal centre (ice-divide), which is likely to have been positioned over the Loch Doon/Cairnsmore of Carsphairn area. A series of rapid ice-divide shifts between positions of relative stability has been suggested by Clark (1993) to induce the development of a series of superimposed forms (Section 3.2.1.2).

The author considers that the cross-cutting within flow stage D-2 can only be explained adequately by adopting the last scenario of rapid ice-divide shifts. This conclusion is based upon the continued variation in discrete flow directions which are suggested to exist throughout south-west Scotland during the last glacial period.

4.2.3.7: Flow stage E (set 18)

This shows consistent south-west ice flow into the north Irish Sea Basin from an ice centre which has shifted position since the development of flow set 17 of stage C representing south-south-west flow (Figure 4.5h). There is no direct evidence for its age but it almost certainly has been superimposed over the lineations of set 17, and so can be considered to post-date this stage. The distribution and concentration of the drift bedforms appears greater than those defining flow set 17, however they are again restricted in the northern half of the peninsula to the western side, for the same reasons as given in Section 4.2.3.5. Only in the southern part of the peninsula have the lineations been recorded throughout the whole area, though fail to extend to the southern most promontory.

4.2.3.8: Flow stage F (sets 1-3)

This stage is thought to reflect the initial phase of topographically (valley) constrained flow of more or less independent glaciers occupying the Ken (set 1) and Cree (set 3) valleys. It is uncertain whether these sets are in fact synchronous but for convenience they have been grouped together into flow stage F, as both have been substantially overprinted by a later phase of valley ice flow (Section 4.2.3.9). There is no direct evidence for the age of this stage but it is thought to be directly related to a period when local ice was receding towards the upland region in the north, and clearly was influenced by the nature of the underlying topography.

The lineations which have been categorised into flow sets 1-3 of stage F are each restricted to separate enclaves within their respective valleys, so providing very little insight into the event during which they were initiated. It could be speculated that each flow set produced a greater distribution of drift bedforms throughout the whole valley which were later remodified by a late readvance of local ice (Southern Upland Readvance?) along the Ken and Cree valleys from a position at the head of each valley. However, this scenario is uncertain and awaits further evidence. It is assumed, though, that the source area for stage F was centred around the upland areas peripheral to these valleys. Flow set 3 is thought to have been developed from ice derived from the western flanks of the Merrick range, Lamachan Hill and doubtlessly Cairnsmore of Fleet. To the east, flow set 1 appears to reflect the movement of ice along the Ken valley principally originating from the Rhinns of Kells to the north-west, Cairnsmore of Carsphairn further to the north, and possibly intensified by tributary ice flowing down the Dee valley from the eastern flanks of Merrick and Lamachan Hill.

4.2.3.9: Flow stage G (sets 2-4-5-6-7)

The pattern of radial flow shown by the lineation sets of this final stage (Figure 4.5j) is very similar to the pattern of flow associated with the reconstructed retreat of the Late Devensian ice sheet over the Southern Uplands, derived predominantly from the dispersal pattern of erratics (Charlesworth, 1926a; Figure 4.7). The author suggests that stage G equates with Charlesworth's (1926b) 'Southern Upland Readvance' of local ice within the main valleys radiating out from the upland massif: the Ken valley (flow set 2); Fleet valley (flow set 6); Cree valley (flow set 4); and Doon valley (flow set 7) (Figure 4.9). The drift lineations of flow set 5, on the eastern periphery of the northern Rhins peninsula, reflect the constrained flow of ice along the Loch Ryan Basin which have also been correlated with the Southern Upland Readvance, but are thought to have developed from the readvance of Highland ice (Section 4.2.4).

Charlesworth (1926a) proposed that the mode of recession of the Southern Upland ice, reflected by flow sets 2-4-6-7, could also be defined by the positioning of moraines and outwash fans, which were reported to be 'finely developed' in connection with all the major independent valley glaciers identified in the region (Figure 2.13) (the Cree, Fleet, Ken, Doon, and Stinchar glaciers) with other glaciers (Nith, Annandale and Upper Clydesdale) existing outwith the area of investigation. In an attempt to correlate the moraines of the widely separated valley systems, Charlesworth (1926a) adopted the Cree glacier as the standard of comparison for reasons expressed in Section 2.5.4.4. Five stages of recession were observed in connection with this glacier, and used to correspond with other stages of constrained glacier retreat in south-west Scotland. From this hypothesis, two significant conclusions were proposed by Charlesworth (1926a, Section 2.5.4.4):

i) Possibly all five retreat stages, but at least some, mark the limits of readvance of the Southern Uplands ice.
ii) To the west of Wigtown Bay, drumlins with a discordance of trend to a second set, orientated at right angles, provide evidence which supports a readvance of the Cree glacier during the 'Kirkcowan stage'. From the positioning of the two drumlin fields, and the association with nearby moraines, the two drumlin sets developed from two 'distinct glaciations' separated by a retreat of unknown extent.

The author suggests that the Southern Upland readvance episode did occur, based on the substantial superimposition of drift bedforms of flow sets 1 and 3 by those of flow sets 2 and 4 respectively, and the concordance of lineations with preserved striations presumably representing the most recent flow (Figure 4.9). However, the congruous alignment of the most recent flow sets throughout the entire length of the Cree and Ken valley leads the author to argue that this evidence recognises only one major readvance of local ice prior to gradual retreat, as represented through the five stages identified by Charlesworth (1926a, Section 2.5.4.4). The discovery by Charlesworth (1926a) of two cross-cutting flow sets to the west of Wigtown Bay is substantiated by the author (Figure 4.5j), and thought to represent the readvance of Southern Upland ice (Flow set 4) over earlier lineations developed from Lake District ice (flow set 12 of stage Post-B) (Figure 4.6, Section 4.2.3.4).

Overall , it is believed that Charlesworth's lack of insight into the widespread cross-cutting of lineations, demonstrated in Carrick and Galloway by the author, has inevitably led to a reconstruction of the Late Devensian glaciation which is somewhat over-simplified, and fails to accord with the multitude of ice flow phases which have been shown to have occurred at

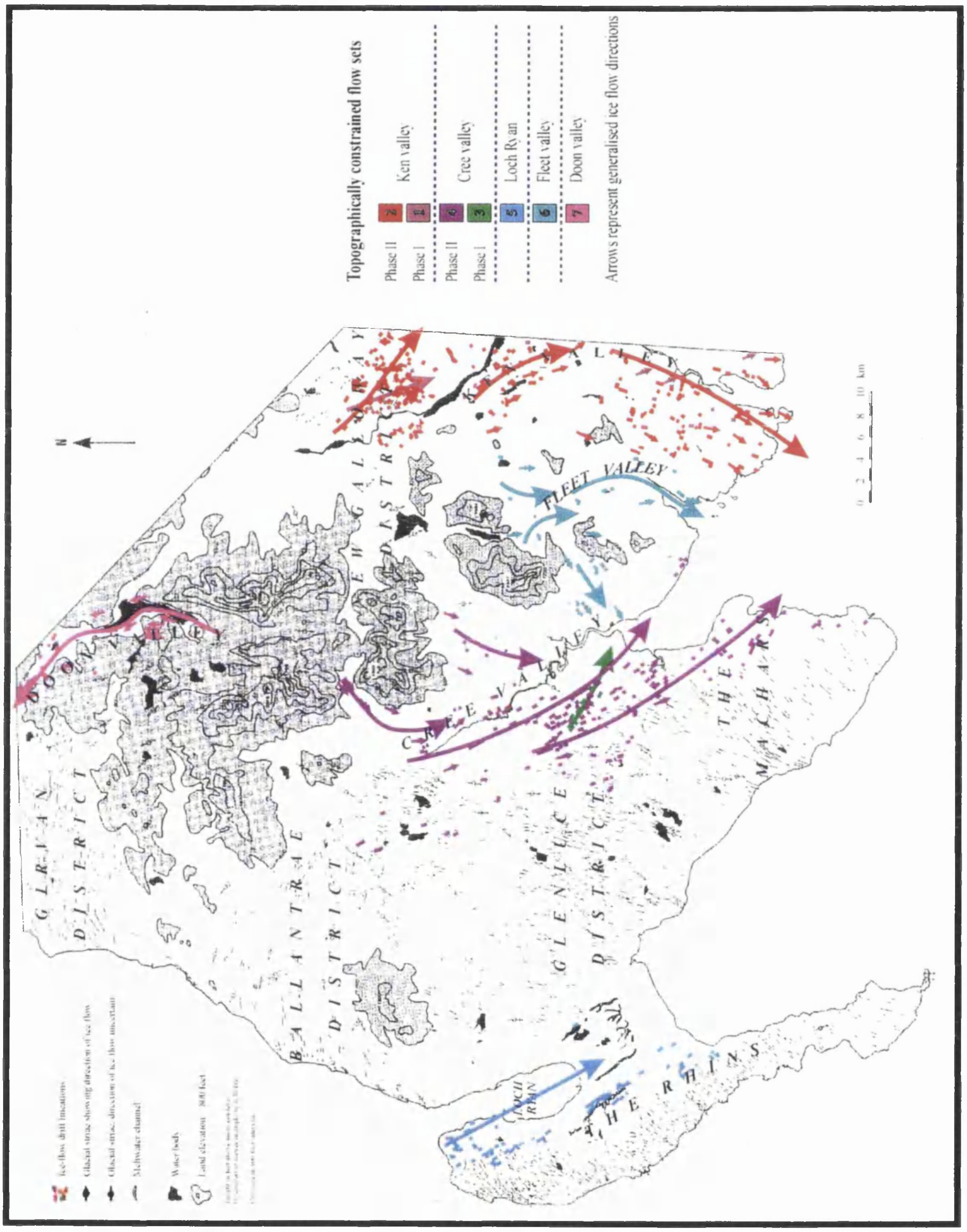


Figure 4.9 Topographically constrained flow sets within south-west Scotland.

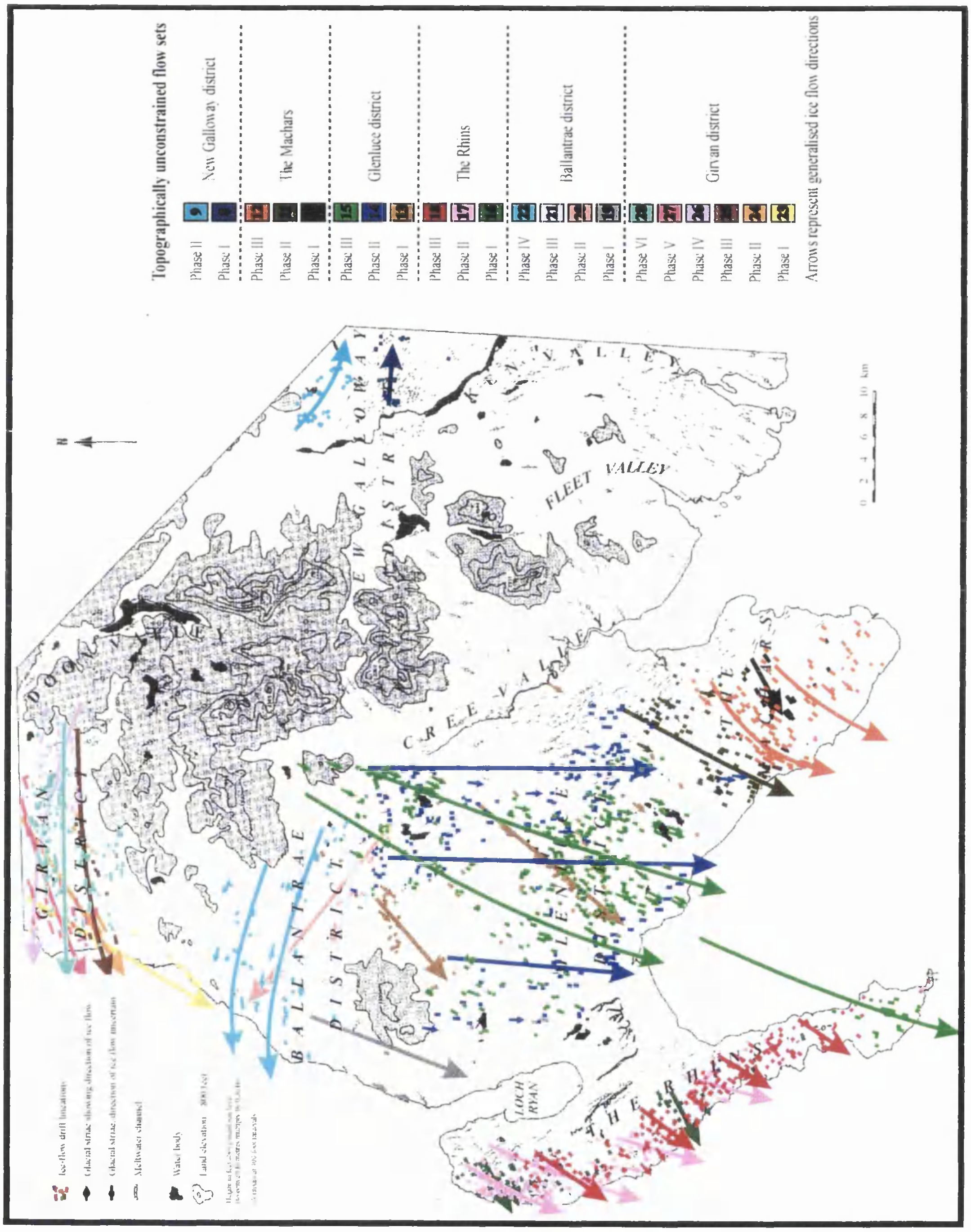


Figure 4.10 Topographically unconstrained flow sets within south-west Scotland.

various phases during the last glaciation. It appears that from the outset, the major moraine systems detected by Charlesworth have been assigned specifically to development under the influence of constrained valley ice, and never has there been any consideration to the viability of their positions relating to the recession of unconstrained ice flow, discordant to the trend of main valley axes (Figure 4.10).

4.2.4: The Highland–Southern Upland Readvance

As introduced in Section 2.5.4, it has been claimed by various workers (e.g. Trotter, 1929; Gresswell, 1967; Huddart, 1977, 1981b, 1991, 1994) that marginal ice readvances during the retreat from the Late Devensian maximum were possibly common within the Irish Sea Basin. By contrast, several workers have disputed the evidence used to infer such readvances arguing that it is insignificant, and instead simply reflects short-distance, localised oscillations of the ice margin during deglaciation (e.g. Thomas, 1985). Other workers have completely dismissed the need to invoke any glacial readvances during Irish Sea ice sheet recession arguing that the entire concept is largely contrived from the personal aspirations of the researchers involved (e.g. Pennington, 1970; Mitchell, 1972; Sissons, 1974).

In the regions of Carrick and Galloway, south-west Scotland a sediment-landform assemblage (Lammermuir–Stranraer moraine) has been reported which is suggested to mark the former ice-marginal limit of a synchronous readvance event of both Highland ice (Highland readvance; Charlesworth, 1926b) and Southern Upland ice (Southern Uplands readvance; Charlesworth, 1926a) (Section 2.5.4.3)*. Many workers have come to accept this moraine as representing an important phase in the deglaciation of south-west Scotland (e.g. Goodlet, 1970; Greig, 1971) though despite this, almost all local workers in this area, with the exception of Charlesworth, are opposed to the existence of such a stage (e.g. Cutler, 1979; Price, 1975, 1977).

This section reflects upon the observation of distinct drift lineations on the eastern periphery of the northern Rhins peninsula (flow set 5 of stage G; Figure 4.5j, 4.9) which the author argues suggest late-stage flow of topographically-constrained ice (readvance?) along the Loch Ryan basin. In order to substantiate the occurrence of a readvance event in this area, the geomorphological criteria which was original used by Charlesworth (1926b) in formulating the Highland readvance model is addressed once again from an objective standpoint. A review of the general character of the readvance moraines within the Loch Ryan Basin and the diagnostic criteria used by Charlesworth to promote this model are addressed in Sections 4.2.4.1 and

* The readvance of both Highland and Southern Upland ice has been termed the 'Highland–Southern Upland readvance' by the author so as to express their synchronism.

4.2.4.2. Section 4.2.4.3 provides a modern reappraisal of the landform distribution throughout the Loch Ryan Basin and considers their implications with respect to the nature of the palaeo-ice dynamics which initiated development. A comprehensive discussion of the findings is presented in Chapter 5.

4.2.4.1: General character of the readvance moraines of the Rhins of Galloway and Stranraer

In Chapter two (Section 2.5.4.3; 2.5.4.4), reference was made to Charlesworth (1926b) who has identified at Stranraer and in the Rhins of Galloway to the west, a complex 'kame-moraine' (the western section of the 'Lammermuir–Stranraer Moraine') which is suggested to have developed marginally to a major southerly readvance of Highland ice demarcating its position ('Highland Readvance') (Figure 4.11). The morainic deposit within the Loch Ryan Basin has additionally been correlated with the more localised 'Southern Upland Readvance' (Charlesworth, 1926b), whose characteristics have been described earlier (Section 2.5.4.4; 4.2.3.9).

To reiterate, Charlesworth (1926b) recognised that the internal constitution of the moraine varied considerably, consisting for the most part of 'water-worn sands and gravels' that exhibited distinct stratification where they were associated with a high degree of drainage. Externally, Charlesworth (1926b) also provided an impression of a 'chaotic composition with many genetically different formations', with superimposed 'pitting' (kettle-holes) reflecting the former position of stagnant ice masses. The surface has been described as 'rougher' where gravel was discovered to be the main constituent, while being characterised as more 'softened' where fine-grained till predominated. In areas where sand was considered to be the major component, a characteristic dimple effect associated with this substrate was recorded on the surface. Prominent meltwater channels have also been described throughout the area, although the presence of eskers associated with the channels of subglacial streams were found to be rare.

Upon reflection of the observed principal characteristics, a complex and dynamic depositional environment appears to have been envisaged by Charlesworth (1926b, p. 26) as depicted in the following summary:

> "The moraine partakes of the nature of true moraine, kame- or kettle-moraine, overwash and outwash fans, lake deltas and floor deposits. Streams descended the ice-margin, subglacial rivers formed eskers at their place of emergence or possibly within their walls, debris rolled down or was shot over the face of the ice, while further confusion was added by subsequent stream erosion and deposition, by the overriding of earlier deposits during temporary readvances or oscillations, with the resultant erosion, disturbance, stau-effects, the

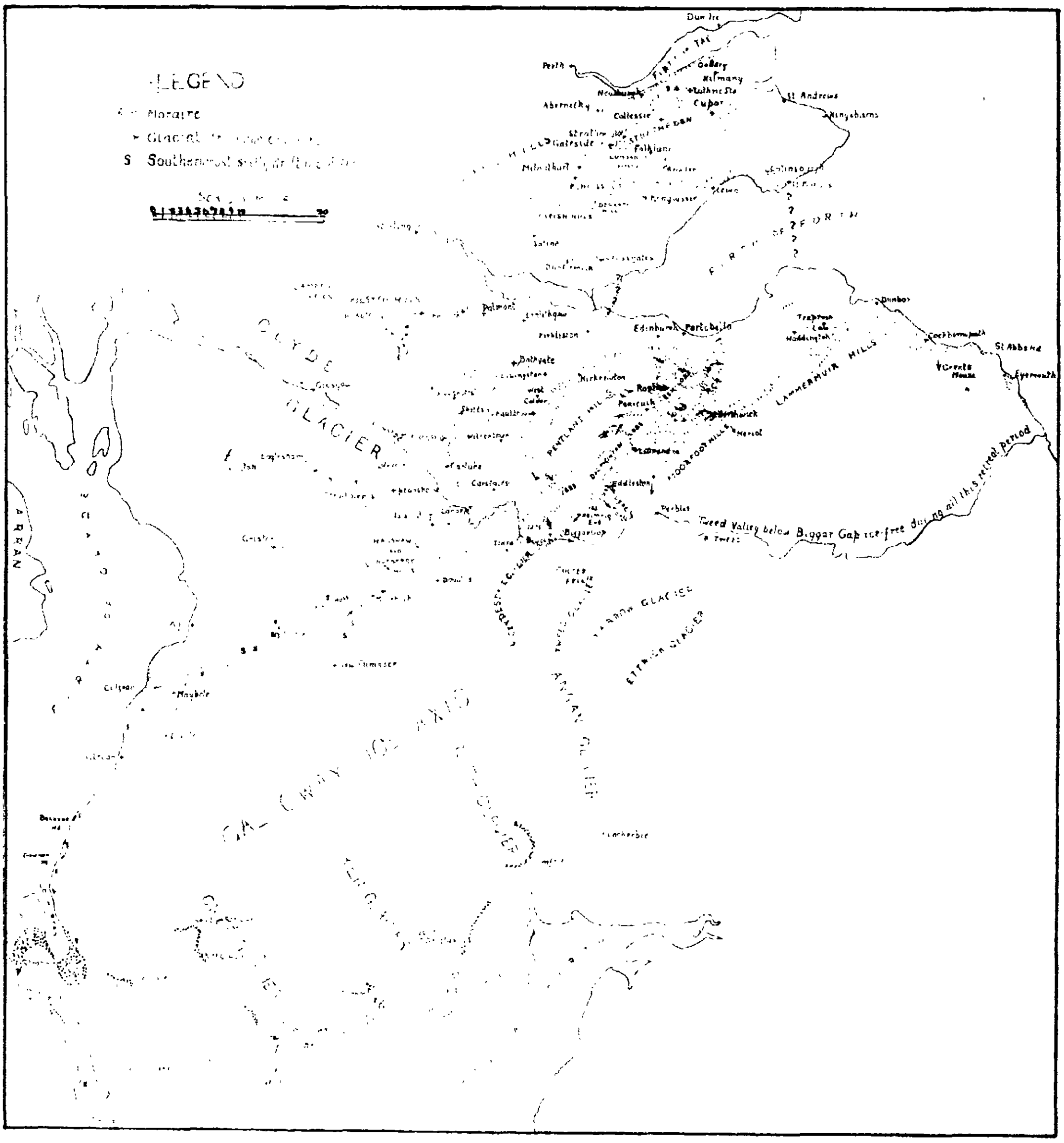


Figure 4.11 Charlesworth's map of the south of Scotland showing the position of kame-moraines. In Galloway region are shown the moraines of the "Kirkcowan stage". (From Charlesworth, 1926b).

> intercrossing of single ridges, and the covering of earlier morainic ridges by later outwash materials".

Oscillations within the moraine belt were suggested by Charlesworth (1926b) after the discovery of outwash sands and gravels intercalating with till, suggesting possible overriding of the proglacial deposits by subsequent ice-margin advance. Charlesworth exercised caution when applying this interpretation, suggesting that the evidence did not necessarily imply oscillations of the ice, but that in this region they seem to be of this origin. Such oscillations are considered to have been further substantiated by the following evidence:

1. dissection by marginal streams of earlier deposits from the initial readvance phase;
2. deepening of previously abandoned marginal channels;
3. 'choking' of the marginal channels at their mouths by moraines; and
4. the entire obliteration of the marginal channels by subsequent infilling.

Upon gradual retreat of the readvance ice, Charlesworth (1926b) anticipated that the dissection of earlier deposits occurred by marginally directed streams, with resultant production of residual kamiform ridges and mounds, and the redeposition of the material as deltas or alluvial fans.

4.2.4.2: Early evidence of the readvance origin of the Rhins of Galloway and Stranraer moraines

In the original reconstruction of the Highland readvance, Charlesworth (1926b) suggests that the northern ice over south Ayrshire was confluent with westerly-flowing local glaciers of the Southern Uplands; a hypothesis based mainly upon the absence of lateral moraines of this episode. If this scenario is accurate, it would be logical to assume that following the expansion of Highland ice over the Rhins of Galloway during the pre-readvance period (flow set 17 of stage C; Figure 4.5e), it later receded to an indeterminable position to the north of Loch Ryan. In order, however, for the ice sheet to retain its confluence with local Southern Upland ice (flow sets 21-22 and 25-26-27-28 of the synchronous D1-D2 stage; Figure 4.5g), it is presumed that the Highland ice failed to retreat to a latitudinal position much further than 55° 04′N.

It is suggested by Charlesworth that only in areas where the Highland ice was allowed to maintain a 'free edge', did it display conspicuous moraines marking the limit of the ice margin. This is thought to have been observed at the head of Loch Ryan where the moraine complex conforms to what has been described as a 'broad, hummocky belt' extending on to the foot of the hills to the north-east of Stranraer. Although it is commonly visualised as one great morainic belt, Charlesworth (1926b) indicated that it was possible to single out individual ridges trending across the basin from side to side. Beyond the readvance limit, a 'great outwash plain' to the south is also thought to have been observed, grading imperceptibly into a defined '60-foot beach' (18.3 m). The following evidence is also presented by Charlesworth (1926b) to substantiate the interpretation of a kame-moraine with outwash, and to reject the alternative suggestion of a beach origin:

1. A southerly direction of stratification throughout the deposit, which in the northern section of the deposit appears inconsistent with a beach origin.
2. The occurrence of the 'wings' of the deposit at elevations above those normally regarded as the maximum altitudes for the Lateglacial sea-level.
3. The occurrence of undoubted moraines over the northern Rhins peninsula, occurring up to 300 feet or so above sea-level.
4. The difficulty in initiating such a beach deposit across the present isthmus due to tidal scour, assuming that Loch Ryan and Luce Bay were continuous (Section 1.3.3.1). Following the filling in of the isthmus by the morainic deposits and outwash, it would be easy to preserve the formation of a beach, as seen in the present conditions.

Charlesworth (1926b) also observed what were thought to be moraines corresponding to those at Stranraer in the Rhins of Galloway, the general trend of which is illustrated in Figure 4.11. A moraine on the eastern seaboard of the northern Rhins was believed to join the Stranraer complex near Leswalt [NX 019639] and thought to continue through west and south-west to Lochnaw castle [NW 992628], Balgracie [NW 987607], and south to Portslogan [NW 987587].

4.2.4.3: Modern reconstruction of the Highland–Southern Upland readvance

In a reappraisal of the landform distribution throughout the Loch Ryan Basin the author has accurately mapped the major glacial landforms using the procedure detailed in Section 3.2.2 (Figure 4.12). A description of each mapped landform and its distribution pattern are discussed below.

1. Lateral moraines. The lateral moraine discussed here is clearly identified on aerial photographs as a dominant large-scale feature occupying an area on the south-western side of Loch Ryan (Figure 4.12). It is arranged in such a way which reflects a strong degree of containment of former Loch Ryan ice (flow stage G; Figure 4.5j) within the basin by the north-north-west to south-south-east upland spine on the Rhins peninsula, to which this linear moraine is parallel. The morphometry and positioning of the moraine indicates that it

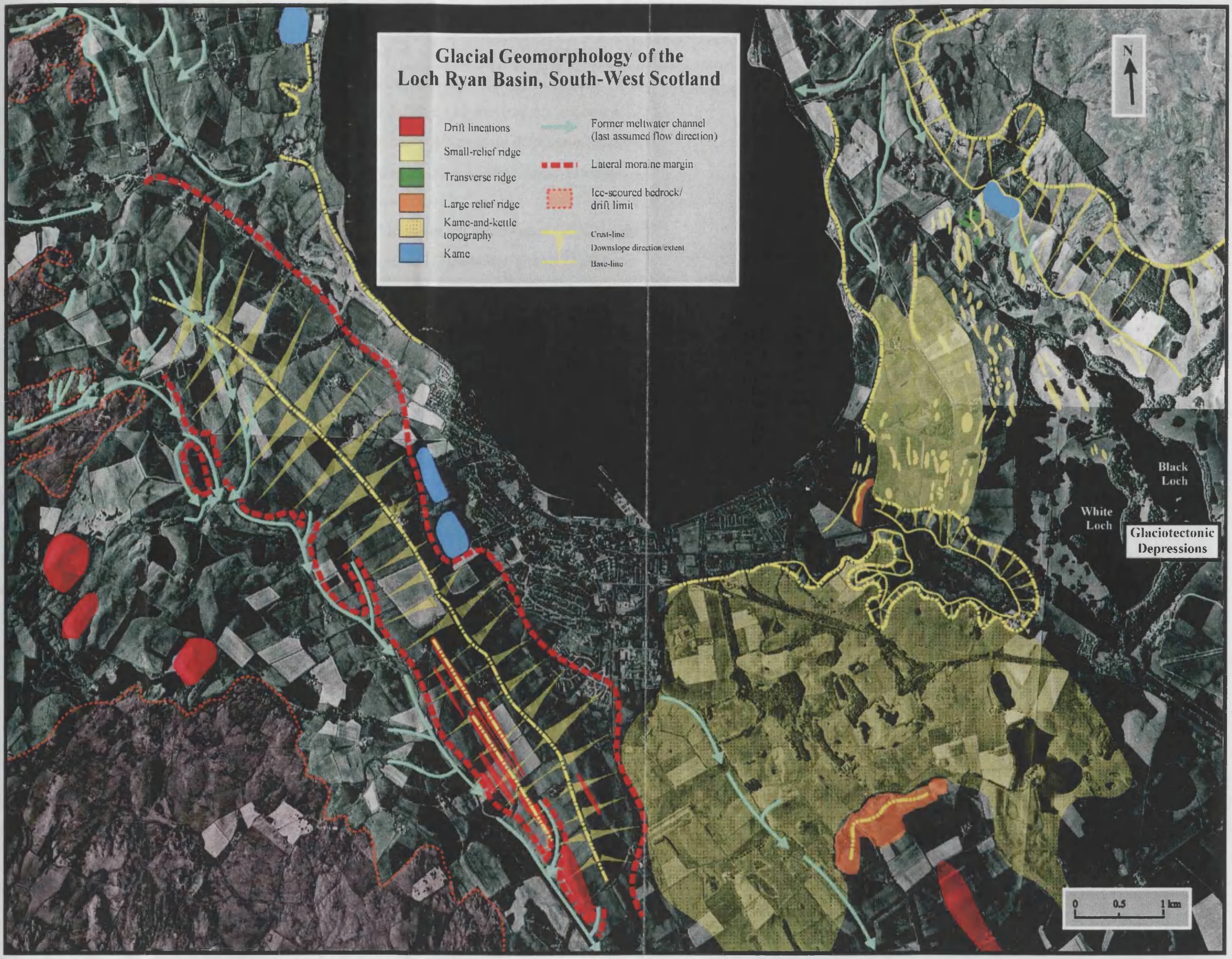


Figure 4.12 Modern evaluation of the glacial geomorphology of the Loch Ryan Basin

undoubtedly developed in association with a lobe of ice draining south-eastwards along the present Loch Ryan Basin from an ice sheet situated to the north. The landform in its entirety displays a single ridge (crestline) along its length, but to the south-east contains adjacent parallel ridges of similar elevation reflecting modification and streamlining of material at a later stage (Point 3 below). On the eastern seaboard of Loch Ryan, there is no indication of lateral moraine development. It is thought, therefore, that this asymmetrical distribution is most likely indicative of the availability of subglacial sediment at the time of moraine development.

2. Depositional ridges. The category 'depositional ridge' includes (a) small-relief ridges on the floor of the basin to the north-east of Stranraer; (b) transverse ridges to the east of Loch Ryan; and (c) large-relief ridges. The three types of depositional ridge which have been identified all show up distinctly in aerial photographs, and a good accuracy of the distribution of these is anticipated. However, there may be small-relief ridges in the forested area to the west of Black Loch and White Loch which have not been detected. The total distribution of the depositional ridges is shown in Figure 4.12.

a) The generic term 'small-relief ridge' has been used here to represent minor ridges (>5 m high) whose form is often asymmetric, with gentle proximal and steep distal slopes. In planform, the small-relief ridges are somewhat inconsistent with one another, displaying both regular and irregular form. The absence of stratigraphic information (no exposure) inhibits a definitive interpretation of these landforms, however the morphology and arrangement of the individual elements suggests one of the following landforms:

i. Push moraine: A push moraine is a small moraine ridge, usually less than 10 m in height, produced by minor (often annual) glacier advances (Worsley, 1974; Boulton and Eyles, 1979; Rabassa *et al.*, 1979; Rogerson and Batterson, 1982; Sharp, 1984). The term used here is adopted from Benn and Evans (1998) which excludes any indication of glaciotectonic processes, but instead implies formation by 'bulldozing'. Push moraines commonly have asymmetrical cross-profiles conforming to the pattern displayed by the small-relief ridges in the Loch Ryan Basin.
ii. Squeeze moraines: These features are minor moraine ridges formed by the squeezing of saturated sediment from beneath the ice margin (Price, 1970; Worsley, 1974). They are somewhat different in morphology from the mapped small-relief ridges in that they have steep or vertical sides and are rarely more than 1 m high. Furthermore, these features tend to degrade rapidly or become modified by ice push, so that intact squeeze moraines are believed to be rare in the geomorphological record (Benn and Evans, 1998).

iii. Small composite ridges: Such landforms are composed of multiple slices of up-thrust and contorted bedrock and/or unconsolidated sediment (Prest, 1983; Aber *et al.*, 1989). In planform, composite ridges comprise arcuate suites of subparallel ridges and intervening depressions reflecting imbricately stacked sheets of up-thrust material. The diagnostic features of this landform therefore fail to correspond, in terms of their morphological expression, with the small-relief ridges which have been identified.

iv. Eskers: The term 'esker' refers to elongate, sinuous ridges of glaciofluvial sand and gravel (Warren and Ashley, 1994). Their planform is extremely variable, taking the form of a single continuous ridge of uniform cross-sectional profile, single ridges of variable height and width, complex braided systems of ridges, or single low ridges linking numerous 'beads'; the latter of which would be most appropriate to select in light of the mapped distribution of small-scale ridges. The alignment of eskers is commonly subparallel to the direction of former glacier flow, reflecting former meltwater flow towards the ice margin. Visually linking up the individual mapped ridges does depict a subglacial or englacial channel system flowing toward the ice margin. However significantly, this interpolation also provides the impression of a curvilinear pattern consisting of subparallel ridges which conforms with the modification of proglacial material through oscillations of the ice margin.

In summary, the author is confident that the landforms mapped as 'small-relief ridges' match the criteria designated for push moraines, both in form and positioning. It is felt, however, that a small number of the ridges mapped in the northern extremity might be eskers as their form displays a more sinuous character and are more inclined to be interconnected with neighbouring ridges.

b) The generic term 'transverse ridge' has been used to identify minor ridges whose alignment is generally cross-valley, and so are transverse to the orientation of the more prominent small-relief ridges mapped within the Loch Ryan Basin. The location of the transverse ridges is within an isolated pocket on the lower terrain to the east of Loch Ryan. Consideration of their origin is uncertain with the following interpretations appearing plausible: (i) push moraines, though their alignment would imply the bulldozing of proglacial material by south-westerly flowing ice which seems unlikely; (ii) eskers aligned transverse to former glacier flow; or (iii) crevasse-fill ridges, with the ridges making intersecting 'waffle' or 'box' patterns reminiscent of crevasse patterns.

c) The generic term 'large-relief ridge' has been applied to landforms which exist as isolated mounds and maintain a morphology which is distinct, either in cross-sectional profile or scale, from the features mapped as small-relief ridges. Within the Loch Ryan Basin, two such landforms have been identified, both of which exhibit an asymmetric profile, with steep proximal and shallow distal slopes, and an irregular planform. It is uncertain how these landforms originated but it is tentatively suggested that they might represent proglacially thrusted blocks of sediment marking the former margin of ice within the Loch Ryan Basin. There is, however, no available stratigraphic evidence which might reveal diagnostic internal structures such as thrust faults which are commonly associated with this landform. No other viable explanation appears to account for their origin.

3. Drift lineations. The drift lineations are here subdivided into drumlins and large-scale flutings (megaflutings, cf. Clark, 1993; Section 3.2.1.1). Drumlins are well-defined individual streamlined landforms, while fluting is defined here as a 'furrowing' of the land surface, for the most part superimposed on other landforms. Comparison of the present map with the glacial lineation map of Charlesworth (1926, Figure 2.13) and Kerr (1982a, Figure 1.8) reveals the inclusion in the present map of megaflutings with an inferred ice-flow direction of 149°. These features are situated to the south-west of Stranraer and are superimposed on top of the lateral moraine in this area. The occurrence of the superimposed megaflutes is interpreted to reflect the nature of ice flow within the Loch Ryan Basin during a final-stage flow event; principally sustained wet-based ice flow which is believed to generate large lineations (Hätterstrand, 1997). Those lineations conforming to the criteria of the drumlin for the most part do not reflect the same alignment as the megaflutings, and are therefore thought to belong to earlier ice-flow sets (i.e. flow sets 17 and 18).

4. Kames and kame-and-kettle topography. The term 'kame-and-kettle topography' is used here to define a landform assemblage containing tracts of upstanding masses of glaciofluvial deposits (kames), and intervening hollows (kettles or kettle holes) representing areas of subsidence caused by the melting of buried ice (ice stagnation). The existence of kame and kettle topography is diagnostic of the reworking of large quantities of debris by supraglacial and englacial drainage systems during the final stages of glacier wastage (Cook, 1946a, b; Clayton, 1964). This assemblage dominates the present map in the region to the south and south-east of Stranraer.

5. Glaciotectonic depressions. This term is applied in the present map to indicate areas which are suspected of having experienced basin excavation by glaciotectonic thrusting of material beneath an ice margin. Commonly, this form of depression is closely linked to a discrete hill of similar size and shape, situated a short distance downglacier (hill-hole pair; Bluemle and

Clayton, 1984). However, the depressions within the Loch Ryan Basin fail to exhibit this relationship, which inevitably places speculation upon this interpretation. Nevertheless, glaciotectonic depressions without associated hills have been recorded in the past (e.g. Ruszczynska-Szenajch, 1976, 1978; Andriashek and Fenton, 1989). Such depressions within the present area are clearly represented as elongate water bodies (e.g. Black Loch and White Loch) which are parallel to the fluted surface, and hence inferred ice-flow direction.

6. Meltwater channels. Glacial meltwater channels are included in the map to present an indication of their distribution and the approximate ice surface slope direction. These areas are interpreted to represent zones where considerable meltwater drainage was taking place and appear, for the most part, to have been confined to lateral and pro-glacial areas, as shown by the distribution of arrows. The channels represented on the present map are assumed to reflect the last flow direction and any discrimination between cross-cutting channels has not been attempted. The accuracy of the distribution of meltwater channels varies over the mapped area. In areas beneath the drift limit, glacial meltwater channels are easily mapped and few errors are anticipated. The majority of channels in these areas are of medium and large size (>5-10 m deep) while in higher elevated areas, where ice-scoured bedrock is present at the surface, small channels are often obscured by the geological signal resulting in a possible under-representation of these channels.

4.3: Palaeo-Depositional Settings

4.3.1: Background

This section presents the field observations on the sedimentology and stratigraphy of specific sites within Carrick and Galloway, south-west Scotland (Figure 4.13). As discussed in Section 3.3, it is anticipated that the character and distribution of the glacial deposits within the investigation area will provide an intuitive insight into some of the following possible activities within the palaeo-depositional environment:

1. shifting patterns of active depositional environments;
2. deformation, reworking and resedimentation of sediment by active and/or stagnant ice;
3. redistribution of sediment by flowing meltwater, gravity, and perhaps aeolian processes;
4. possible reworking of sediment deposits during periglacial or paraglacial activity.

The format of this section has been largely controlled by the distinct absence of sediment exposures throughout the investigation area, which has supported the earlier impression of Dr. John Walden (*pers. comm.*) during an unrelated field survey of sediment exposures in the

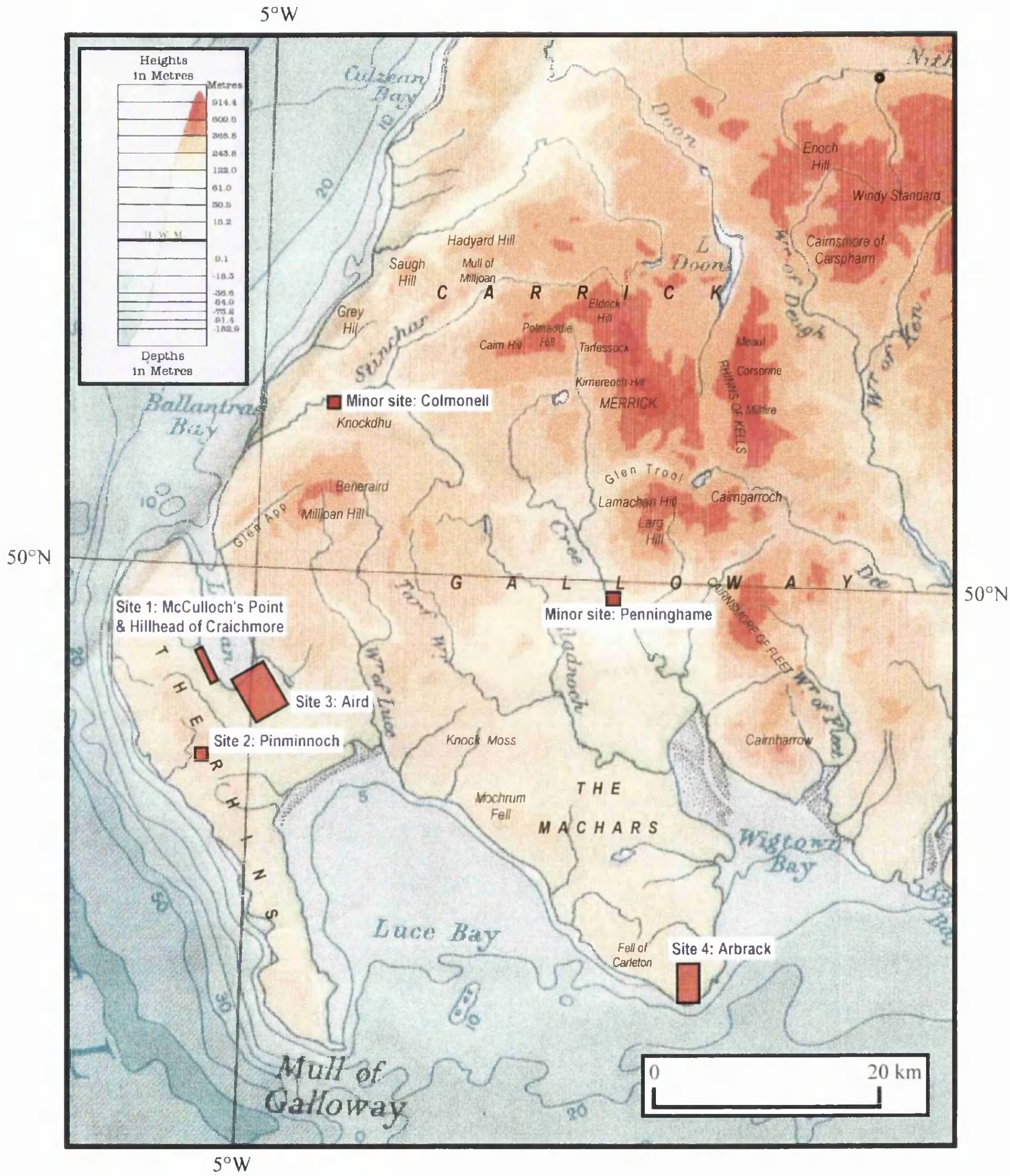


Figure 4.13 Location of field sites as presented in Section 4.3. (Adapted from the 1; 1 000 000 O.S. map of XVII Century England, 1930).

Galloway region. The described sites in Section 4.3.2 therefore reflect the only prominent exposures of glacigenic sediment which have been discovered, or could be accessed. A number of unrelated factors appear to explain the scarcity of stratigraphic sections in Carrick and Galloway: (1) lack of commercial extraction of sand and gravel; (2) absence of eroded coastal exposures; (3) natural plant regeneration or landscaping of road and railway cuttings; and (4) the overall absence of incision of the main drainage routes.

In Chapter 1, the area of south-west Scotland was divided into five sub-regions (Rhins of Galloway; Whithorn; Wigtownshire; south-west Ayrshire; and Carrick and Loch Doon) using the criteria set out in Section 1.3.1, so that a comprehensive description of the regional physiography (Section 1.3) could be presented for each region. Although such a format would also seem to be the most appropriate method of presenting a large quantity of similar field-related information in this section, the poor spread of sedimentary exposures throughout Carrick and Galloway has significantly curtailed the use of this approach. Instead, the investigation area is addressed here as a whole, but every effort has been made to present the field observations in the order that the sub-regions are addressed in Section 1.3. Radiocarbon dates are also included in this section.

The selected format of site description (Section 4.3.2) has been chosen because of the comprehensive and logical manner in which this ordering conveys information relating to field sites, thus this follows the sequence of sediment and stratigraphic description followed by interpretation. The author has included geomorphological information relevant to the interpretation of each field site description. In Chapter 5 the sedimentological and stratigraphic information for each of the sites is reconciled with the description of the geomorphology detailed in Section 1.3.

Throughout Section 4.3.2, a facies code is provided after the description of a particular stratigraphic unit in order to allow the reader to be able to identify this on related horizontal and vertical logs. A full description of each facies code is provided in Figure 3.15.

4.3.2: Sedimentology and stratigraphy

4.3.2.1: Site 1 - McCulloch's Point and Hillhead of Craichmore, Rhins of Galloway

Two >20 m long exposures through Quaternary sediments along the western shore of Loch Ryan both reveal a predominantly massive matrix-supported diamict (Figure 4.13, 4.14). The diamictons at both sites have been internally thrust faulted and disrupted by a network of Riedel shears picked out by wisps of sand (consistent cross-cutting faults) thereby displaying a)

b)

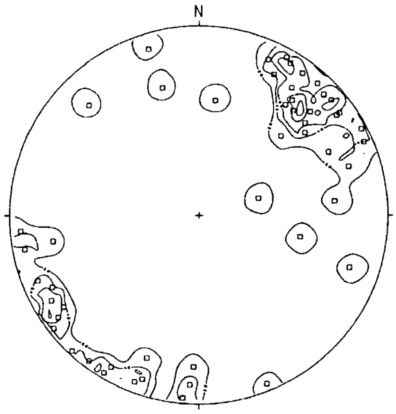
Figure 4.14 Massive matrix-supported diamict exposed at McCulloch's Point (a) and Hillhead of Craichmore (b), on the western shore of Loch Ryan.

evidence of glacitectonic disturbance (brittle deformation). The exposure at McCulloch's Point [exact position - NX 048624] and that at Hillhead of Craichmore to the north-north-west [exact position - NX 038634], are separated by approximately 1500 m of heavily vegetated banking above the upper shoreline. Both sites reveal sections of fairly consistent thickness varying between 2.6 m and 3.9 m, with this depth reflecting a vertical distance from the upper shoreline to the land surface. At both sites the sequence of massive diamict is capped, for the most part, by a thin veneer of colluvium (< 0.5 m). Aside from the occurrence of shear planes within the deposit, both exposures contained no other obvious sedimentary structures. Within the massive diamict at both exposures is a heterogeneous and variably-sized clast content within an extremely compact fine-grained matrix of silts and clay.

Two clast macrofabrics, comprising 50 clasts per fabric from the upper and lower portion of the diamicton (*c.* 2 m apart), were collected from McCulloch's Point and also at Hillhead of Craichmore (Figure 4.15) to compare the fabric signatures with tills and glacitectonites (e.g. Benn, 1994, 1995; Benn and Evans, 1996; Evans *et al.*, 1999). The envelopes on the fabric shape triangles in Figure 4.16 have been determined from:

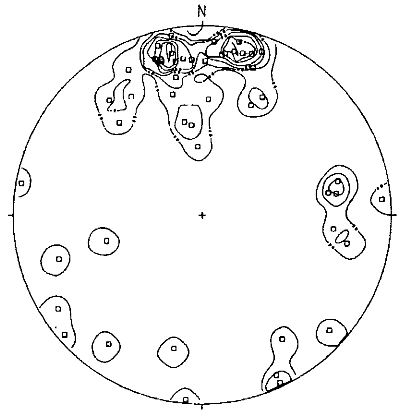
i. Fabric shapes for deformation till and glacitectonites from Drumbeg Quarry near Loch Lomond in central Scotland (Benn and Evans (1996).
ii. Fabric shapes for deformation till from Breidamerkurjökull, Iceland (Benn, 1995). Within a two-tiered till structure discovered in this area, the upper (A) horizon has been shown by Boulton and Hindmarsh (1987) to be undergoing ductile deformation, while the lower (B) horizon is deforming by discontinuous brittle or brittle-ductile shear.
iii. Fabric shapes for a hybrid lodgement and deformation till from Slettmarkbreen, Norway (Benn, 1994).

An important detail stressed by Evans *et al.* (1999) is that in many subglacial sediments a continuum of fabric strengths must exist which reflects the maturity of the subglacial sediment (Figure 4.16), or more specifically its strain history (e.g. Evans *et al.*, 1998). Therefore, clasts which have been subjected to a prolonged period of intense strain within a thin layer undergoing failure by brittle shear are likely to possess a strong fabric cluster (Evans *et al.*, 1999). Conversely, the shearing and alignment of smaller particles requires only shorter periods of strain producing strong microfabrics. These characteristics consequently have to be taken into consideration when using a macrofabric alone to assess palaeo-ice flow directions.



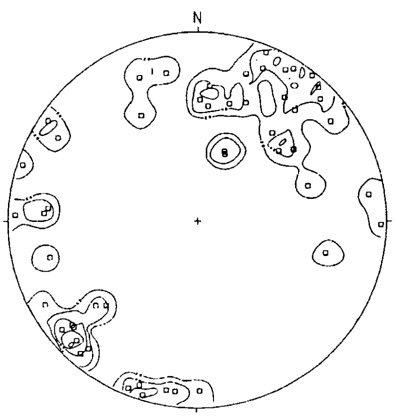
McCulloch's Point (lower diamict)

MLA	043.8
S_1	0.731
S_2	0.181
S_3	0.088



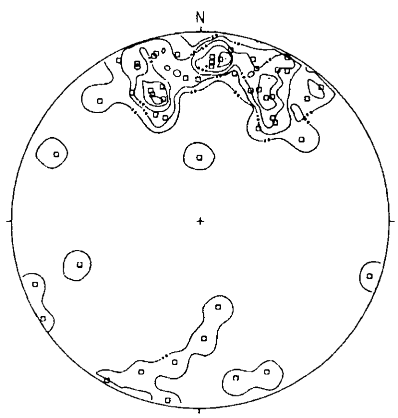
McCulloch's Point (upper diamict)

MLA	358.5
S_1	0.656
S_2	0.242
S_3	0.102



Hillhead of Craichmore (lower diamict)

MLA	036.8
S_1	0.642
S_2	0.250
S_3	0.108



Hillhead of Craichmore (upper diamict)

MLA	007.2
S_1	0.711
S_2	0.186
S_3	0.103

Figure 4.15 Schmidt stereonet plots of the McCulloch's Point and Hillhead of Craichmore clast fabrics (upper and lower diamict) produced by Rockware software. MLA, mean lineation azimuth (in degrees); S_1–S_3, eigenvalues.

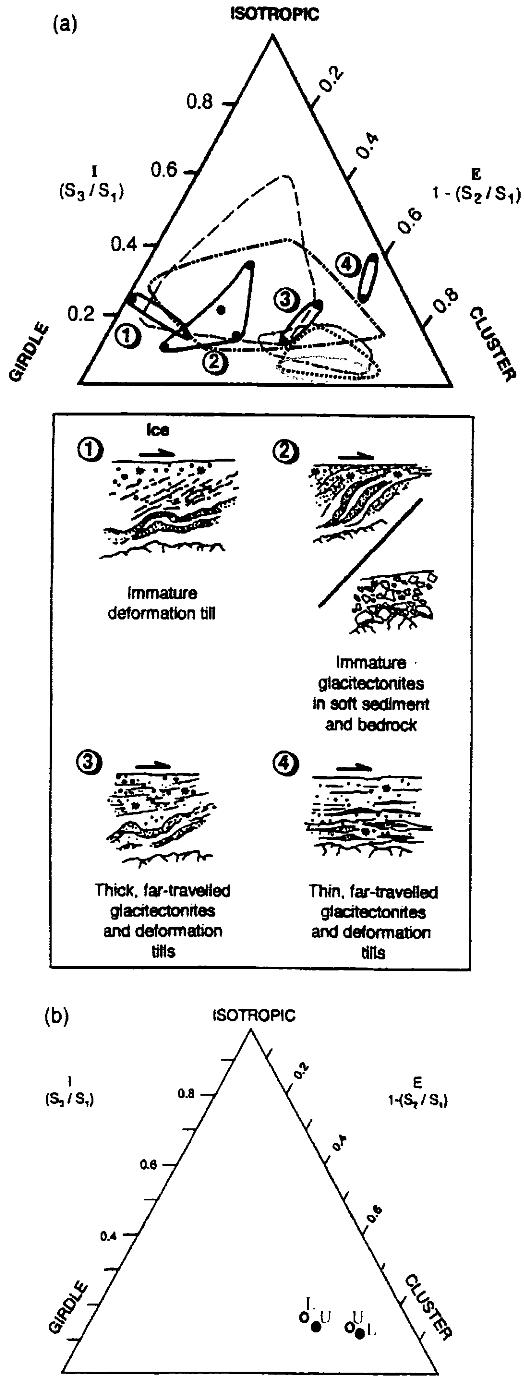


Figure 4.16 (a) Fabric-shape triangle (after Benn, 1994) containing envelopes of previously reported subglacially deformed materials (sample points represent fabric samples). Broken line: deformation till envelope (Benn and Evans, 1996); dot-dash line: glacitectonite envelope (Benn and Evans, 1996); 2 dot-dash line: upper till envelope, Breidamerkurjöjull (Benn, 1995); dotted line: lower till envelope, Breidamerkurjökull; solid grey line: hybrid lodgement and deformation till envelope (Benn, 1994). Solid line envelopes represent a range of deformation till–glacitectonite fabric shapes arranged according to maturity / travel distance (Evans *et al.*, 1998). (b) Fabric-shape triangle containing the McCulloch's Point (filled circles) and Hillhead of Craichmore (unfilled circles) clast fabric samples. U, fabric sampled from upper diamicton; L, fabric sampled from lower diamicton. (Modified from Evans *et al.*, 1999).

The position of the four fabrics in the fabric triangle, within the glacitectonite envelope, confirm the initial conjecture that the massive diamict has been subjected to former glacitectonic disturbance (Figure 4.16). The glacitectonite appears to be far-travelled based upon the fabric shape envelopes in Figure 4.16a and comparison with data from Evans *et al.* (1998). The striking contrast between the fabrics taken from the upper and lower portions of the two exposures is the difference in the fabric signatures (Figure 4.15). The dip direction remains fairly similar in the upper and lower diamicton samples, however, the Schmidt nets for both sites display a shift from the north-east to north quadrants upsection (mean lineation azimuth of 36.8°-43.8° for the lower diamicton compared with 358.5°-7.2° for the upper). This appears to strongly indicate that there has been a change in the palaeo-ice flow direction along the Loch Ryan Basin from a transverse flow to one corresponding closely with the general trend of the axis of the basin. The stratigraphy at both exposures does not provide any evidence to suggest a multiple till sequence. Therefore although the general change in the fabric signature upsection might correspond to a readjustment of a single phase ice flow, it is thought that the fabric of the upper diamicton most likely records a superimposed strain signal resulting from glacitectonic modification by a later flow stage.

4.3.2.2: Site 2 - Pinminnoch, Rhins of Galloway

Extraction of sand and gravel from a site at Pinminnoch [exact position - NX 026546] has created two exposures through Quaternary sediment which are offset from one another, thereby providing an insight into the three-dimensional stratigraphy of the depositional setting (Figure 4.13, 4.17a, b). The pit is developed in the south-west extension of a 600 m long and 360 m wide elongate drift mound (orientated 034° - 214°). The site is located within an area of subdued and fairly regularly distributed drift ridges among occasional flat-bottomed depressions, with several palaeo-channels on the surface additionally contributing to the undulating topography (Figure 4.18). The main south-westerly facing exposure is approximately 50 m long. However, the stratigraphy has been concealed along a large portion of the section due to the deposition of recently extracted material up against the old exposure.

An undisturbed section situated on the western extremity of the main face reveals, for the most part, a complex cyclical sequence of horizontally-bedded and laminated medium- to coarse grained sands (Sh, Sl) interspersed with well-sorted to moderately-sorted granules and gravels (Gms, Gh, Guf, GRh) containing occasional lenses of coarse-grained massive sands (Sm) (Figure 4.17a). Beneath this 2 m sequence lies a 15 cm thick layer of brown-purple massive clay separating the above cyclical sequence from an underlying unordered sequence of medium-grained massive sands (Sm), horizontally bedded coarse-grained sands (Sh), and

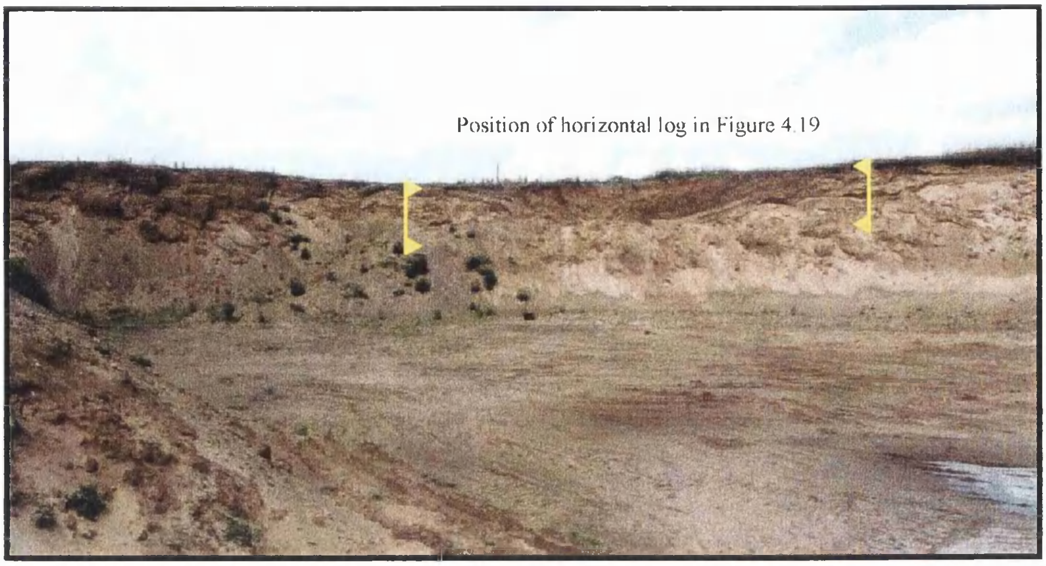


Figure 4.17a The main south-westerly facing exposure at Pinminnoch.

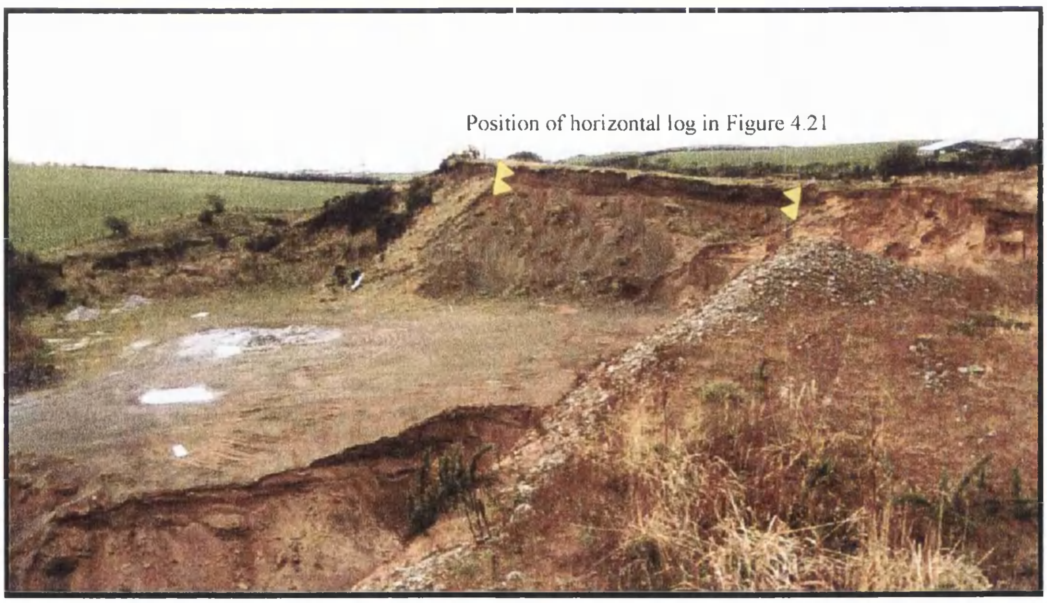


Figure 4.17b The adjoining south-easterly facing exposure at Pinminnoch.

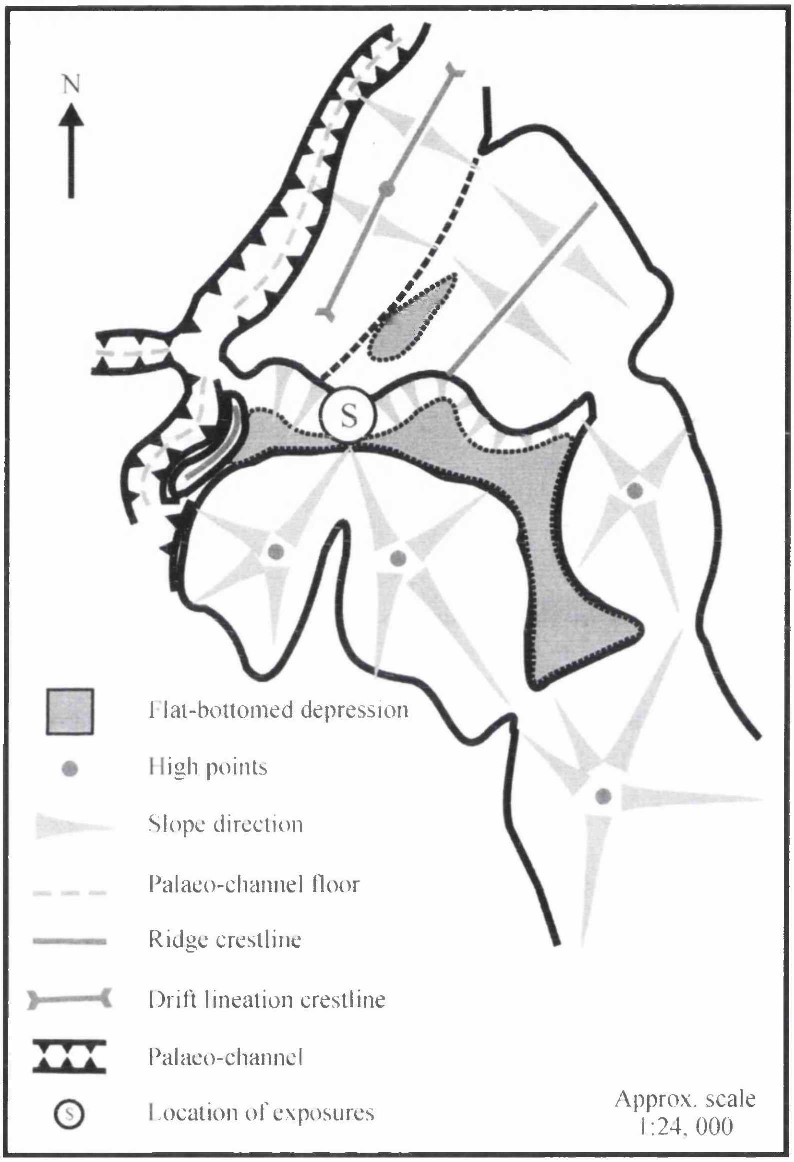


Figure 4.18 Pinminnoch site location map, indicating the positions of the major morphological features.

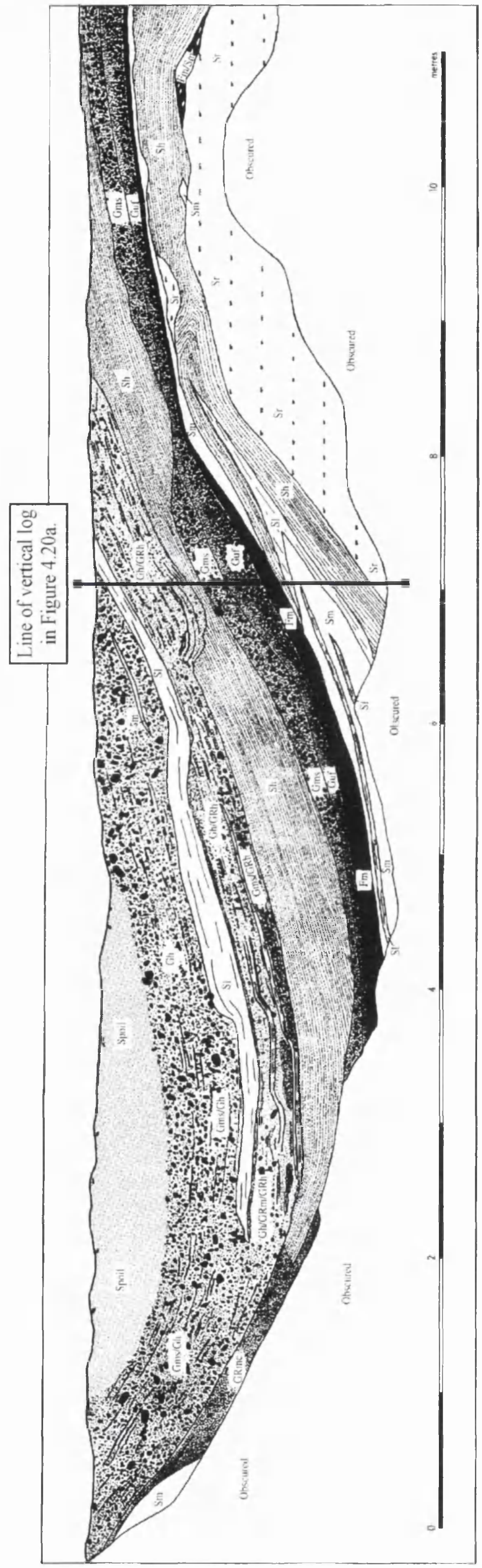


Figure 4.19 Horizontal log of the south-westerly facing exposure at Pinminnoch. See Figure 3.15 for explanation of facies code.

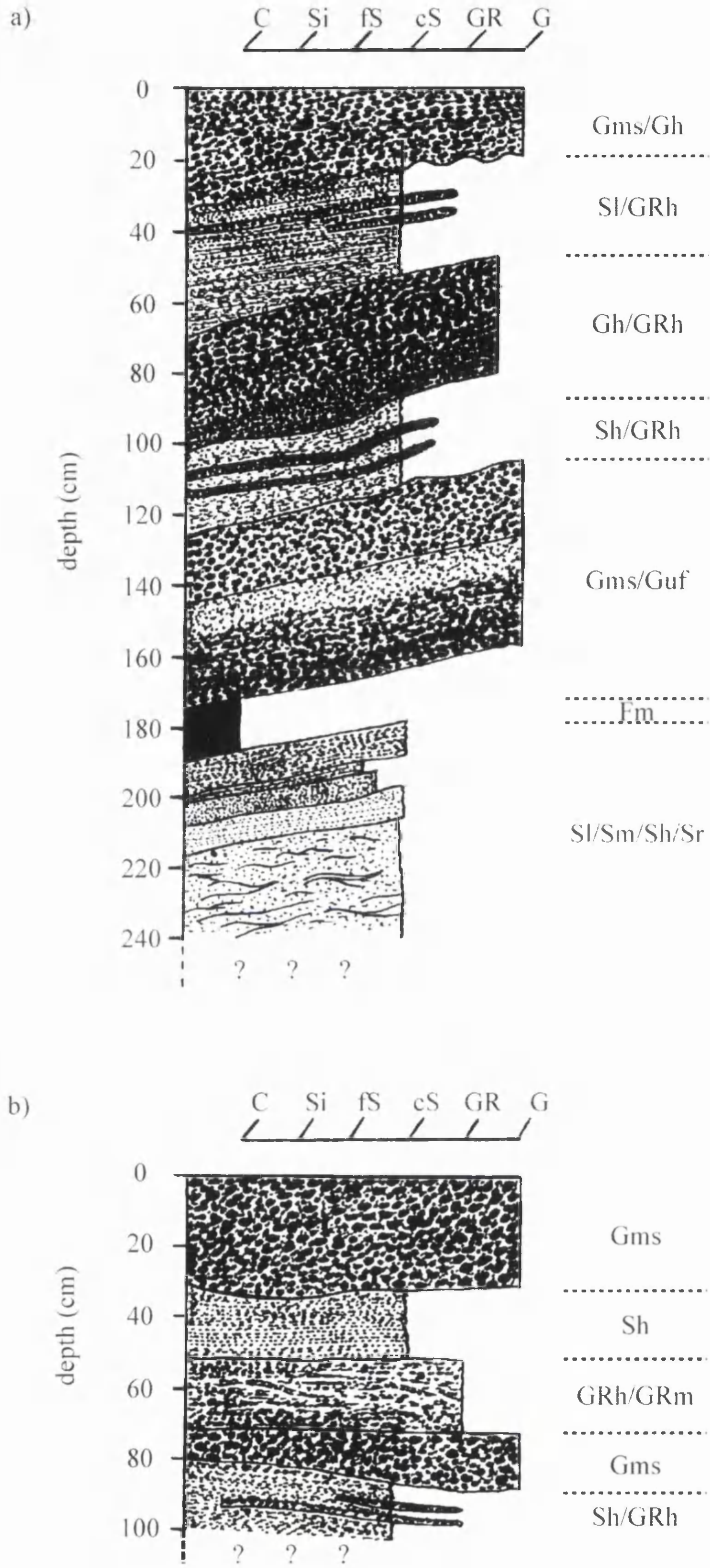


Figure 4.20 Vertical profile log of the south-westerly facing exposure (a) and adjoining south-easterly facing exposure (b) at Pinminnoch (positioning of logs indicated in Figure 4.18 and 4.19 respectively). See Figures 3.15 and 3.16 for explanation of facies code and symbols. C, clay; Si, silt; fS, fine sand; cS, coarse sand; GR, granules; G, gravel.

laminated fine- to medium-grained sands (Sl). The rest of the underlying exposure reveals predominantly cyclic medium-grained sands with ripple forms (Sr) (Figure 4.19, 4.20a). The adjoining south-easterly facing exposure (approximately 25 m long), which would relate to the top third or so of the main south-westerly facing exposure, depicts a more unordered, chaotic sequence (Figure 4.17b). A predominantly matrix-supported gravel (Gms) is revealed with intercalating thin sheets of horizontally-bedded coarse sands (Sh) and laminated sands (Sl), lenses of massive sands and gravels (Sm, Gm), and pockets of upward-fining gravels (Guf) which are sometimes repeated (Gruf) (Figure 4.20b, 4.21).

In summary there are seven major facies which have been identified from the two exposures at Pinminnoch:

1. Poorly-sorted, subrounded gravels, occasionally weakly stratified, with a modal grain-size of approximately 30 cm supported in a matrix of medium-grained sand (rarely clast-supported). Sand interbeds in places.
2. Cyclic rippled medium-grained sands greater than 1 m.

These two facies quantitatively dominate both exposures although facies 2 is only exposed at the base of the south-west facing section. Intercalated within the matrix-supported gravels of both exposures are:

3. Horizontally stratified, coarse- and medium-grained sand units up to 25 cm thick.
4. Fining upwards sequences of gravel up to 20 cm thick.
5. Parallel laminated medium-grained sand units up to 25 cm thick.
6. Thin sequences of massive and horizontally-bedded granules (usually <10 cm thick)
7. Massive clays between 5 and 15 cm thick.

The facies can be differentiated into two lithofacies associations: (LFA 1) an upper, coarse-grained lithofacies association comprising gravels and stratified granules, and (LFA 2) a lower, fine-grained lithofacies association comprising clay, and stratified and rippled sands. LFA 1 in the main south-westerly facing exposure has an average thickness of around 170 cm and a lower surface which appears conformable with underlying LFA 2 whose upper surface is represented by the massive clay facies. Minor folds are occasionally found in the coarse lithofacies association, although no faulting has been observed. The extensive fine lithofacies association represents a subaqueous environment of standing or low flow regime sedimentation, and is closely comparable with sediments described by Donnelly and Harris (1989) from a small ice-dammed lake in Norway. Ubiquitous ripple sequences indicate deposition from suspension during waning flows (Paterson and Cheel, 1997). The absence of dropstones indicates shallow water. The coarse lithofacies association has the characteristics of

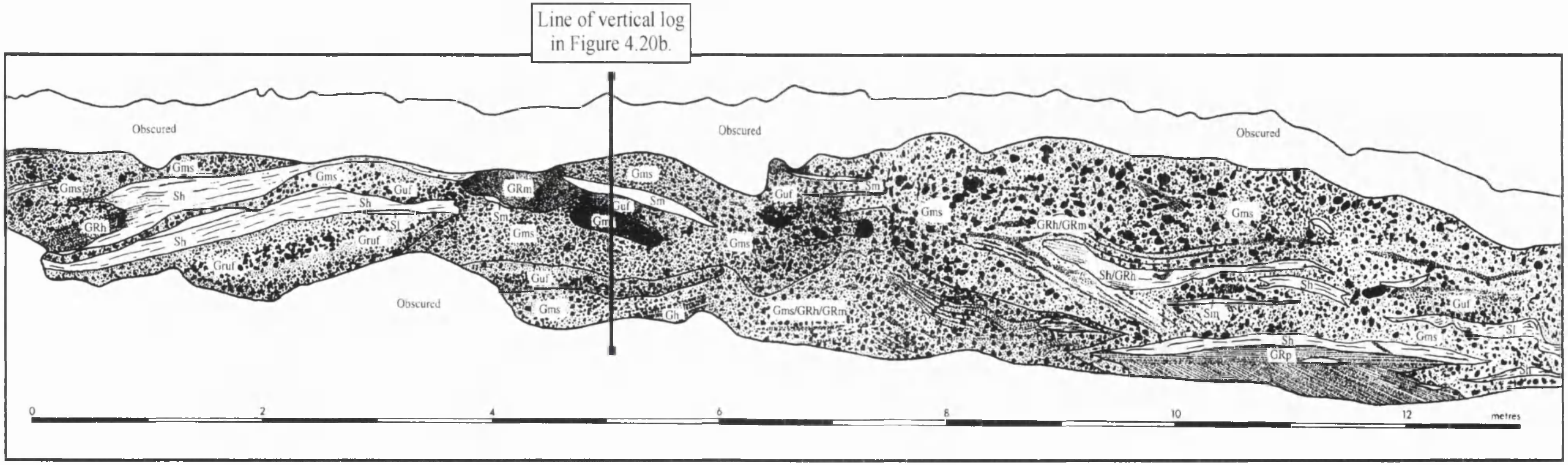


Figure 4.21 Horizontal log of the south-easterly facing exposure at Pinminnoch. See Figure 3.15 for explanation of facies code.

fluvial channel deposits from high sediment concentration flows (e.g. Saunderson, 1977; Cheel and Rust, 1982; Maizels, 1989, 1991) and the gravel units suggest proximity to a sediment source (Benn, 1996). Pebble gravel deposited by the strongest flows passes up through stratified granules into stratified sands. These successions most probably represent pulsed discharge events showing an alternation between high and low sediment input and between periods of high and low flow regimes (variations in meltwater discharge). Measurements of suspended sediment concentrations and total loads in glacial outwash streams in the Alps, Scandinavia, the cordillera of North America, and the Karakoram Mountains (e.g. Collins, 1990; Kelly, 1990; Kruszynski, 1993) have revealed that sediment pulses can be associated with, or independent of, discharge changes. These have been demonstrated to be due to (1) hydrological conditions (e.g. disruption of drainage in surging glaciers has been observed to produce both discharge and sediment pulses); (2) climatological conditions (e.g. floods with extreme sediment loads induced by storm events); and (3) glaciological conditions (e.g. mobilisation of sediment close to the glacier terminus produces sediment pulses) (Johnson, 1997).

The sharp contrast between lower LFA 2 and upper LFA 1 could infer the incision of a fluvial system into the underlying fine lithofacies association, however, the absence of an obvious erosional contact nullifies this hypothesis. It is therefore believed that LFA 2 represents rapid deposition within a ice-distal, low-energy, aqueous environment (in an area opened up by ice retreat in which ponded water existed) which slowly drained leaving a succession of sands capped with a clay carapace. A flat plain would have been left after the shallow water body drained. Evidence of this is found in the area to the south of the pit which exhibits a flat-bottomed depression surrounded on all sides by mounds and ridges of higher relief, except for two narrow routes to the west, which enter into a palaeo-channel system (Figure 4.18). Following the gradual draining of the water body, over a period of indeterminate time, it is inferred from LFA 1 that there was a triggering of pulsed alluvial deposition causing subsequent deposition of gravel units over the fine lithofacies assemblage. In summary. therefore, the depositional history of the Pinminnoch site can be divided into two phases: (1) deposition of the subaqueous deposits; and (2) deposition of the glaciofluvial sediment. The spatial relationship between the two lithofacies associations is thought to reflect one of the following scenarios:

1. Renewed proximity to an ice margin and sediment source, inferring a readvance of ice, whose margin became more proximal to the site at Pinminnoch (Benn, 1996)

2. The release of a high, pulsed discharge event from a relatively stabilised ice margin depositing a high sediment concentration, subsequent to a quiescent phase in sediment discharge.

4.3.2.3: Site 3 - Aird, Stranraer

A working sand and gravel quarry at Aird [exact position - NX 098602], three and a half kilometres east-south-east of Stranraer reveals a clear exposure through predominantly graded, plane-parallel stratified units comprising very coarse gravel and intercalated medium-coarse grained sands (Figure 4.13, 4.22). This component of the exposure rests on top of what are undoubtedly clinoforms composed of a variety of sand and gravel facies with varying modal grain sizes. The nature of the quarry is such that the extraction of material takes place on essentially two main tiers: (1) an upper tier which reveals the horizontally-parallel units; and (2) a middle tier exposing the clinoform stratigraphy; with the base of the upper tier possibly representing the approximate position of the boundary between the two lithofacies, though this is speculative and cannot been confirmed (unexposed). The base of the quarry (lower tier) has been back-graded in most places and therefore exposed stratigraphic sections are absent.

Stereoscopic examination of the quarry site using a pair of 1:24 000 aerial photographs, together with field reconnaissance, indicates that this depocentre forms the entire vertical extent of the south-easterly embankment of a generally south-east to north-west aligned, flat-bottomed, wetland basin (5-10 m above sea-level), whose north-easterly margin opens out through a fairly narrow breach onto the seaboard at the head of Loch Ryan, near Stranraer (Figure 4.22). This steep rise, into which Aird quarry is cut, continues around the lateral margin of the basin on both the north-east and south-west sides before eventually becoming less pronounced in height, and more integrated with the surrounding relief. To the west and south of Aird the land surface, corresponding with the upper most level of the quarry, is characterised by both kame-and kettle topography and intermittent meltwater channels as described in Section 4.2.4.3, while to the south-east and east, the flat upper level of the quarry exposure grades imperceptibly into gently undulating terrain of low-relief.

The characteristics of the deposits broadly outlined above are thought to closely correspond with the upper and middle components of a Gilbert-type delta complex: upper horizontal topsets (top tier); and inclined foresets (middle tier) (e.g. Nemec, 1990; Prior and Bornhold, 1988, 1990). The third component normally associated with Gilbert-type deltas, bottomsets, are not revealed in Aird quarry, though may possibly exist below the level of current sand and gravel extraction. Despite the prominence of a plateau capping the delta at 27 m above sea-

Figure 4.22 Aird site locational map, indicating the positions of the major geomorphological features within the vicinity of the quarry.

level, there is no evidence to suggest the termination of the plateau in a steep ice-contact slope. It is therefore strongly argued that this delta was connected to a subaerial feeder river receiving sediment from glacial meltwater streams which have subsequently been infilled, and therefore cannot be detected on aerial photographs. The two main sedimentary successions which are described below are recorded from a northerly facing stratigraphic section on the upper tier (topset units), and a north-easterly section on the middle tier (foreset units). Implications of these deposits are discussed where appropriate.

The topset units (Figure 4.23), representing the most elevated part of the delta, include the following main facies within a 5.5 m exposed section: (1) massive to weakly stratified, medium-bedded coarse and very-coarse grained, matrix-supported gravels represented as extensive tabular sheets (Gms, Gh, Gm); (2) lenses of poorly sorted sand and gravel (Sm, Gm); (3) extensive tabular sheets of horizontally bedded and massive sand (Sh, Sm); and (4) laterally discontinuous units of granules, with oversized clasts interbedded with wisps of coarse sand (GRh. GRmc, GRch). No molluscan fauna were discovered among the topset component, despite an extensive search. Of significance, with respect to a preliminary interpretation of this stratigraphic sequence, has been the discovery of a number of the units which commonly display shallow erosional (channelised) bases, and the observation that nearly all of the facies within the sequence generally lack any noticeable degree of sorting.

It is apparent that a number of characteristics described above are strongly indicative of fluvial deposition of these sediments. Gravel sheets, in particular, are common on low-relief parts of river beds, such as longitudinal and bank-attached bars where they develop through clast-by-clast accretion (Hein and Walker, 1977; Miall, 1977; Collinson, 1986). In addition, coarse-grained sands with horizontal bedding tend to arise where sedimentation has prevailed under a low flow regime. Several lines of evidence therefore suggest that the architectural element of the topsets strongly reflects the distal zone of a fluvial system, and perhaps characterise a sandur environment. Low-relief gravel surfaces form a major element of many sandar (Miall, 1992). Among the exposed topsets, occasional slump structures within the uppermost facies are suspected to represent the position of former lodged blocks of ice which have been incorporated into the fluvial system and laid to rest among the sandur deposits, from which position they have melted out. Overall, these reasonably complex units are recognised as topsets because of their stratigraphic position and the fact that they erosionally truncate and overlie the steeply dipping foreset strata. It is therefore inferred that this lithofacies association preserves a record of the complex interface between a sub-aerial feeder channel and a

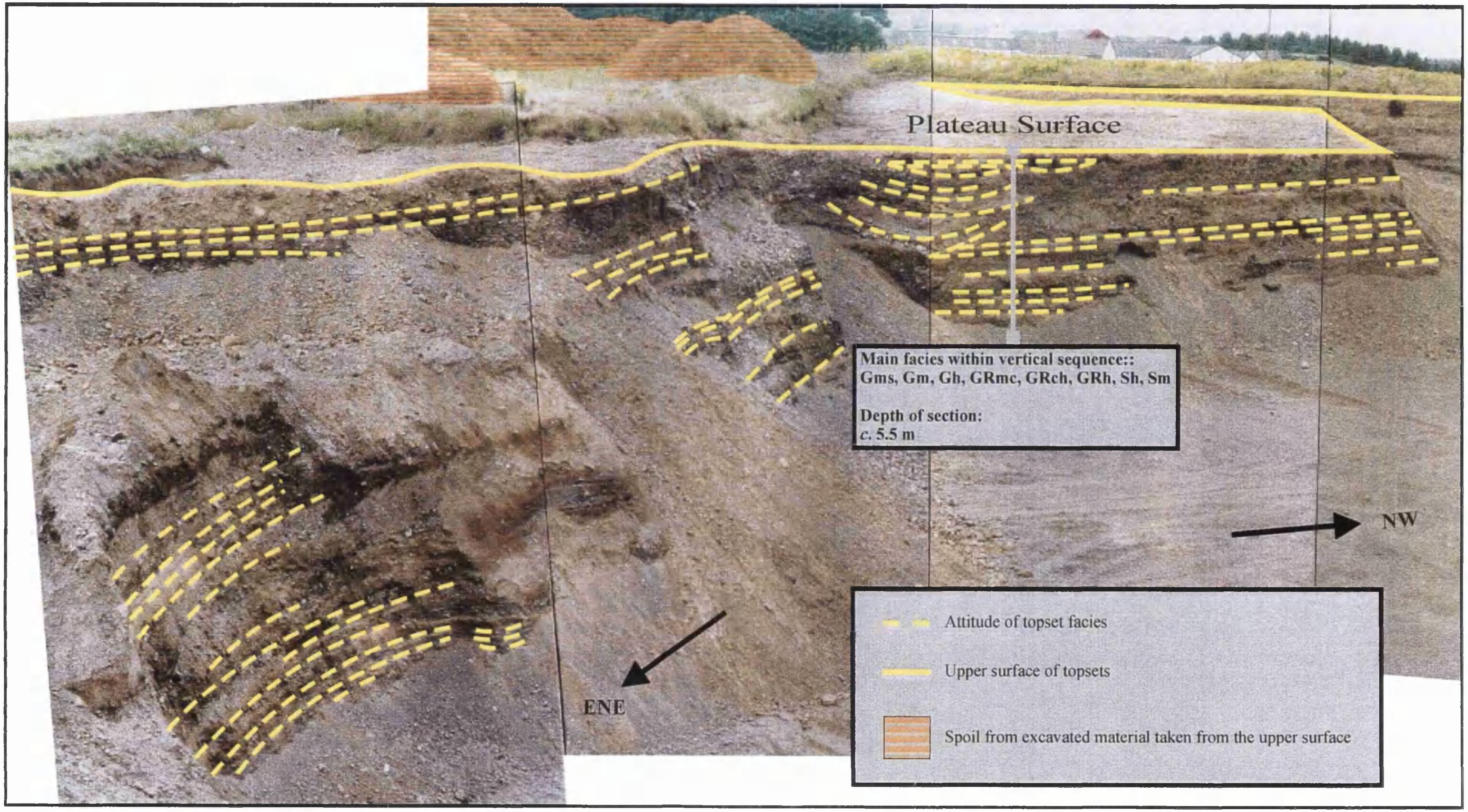


Figure 4.23 Schematic diagram of exposed topsets above the upper tier of Aird quarry.

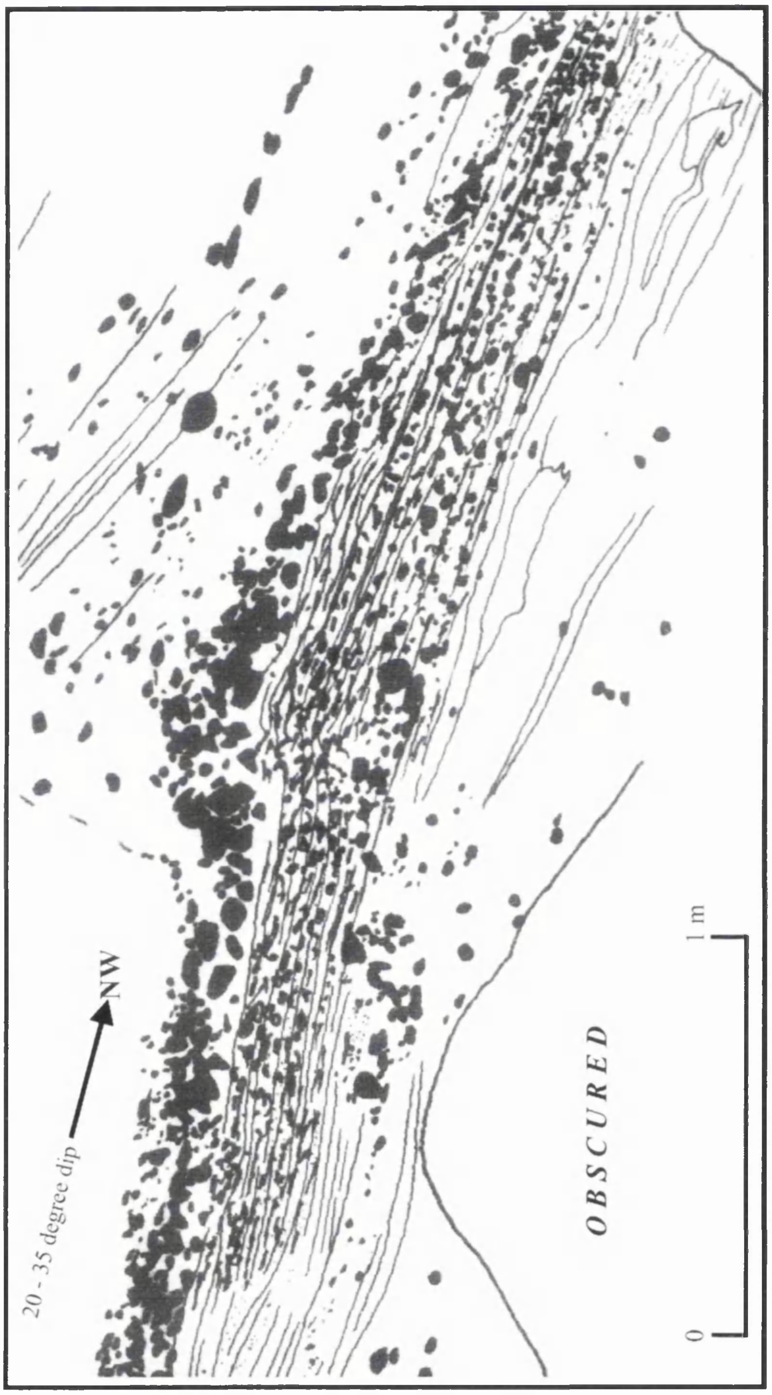


Figure 4.24 Schematic diagram of part of the exposed foresets at Aird quarry.

subaqueous depositional environment whose context with respect to this depositional setting is discussed later.

The underlying foreset units (Figure 4.24) exposed on the middle tier of the quarry (upper level thought to be *c.* 15-20 m above sea-level) provide the distinctive facies geometry and stratigraphy that allows the recognition of a Gilbert-type fan delta. Where it is prominently exposed, the foreset sequence is up to 9 m in thickness though it is suspected it may be thicker than this. Overall, the section is characterised by steeply dipping units, with dips towards the north-west (305°), and with dip angles from 20° to 35°, although they are observed to become more shallow towards the north-west (15°-20°). The foreset units typically consist, for the most part, of (1) massive open-work gravels (Go, Gm), (2) single and repeated upward-fining coarse-grained gravels (Guf, Gruf); (3) graded gravels and sands (Gh, Sh, Sl); and (4) clast-supported boulders (BGfo, BGcg). Other small exposed sections (*c.* 1-2 m^2) have allowed the gravel foresets to be traced from the main section detailed above, on the middle tier, for approximately seventy metres to the north-west (down-dip), which clearly suggests a lateral hydraulic sorting of grain-size, with much finer, compact foreset units (matrix-supported gravels, massive gravels with occasional fining upwards, and interbedded coarse-grained sands) becoming more commonplace in what would equate with the lower delta foreslope.

Lowe (1982), Postma and Roep (1985) and Nemec (1990) suggest that such evidence invokes the 'freezing' of traction carpets as the energy of flow decreases, resulting in the rapid deposition of coarse gravelly units, while the more turbulent and fluid sandy materials continue travelling down the delta where they become deposited further downslope, or eventually as bottomsets. Alternative explanations are that the proximal to distal fining indicates turbulent sediment-gravity flows or turbidity currents beneath the standing water. As the flow slows down or becomes less turbulent, these processes give rise to the lateral sorting of beds, reflecting the rapid deposition of coarse material and the transport of finer material into deeper parts of the water body (Nemec, 1990).

Beyond the lower tier of Aird quarry to the north-west, an exploratory investigation has been carried out on the flat-bottomed basin to verify the composition of the sediment in order to (a) ascertain the depositional setting of the delta as either glaciomarine or glaciolacustrine, and (b) to provide further evidence to invoke the energy regime of this environment. As no stratigraphic sequences are revealed within the basin, manual coring was undertaken whereby 0.8 m cores were taken at a series of depths (1 m intervals) using a Russian corer and 1 m extension rods in order to provide a complete stratigraphic profile beneath the surface. The length of the entire sequence was determined by the practical limitation imposed by the nature of the coring technique, and the water-saturated conditions of the site. Sampling was conducted along a transect at regular twenty metre intervals from the south-east margin of the basin (*c.* 150-200 m from the exposed foreset beds) towards the centre, until a position was reached where underfoot conditions became too water-saturated to allow satisfactory retrieval of the corer (100 m from the initial point) (Figure 4.22). The stratigraphic signature revealed by five cores (4-5 m in length) all conform to a basic sequence of poorly-developed peat and plant detritus down to a depth of 120-160 cm, followed by an abrupt transition to predominantly massive black-grey clays (referred to as lithofacies 1), followed by a gradual gradation at around 225 cm to dark-grey clays and intermittent light-coloured silts (referred to as lithofacies 2), with very occasional organic matter. The only variability between the cores was in the thickness of the upper peat horizon which became gradually greater as the sampling progressed away from the basin margin. Within one of the cores, the author was fortunate enough to sample a complete mollusc within lithofacies 1 at a depth of 210 cm, which corresponds with an altitude of approximately 6 m above present sea-level (Figure 4.25). The shell yielded an AMS age of 6 075 ± 55 years B.P. (AA-33830), and is taken to infer a minimum deglacial date for this area if it is *in situ*.

From the observations provided by the stratigraphic profiles the following points are introduced which are thought to be significant in characterising the depositional environment at Aird:

1. The predominant fine-grained nature of lithofacies 1 and 2 beneath the peat accumulation invokes the settling of suspended sediment within a low-energy subaqueous environment, for example a glacial lake, lagoon, or sheltered marine embayment.
2. The lateral extent of lithofacies 1, within the basin, might suggest that sediment was carried into the water body as high-level overflows and interflows, from which it eventually settled out through the water column. Alternatively, if sea-level was approximately 5-10 m above present, the basin would be inundated by shallow marine waters which would promote the deposition of tidal sediments within the embayment, producing a mud flat (lithofacies 1?). A date of 6 075 ± 55 years B.P. for the mollusc found at an altitude of *c.* 6 m above sea-level provides a minimum sea-level height for this time (on the premise that the shell was found in its growth position), and closely supports the height of sea-level inferred by Lambeck's sea-level curve at 6 000 years B.P. for the west Solway Firth (Wigtown Bay and Luce Bay) (Lambeck, 1993b; Figure 2.29). Thus, such a scenario fits with the available evidence and accounts for the widespread extent of lithofacies 1.

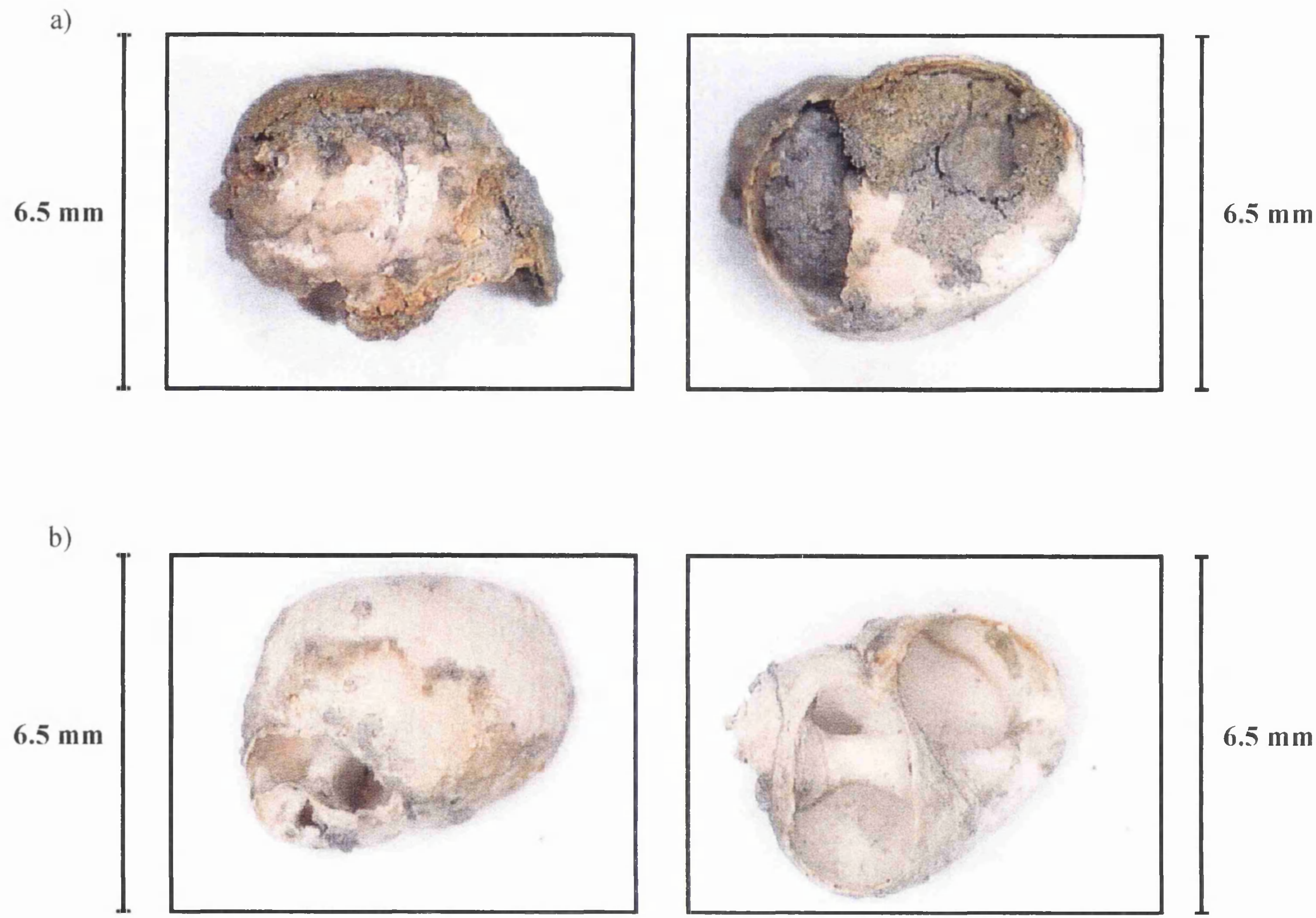


Figure 4.25 Mollusc sampled in massive clays from a position 210 cm below the surface of the basin, north-west of Aird quarry. The shell yielded a date of 6 075 ± 55 years B.P. (a) sample before removal of sediment; (b) sample after removal of sediment.

3. Massive clays are known to commonly develop below shallow water owing to sediment mixing by wave disturbance, wind generated water currents and bioturbation.

The evidence presented from this site does not unequivocally classify the type of aqueous environment into which the foresets and basin sediments were deposited. Points 1 and 3 above do, however, strongly invoke a shallow water, low-energy environment, leading to two possible types of depositional environment: (1) a low-energy, shallow lagoon (embayment) connected to open marine waters within Loch Ryan Basin; or (2) a lacustrine setting. Upon assessing the depositional setting of this depocentre, it is important to consider that the formation of the Gilbert-type delta occurred presumably during a period when sub-aerial meltwater and sediment was in great abundance, accounting for the vast input of sediment into this delta system. It is logical to suggest, therefore, that this would most likely coincide with the Lateglacial stagnation and retreat of ice from the Loch Ryan Basin.

At such a time, Lambeck's (1996) predicted deglacial palaeoshorelines infer that sea-level in south-west Scotland was slightly lower than at present (Figure 2.30), and so the basin into which the clays and silts were deposited would undoubtedly have been filled with freshwater. It should be highlighted at this point, however, that the accuracy of numerical modelling from which these predictions have evolved has been strongly disputed by McCabe (1997), as is discussed in Section 2.7. Nevertheless, if the predictions of Lambeck's (1996) model are accepted for the purpose of discussion, the characteristics of lithofacies 2 within the stratigraphic cores most likely reflect fluctuations in grain size and the quantity of incoming sediment from the delta, as a result of annual water and sediment discharge cycles into a lacustrine setting. The repeated cycle of clay and silt is not, however, thought to be consistent enough to classify these deposits as varves, owing to the considerable quantity of clay which, on occasions, has accumulated between consecutive silt layers (up to 30 cm). Point 3 above could quite easily explain the absence of typical closely repeated silt-clay sequences. To allow the development of a lake within the basin it is necessary to dam the opening to the north-west, in order to allow the accumulation of approximately 15 m to 20 m of water (to the top of the foreset beds). The most obvious scenario which can be envisaged is the positioning of retreating ice across the Loch Ryan Basin so that it abuts the narrow opening, thereby 'sealing' the lake (Glacial Lake Aird). Eventually when the ice is removed, however, it is anticipated that Glacial Lake Aird drained to the north-west through the breach, exposing lithofacies 2. Eventually lithofacies 2 was submerged by marine waters during the Holocene when sea-level rose to a over 6 m above the present level, thereby filling the embayment with shallow water in which lithofacies 1 was deposited.

One important point that remains to be addressed concerns the positioning of the source which was responsible for feeding meltwater and sediment to the north-west, as recorded by the direction of the foreset beds. It has been invoked, for reasons given above, that the sediment entered into the subaqueous environment by means of a subaerial feeder stream. Clearly, the most obvious scenario which can explain the prograding of the delta along this orientation is by meltwater flowing to the north-west from an ice marginal position to the south-east. However, ice flow to the north-west has not been modelled by previous investigations, and is not represented in the most recent reconstruction of the palaeo-ice sheet dynamics of this area (Section 4.2.3). A conceivable explanation, however, might arise from the positioning of a piedmont lobe within the lowland plain, to the south of Stranraer, during the time of delta development. This would require ice, previously confined within the bedrock trough of the Loch Ryan Basin, to sprawl out over the plain in such a way that the positioning of a lateral margin conformed to a east-west orientation, allowing lateral runoff to the north into Glacial Lake Aird. Although this scenario is highly speculative, it is currently thought to be the only viable theory which accounts for the evidence which has been observed.

4.3.2.4: Site 4 - Arbrack, Isle of Whithorn

An isolated landform approximately two and half kilometres north-west of the Isle of Whithorn, specifically described by Charlesworth (1926a) as a 'fine moraine of water-worn sands and gravels', contains a large exposure resulting from the commercial extraction of its constituent sand and gravel [exact location - NX 456373] (Figure 4.13, 4.26). Stereoscopic examination of this area, using a pair of 1:24 000 aerial photographs, indicates that the landform is actually composed of a small elongate mound (*c.* 100 m long) that is partially coalescent with a larger mound of similar morphology (*c.* 300 m long), from which the material has been extracted creating the exposure (Figure 4.27). Both mounds possess an asymmetrical cross-sectional profile (steep slope facing to the north-east and a comparatively gentler slope to the south-west). In planform, the strike line of the north-east facing slope of both mounds is nearly constantly linear throughout, whereas the other slopes splay out on the opposite side in all directions (west, through south-west, to south and south-east) from the individual high points of the mounds.

The presence of this landform is somewhat uncharacteristic of the general topography of the area which is subdued and gently undulating due to a multitude of near-parallel elongate depressions (Figure 4.27). These depressions exist predominantly to the south of the site where they very gently rise from around 30 m above sea-level at Arbrack to an elevation of approximately 70 m above sea-level, before falling back down to 10 m above sea-level adjacent

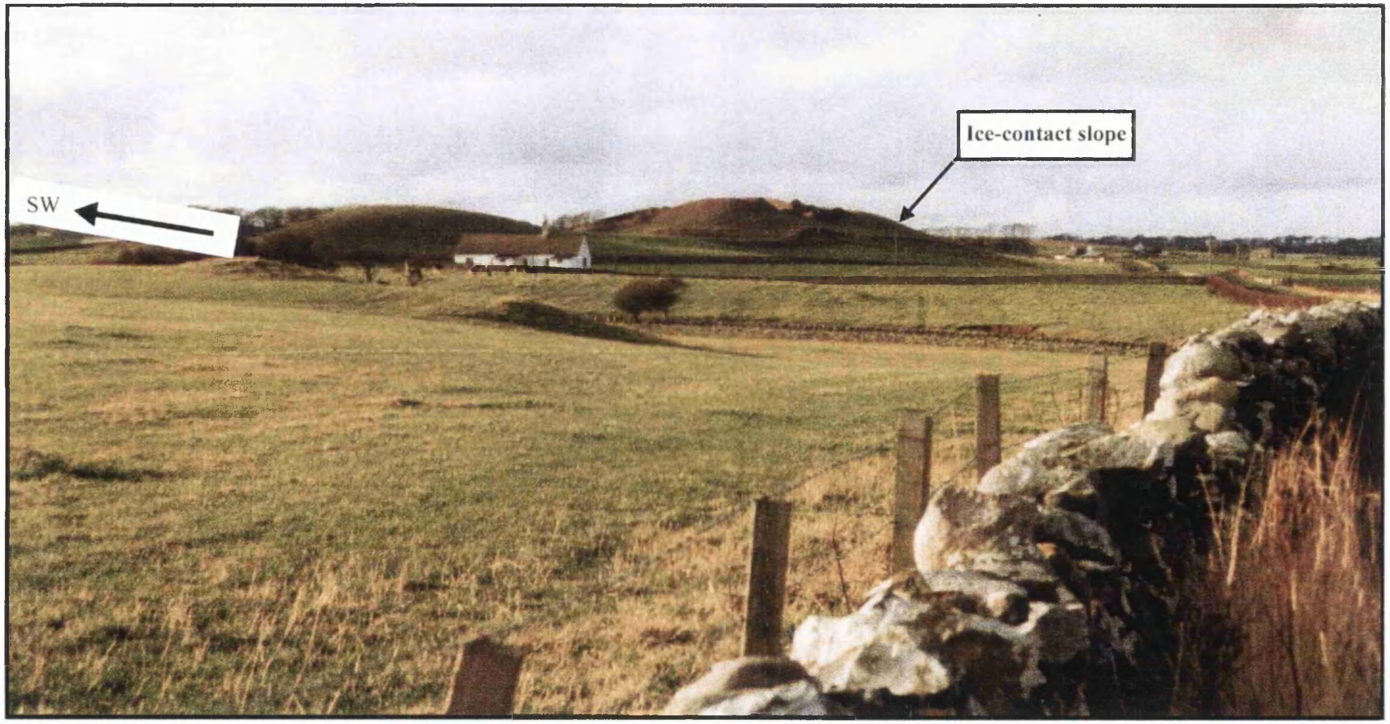


Figure 4.26 Arbrack site in the context of the surrounding relief.

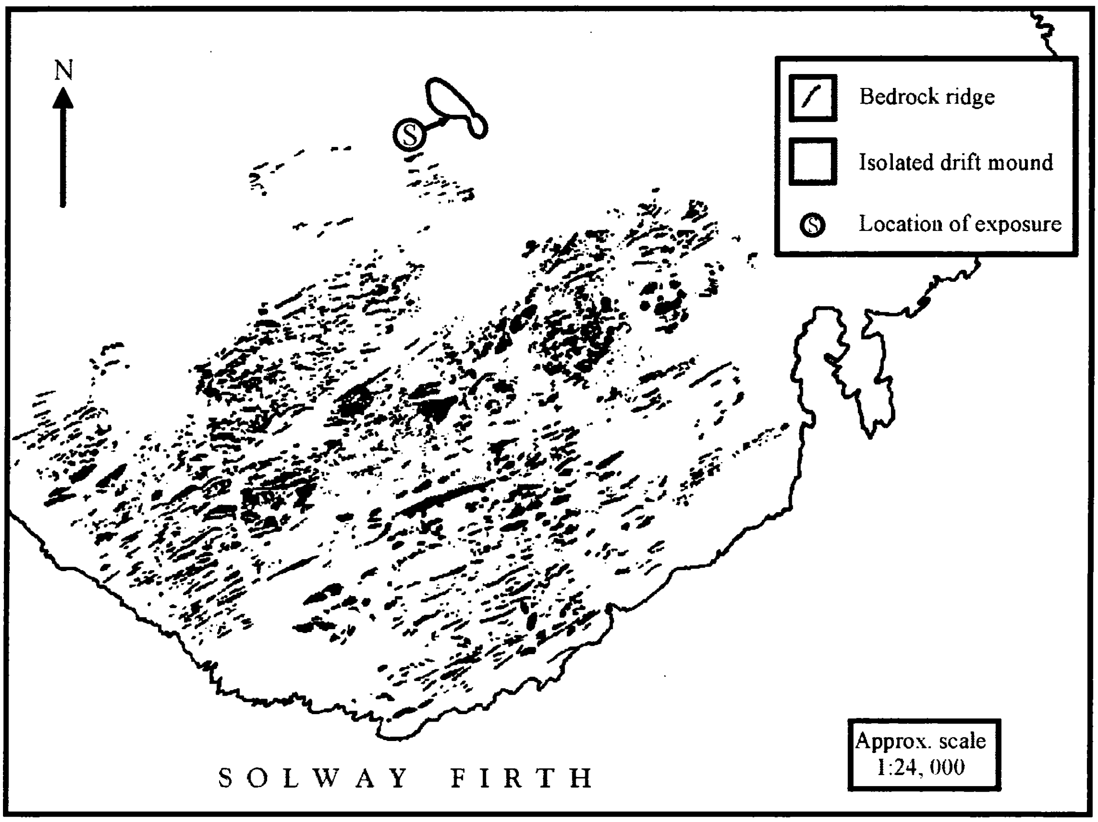


Figure 4.27 Arbrack site locational map, indicating the positions of the major geomorphological features in the southern region of the Machars peninsula.

to the Solway Firth coastline. The depressions, some of which measure up to 500 m in length, 50 m wide and 5 m deep, are formed in bedrock and exhibit many characteristics of P-forms, or plastically moulded forms (Dahl, 1965). Three broad types of P-form have previously been reported according to whether they are parallel or transverse to former ice flow, or non-directional. The P-forms displayed near Arbrack appear to conform, for the most part, with the characteristics of longitudinal forms, with elongate furrows (flow-parallel grooves) aligned parallel to former ice flow (Kor *et al.* 1991). Four media have been invoked to explain the erosion of P-forms, the importance of each having been stressed at particular times through various studies (Gjessing, 1965; Gray, 1981, 1982; Rea *et al.*, 2000): (a) debris-rich basal ice; (b) saturated till flowing between the ice-bedrock interface; (c) subglacial meltwater under high pressure; and (d) ice-water mixtures. The question as to how P-forms develop remains unresolved, with some workers suggesting a hybrid origin for at least some forms (e.g. Boulton, 1974; Gray, 1992), accepting the role of both fluvial and glacial abrasion. An excellent review of the theories behind P-form development is presented in Benn and Evans (1998, p. 317-323). The significance of their development in this location with respect to the palaeo-ice sheet dynamics and deglacial history of the Southern Upland ice sheet is discussed in Chapter 5.

The exposure on the south-west side of the main mound at Arbrack, to the north of the P-forms is approximately 150 m in length, with a maximum depth of between 10 and 15m, tapering in height towards both lateral margins. The stratigraphy is more obviously revealed by three undisturbed sections, illustrated in Figure 4.28, which are separated by slumped material covering the remaining portions of the exposure. The near-vertical section on the west-south-west side exposes a succession of steeply dipping, occasionally repetitive, facies of matrix-supported gravels (Gms), upward-fining sequences of gravels (Guf, Gruf), large boulders (BG), very coarse horizontally bedded sands with isolated dropstones, and homogeneous granules containing a comminuted molluscan fauna (Figure 4.29). A complete bivalve sampled from slumping beneath this section yielded an age of 35 800 ± 770 years B.P. (AA-33829) (Figure 4.30). The dip of the contact between each sediment unit was measured, and found to range between 25° and 28° towards the west-south-west (*c.* 245°). The angle of dip was measured where the contact between adjoining facies could be easily accessed on the exposure. Also revealed in the west-south-west section is an erosional contact between a generally massive granule unit (GRm) (with occasional signs of upward fining (GRuf)) and the lateral margin of several stacked, and at times repeated, facies units of gravels with differing levels of sorting and modal grain size (e.g. Gruf, Gm, Gms, Guf). The 'middle' section of the exposure revealing the upper stratigraphy of the east-south-east flank, is less extensive in size than the above, though exposes very similar facies (matrix-supported gravels (Gms), massive gravels

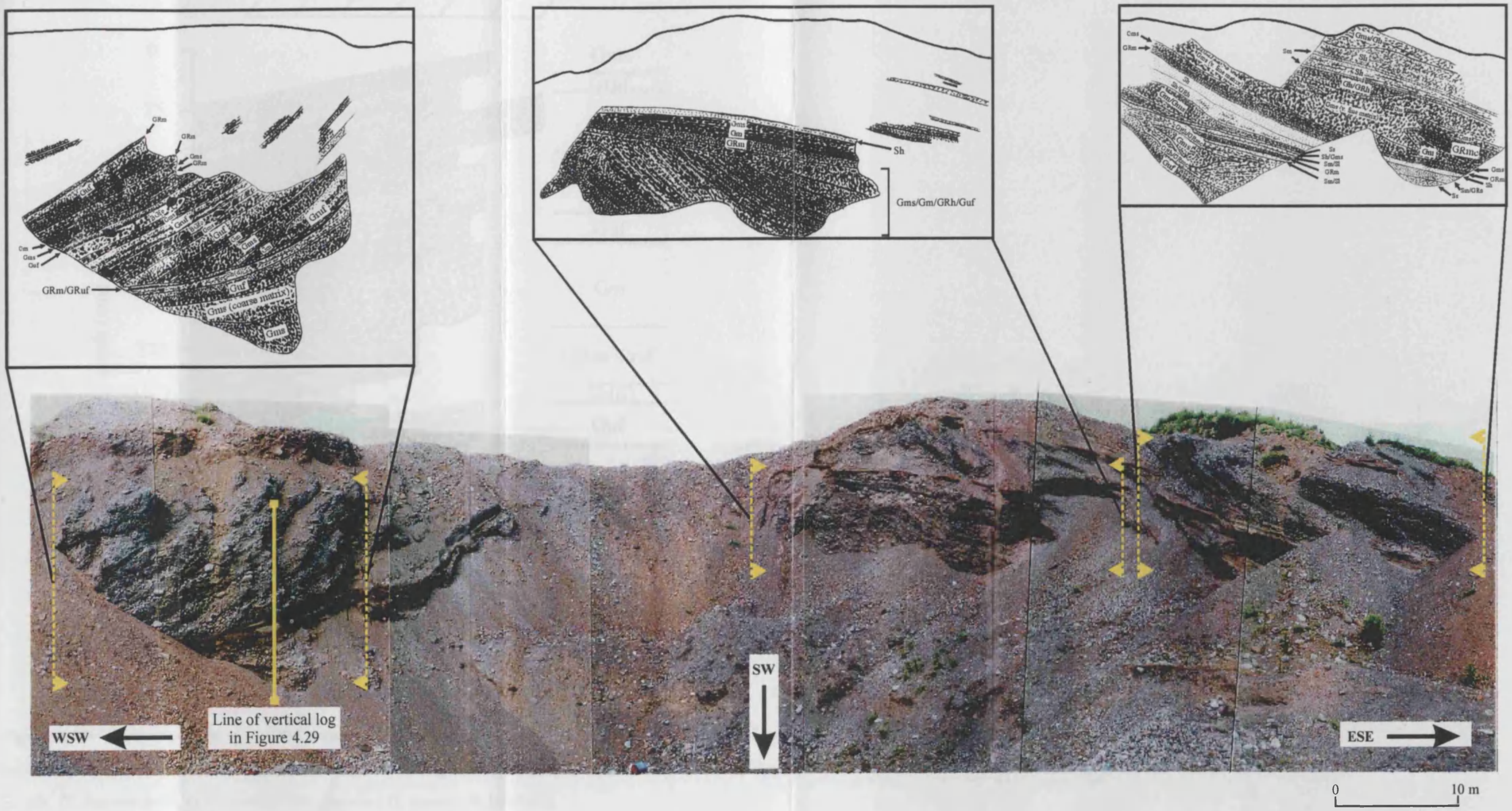


Figure 4.28 Photograph montage of the south-westerly facing exposure at Arbrack, with main stratigraphic sections. See Figure 3.15 for explanation of facies code.

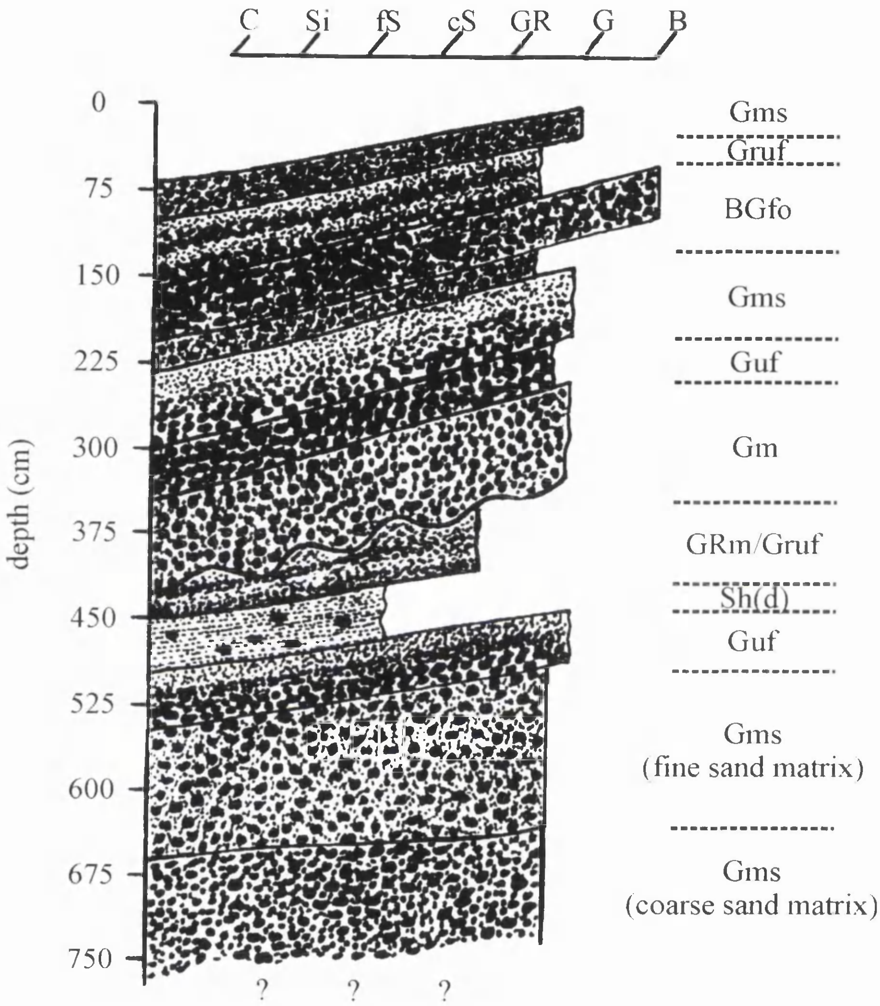


Figure 4.29 Vertical profile log of the western flank from the exposure at Arbrack (positioning of log indicated in Figure 4.23). See Figures 3.15 and 3.16 for explanation of facies code and symbols. C, clay; Si, silt; fS, fine sand; cS, coarse sand; GR, granules; G, gravel; B, boulders.

Figure 4.30 Bivalve mollusc sampled from slumped material beneath the west-south-west section at Arbrack, which dated at 35 800 ± 770 years B.P.

(Gm), gravels with upward-fining signatures (Guf), and horizontally-bedded granules (GRh) and sands (Sh)). The attitude of the sediment beds making up the lower half of this section are, however, strikingly different in that they trend towards the east-south-east, at angles decreasing from 45° to 15° moving along the inclined stacked sequence towards the east-south-east (*c.* 100°) (Figure 4.28). Capping the lower facies sequence, and thereby attributing to the upper half of the section, is a succession of gravels and granules which are less steeply inclined (<15°), and marked by a prominent erosional contact at their base. The third section on the lateral margin of the east-south-east flank appears to overlap the middle section in its upper reaches, and as a consequence reveals more steeply dipping sediment units towards the top (Figure 4.28). The stratigraphy portrays a complex succession of facies with a repeated ordering, moving up-sequence from poorly-sorted matrix-supported gravels (Gms) to horizontally bedded granules (GRh) to stratified coarse-grained sands (Sh). Also interspersed with these units are enclaves of massive gravels (Gm) and granules (GRm), with the latter supporting isolated oversized clasts in places (GRmc).

In summary, there are four major facies which have been defined within the three stratigraphic sections revealed by the south-westerly facing exposure at Arbrack:

1. Poorly-sorted, coarse-grained heterogeneous gravels and granules, occasionally stratified, supported in a matrix of mainly coarse-grained sand.
2. Well-sorted heterogeneous gravels and granules with either single or repeated fining-upwards sequences, with interspersed mollusc fragments (a scarcity of whole specimens).

These two facies quantitatively dominate the three sections within the Arbrack exposure, but in a way which does not demonstrate volume, but rather regularity within the large succession of sedimentary facies. Despite the prominence of upward-fining within individual units, the impression is also provided of a secondary larger-scale succession of repeated upward-fining within the entire stratigraphic sequence. Also making up this repeated succession are the following facies which occur on a less regular basis:

3. Horizontally-bedded coarse-grained sands with occasional lenses of granules and isolated dropstones, which are more prevalent towards the lateral margin of the exposure.
4. Boulder units containing clasts up to 15 cm in diameter in a clast-supported matrix.

On the basis of the observations detailed above, facies 1-4 have been assigned to a lithofacies association which, despite occasional interjection of units disrupting the sequence, reflects a repeated fining upward sequence of (1) boulders; (2) matrix-supported and clast-supported gravels (occasionally stratified); (3) massive granules (occasionally stratified); and (4) stratified sands. This lithofacies association is observed to repeat itself throughout the stratigraphic succession exhibited at the Arbrack site.

The dipping units reported in the three sections, categorised as clinoforms (facies 1-4), are interpreted as foreset beds deposited on a prograding subaqueous slope, the high angle of which indicates gravity-driven transport and deposition. The presence of normal grading and regular sorting among many units indicates that some flows were dilute (in terms of sediment concentration) and turbulent, allowing grains to move relative to each other (cf. Lowe, 1982; Nemec and Steel, 1984; Postma, 1986). The bedded sands revealed towards the east-south-east may represent deposition of tractive bedload, possibly at the base of low-concentration turbulent underflows (cf. Lowe, 1982). Alternatively, the stratified units could infer sediment remobilisation by gravity flow due to oversteepening of the fan surface (cf. McCabe *et al.*, 1984). The presence of sand units overlying massive gravels suggests that some low-concentration flows may have originated by dilution and winnowing of high-concentration mass flows during transport (cf. Benn, 1996). Ice-rafted debris found among the more distal sand units are also indicative of this type of depositional environment. Erosional contacts beneath some units record the erosion and remobilisation of sediment, associated with the influx of new material onto the slope.

The overall grain-size and sorting characteristics revealed in the three sections, and the observation that the foreset beds can be traced downwards from a central apex, indicates that debris delivery to the top of the prograding slope was by sediment-laden meltwater discharging south-westwards into a water body. It is inferred that the meltwater was discharged subaqueously in such a way which created a sediment plume, to account for the prevalence of well-sorted gravel units and the lack of fine-grained sediment. Powell (1990) suggests that such criteria are diagnostic of glacimarine environments where the buoyancy of inflowing waters tends to produce plumes which carry fine-grained sediments away from the depocentre, unless sediment concentrations are high. The suspended sediment spreads laterally until it is able to settle through the water column. Cowan and Powell (1990) suggested from detailed observations at tidewater margins in Alaska that this is most likely to occur during ebb tides, when internal turbulence within plumes is minimal. Sediment plumes can also be generated if the meltwater is discharged into the subaqueous environment from an englacial portal, thereby introducing sediment into the water column at a higher level, allowing fine-grained particles to be transported away to a more distal position.

The effects of varying water and sediment discharge are also thought to have an important effect on fan development in glacimarine environments (Powell, 1990). With low magnitude discharges, efflux jets have low momentum and so rise to the surface of the water column immediately. However at the other end of the scale, high sediment discharges have a high enough jet momentum to delay buoyant rising of the suspended sediment plume (Powell, 1990). It is suggested by the author, however, that a similar high magnitude discharge into a lacustrine environment would also cause suspended sediment to be dispersed some distance from the depocentre, and therefore the lack of fine-grained sediment within the exposure at Arbrack cannot be solely relied upon as evidence to verify a glacimarine setting.

The morphology and facies architecture of the Arbrack site leads the author to conclude that the isolated landform represents a former fan-shaped depocentre, which evolved from the deposition of masses of sediment built out into a water body by englacial and/or subglacial processes, and by a combination of gravitational mass movement and suspension settling below water-level (Powell, 1990; Powell and Domack, 1995; Figure 4.31). The isolated nature of the form, within an area which is otherwise topographically subdued, and the presence of a steep (ice-proximal) face lends itself to the morphological characteristics of a grounding-line fan. It is further suggested that the main landform described above may belong to a coalescing fan complex in that it grades imperceptibly into a morphologically similar, though less extensive, mound to the south-east (Figure 4.27). Whilst there are many similarities between ice-contact fans from both terrestrial and marine depositional settings, the absence of dead ice collapse features on the proximal flank (common to terrestrial ice-contact fans), and the concentration of the coarsest-grain materials at the ice-contact point (with distal reduction in grain-size) suggests a glacimarine setting for the development of this feature (Boulton, 1986). The absence of horizontally-stacked topset beds infers that the deposited sediment did not aggrade to water level, reflecting either a limited re-equilibration period of the retreating ice-margin, or tidal erosion of the upper surface (Powell, 1990). A rounded surface on the top of the stratigraphic sequence lends credence to the former scenario.

If a glacimarine setting were to be accepted then this would have considerable implications for the status of sea-level during the period when ice receded out of the Solway Firth onto this part of the Machars coastline. The upper limit of the foresets at Arbrack are recorded at a elevation of 40 m above sea-level which would imply a minimum sea-level of this height, in the context of a glacimarine model. Jardine (1971), who has carried out extensive observations on palaeo sea-level evidence in south-west Scotland has never identified elevated pre-Holocene marine limits and therefore it has generally been assumed that when the coastal area became ice free

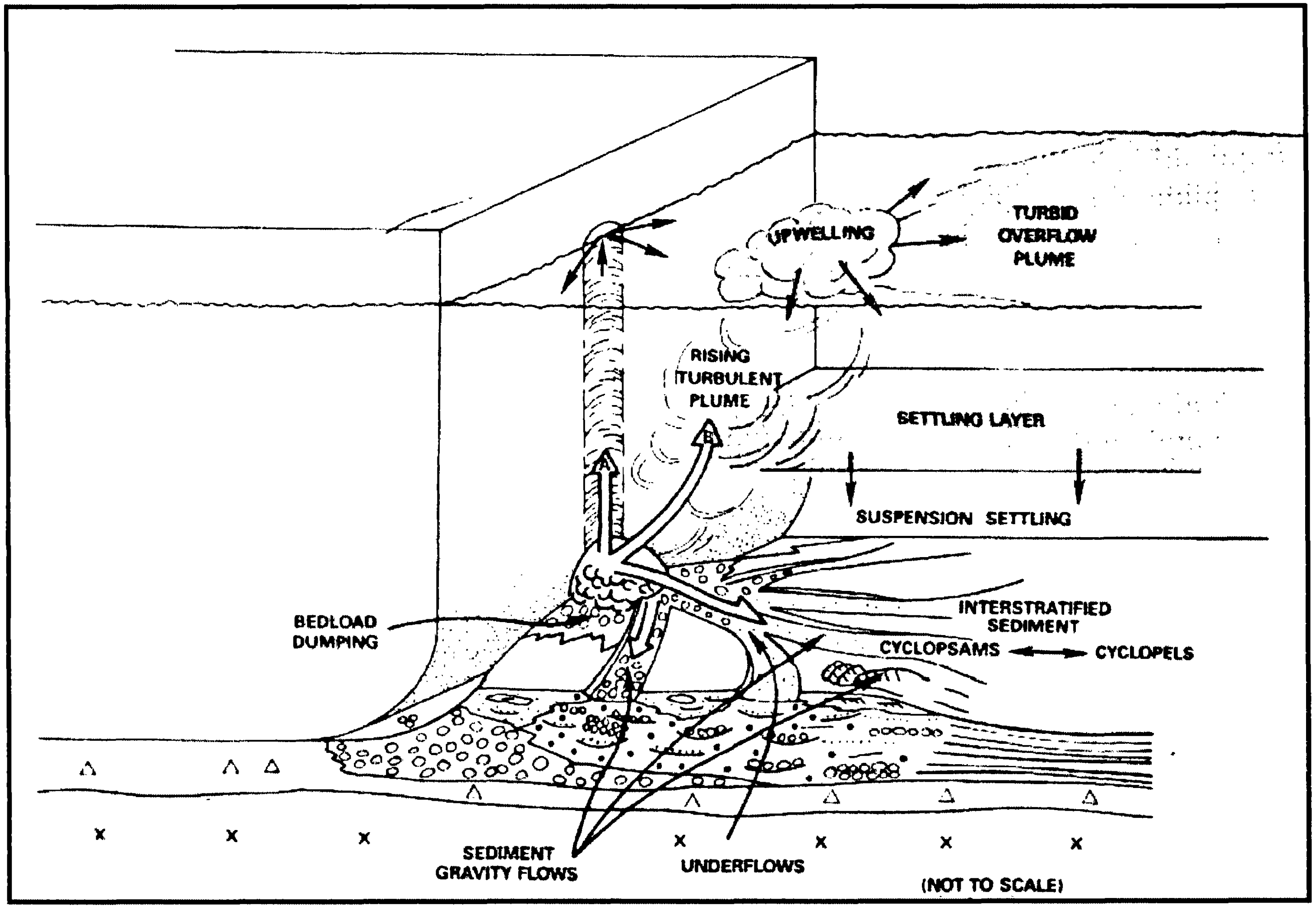


Figure 4.31 Processes and sediment associations of submarine outwash produced from subglacial discharges at tidewater fronts of temperate glaciers. This form of depositional system is invoked to explain the development of the ice-grounded fan at Arbrack. (From Powell and Molnia, 1989).

around 14 000 years B.P., sea-level must have been near or below the current level at that time. Indeed, the oldest indicators of sea-level documented along this coastline occur in the upper reaches of the Solway Firth and span a time interval of 12 290 ± 250 to 10 300 ± 185 years B.P. (Bishop and Coope, 1977) and suggest that mean sea-level in this period was at -3 to -4 m below the present level.

If a glaciolacustrine setting were to be accepted then the damming of water in this location would have to be explained. If consideration is given to the positioning of the grounding-line fan, with respect to the surrounding relief, then the formation of a glacial lake to the south-west of Arbrack is conceivably possible if particular circumstances arose. As highlighted above, the grounding-line fan is positioned in a slight depression at just over 30 m above sea-level and is surrounded to the south-west by a ridge of longitudinal P-forms (*c.* 70 m above sea-level). To the north-west and west the topography rises to an elevation of around 60 m above sea-level. The depocentre at Arbrack is therefore located across a shallow trough that opens to the south-west, in the same direction as the prograding foresets, finally terminating in conjunction with the coastline. The most plausible scenario which could produce a glaciolacustrine depositional setting is through the damming of the depression at the seaboard margin by ice situated along the Solway Firth Basin (Lake District ice?), and by the issuing of water and sediment from a north-easterly retreating Southern Upland ice flow (into the Cree valley?).

The presence of this subaquatic depositional system forms an important component of the record left by the Late Devensian ice sheet, as it provides an intuitive insight into the dynamics of the former Irish Sea ice sheet, and in particular the pattern of deglaciation over this region of south-west Scotland. The full implications of this evidence are discussed in Chapter 5.

4.3.3: Minor stratigraphic exposures

Other discrete exposures revealing sedimentological and stratigraphic evidence have been discovered at localities elsewhere in the Carrick and Galloway regions. The small size of these sections subsequently provides only limited evidence to form the basis for an interpretation of their depositional setting. Therefore, the inferences outlined below are conjectural.

4.3.3.1: Penninghame, near Newton Stewart

A recently excavated trench within a field, 4.5 km north-west of Newton Stewart [exact position - NX 384687] (Figure 4.13), has exposed a channel-like depression with infilled sediments, incised into horizontally-bedded sediments. The width of the channel is estimated to be about 10 m, although just over one-half of the channel width is currently exposed. Its full depth is also not visible, but a thickness of 1.75 m of the channel-fill sediment was measured in the axis of the depression. The sequence of the sediments (lithofacies association 1) reveals a series of horizontally-bedded channel infill deposits which include a number of the facies which Miall (1985, 1992) has recognised as typically defining this style of fluvial deposition. These include a predominantly coarse gravel (up to 0.75 m in thickness), the main constituent of this sedimentary sequence forming the basal unit which occasionally shows discrete fining upwards cycles. Above this facies occurs a compact sequence of massive gravels, granules and thin beds of horizontally stratified sands and thicker units of rippled sands which are laterally more extensive. The base is erosional and is channelled (lies on a concave-up bounding surface) into the underlying basal sediments of the exposure which are of indeterminate thickness (0.4 m exposed). These comprise a medium-grained sand facies on top of well-sorted gravels (Lithofacies association 2).

The fluvial architecture of these deposits appears to indicate, from the evidence that has been made available, a change in fluvial style recorded through the transition from LFA 1 to LFA2. These sedimentological changes could indicate a transition from a single-threaded channel system (LFA 1) to perhaps a multithread channel system (LFA 2), which reflects changing sediment and/or discharge conditions during the localised retreat of ice in this region towards the upland massif.

4.3.3.2: Colmonell, near Ballantrae

A small section (*c.* 5 m^2) on the south-south-east bank of the River Stinchar [exact position - NX 147856] (Figure 4.13) reveals a basic stratigraphy of predominantly medium- to coarse-grained gravels, occasionally stratified, containing what can be described as drapes of fine- to medium-grained sand and silt with occasional thin lenses of clay. These drapes vary in thickness from 2 to 10 cm, though appear to be quite laterally extensive (>1 m) pinching out towards their ends. The nature of the contact between the individual drapes and underlying gravel unit suggests that this finer-grained sediment blankets the underlying deposit, which itself is considerably thicker than the draped facies. On top of the sequence is thought to be a colluvium deposit although this is only partly exposed and therefore indeterminate. A reasonable interpretation of this sediment section is that the gravel facies have formed by accretion within a glaciofluvial system, possibly relating to the retreat of valley-constrained ice towards the centre of ice-dispersal over the western region of the Southern Uplands. The occurrence of draped material of such fine modal grain size over the gravels could imply that there were episodes of waning flow which allowed the finer particles to settle out from suspension, rather than being incorporated in the matrix of the gravels. Miall (1977) suggests that the thinning out of the drapes towards their lateral edges is indicative of the settling of fine-grained particles in pools of stagnant water, or abandoned meltwater channels.

4.4: Synopsis

This chapter has detailed the character and extent of the ice-flow directional record, and described the glacial geomorphology and stratigraphy of a range of depositional settings throughout south-west Scotland. The following sub-sections highlight the principal points which have been raised and are intended as a cross-reference for Chapter 5.

4.4.1: Patterns of drift lineation and relative chronology

- Satellite imagery of south-west Scotland has revealed an extensive coverage of drift bedforms and suites of superimposed lineations of varying directions which deviate significantly from the single lineation pattern used by early workers to infer a radial, unopposed ice-dispersal pattern centred over the Southern Uplands. By promoting a single lineation pattern, it would appear that earlier workers have ignored the significance of the varying lineation orientations which were originally mapped by Charlesworth (1926a) and subsequently detected by the analysis of satellite imagery.
- The application of remote sensing has provided a means of identifying palimpsests of drift lineations representing former ice-flow dynamics beneath the more recent lineation pattern, which previously had remained undetected.
- Twenty-eight coherent flow sets are recognised over Carrick and Galloway which have been correlated, their relative ages determined, and a sequence of ten flow stages established on the basis of their glaciological compatibility.

4.4.2: Age of the ice-flow stages

- No unequivocal evidence exists to provide an indication of the age of the individual flow stages.
- Evidence relating to pre-Late Devensian deposits in south-west Scotland are unreported, therefore it is tentatively suggested that the present drift lineations developed during the Late Devensian.

4.4.3: Evolution of the ice sheet through the last glacial cycle

- *Flow stage A*: Represented by limited NE-SW orientated lineations in Ballantrae district (flow set 13), which have been correlated with lineations of identical orientation over the north-west and west-central Rhins peninsula (flow set 16) (NW-SE ice divide ridge). Two scenarios explain the scarcity of lineations representing this stage: (i) lineations reflect earliest Devensian ice sheet expansion; or (ii) lineations reflect residuals from an early phase of Late Devensian ice flow which survived overprinting and modification by later more powerful ice flow during the period of maximum glaciation.

- *Flow stage pre-B*: Represented by N-S aligned lineations which overprint, and thereby post-date, flow set 13 of stage A over Glenluce district. Two scenarios explain the development of this stage: (i) expansion of Southern Upland ice into Glenluce due to a shift in the ice-divide from an alignment of NW-SE to E-W; or (ii) early dominance of more powerful Highland ice flowing in a southerly direction from the Firth of Clyde, which supports the inference of Charlesworth (1926a).
- *Flow stage B*: Represented by flow set 15 comprising SW orientated lineations close to the upland region which gradually realign to a NNE-SSW orientation throughout Glenluce district, and whose trend further extends over the southern part of the Rhins peninsula. Flow set 15 has been correlated with flow set 11 which is represented by lineations of similar alignment, spatial distribution and concentration. Two scenarios explain the development of this stage: (i) Easing off of Highland ice to the west, based on the acceptance of scenario two of stage pre-B, allowing the overprinting of flow set 14 in Glenluce by developing Southern Upland ice of stage B. (ii) Coalescing and manipulation of Highland ice, based on the acceptance of scenario two of stage pre-B, resulting in the divergence of Highland ice into a SSE flow direction dictated by the local ice stream.
- *Flow stage post-B*: Represented predominantly by lineations of flow set 12 in an area to the SW of the Machars. It is hypothesised that this region marks a coalescing point of local ice (flow set 11 of stage B) and Lake District ice (flow set 12). The scenario which is envisaged is that synchronous W to NW flowing Lake District ice converged and was subsequently reoriented by SSW flowing Southern Upland ice. This is supported by (i) the absence of a topographic barrier which might alternatively explain an abrupt change in generalised flow from NW to SSW; and (ii) the lack of an obvious local ice-dispersal centre which would provide a dominant westerly flowing ice stream.
- *Flow stage C*: Represented by NNE-SSW orientated lineations on the western seaboard (flow sets 17-19-23) reflecting the southerly penetration of ice from a source area to the north of Carrick and Galloway. This alignment supports the evidence used by Charlesworth (1926a) to infer a similar scenario, and further explains the presence of extraneous paisanite erratics on the shoreline of Loch Ryan. Isolated lineation pockets of stage C are explained by: (i) supersedence by lineations of later flow sets (stage D-1 and D-2) reflecting ice of local provenance; and (ii) existence of topographic barriers restricting bedform development in specific areas. The first explanation relies upon temporary retreat, or repression, of the Highland ice sheet, arising in the convergence with the local ice streams of stage D-1 and D-2 off the position of the present coastline, and eventual rejuvenated SSW flow of both ice streams over the Rhins of Galloway.

- *Flow stage D-1 and D-2*: Flow sets of stage D-1 (21-22), thought to be synchronous with those of D-2 (25-26-27-28), are represented by lineations which reflect ice flow to the WNW, through W, to WSW, partly controlled over the seaboard by the underlying topography. Stage D-2 follows a similar pattern and notably cross-cuts older bedforms. Three scenarios explain the cross-cutting relationship: (i) Expansion of Charlesworth's (1926a) hypothesis whereby local ice flowing northwards (e.g. along the Doon valley) was diverted by Highland ice situated in the Firth of Clyde through W into SW, and later overprinted those flow sets which developed from westerly flowing ice from the Southern Uplands. (ii) Successive stages of ice margin recession towards the upland ice-dispersal centre to the east. (iii) Reorientation of the principal flow direction by rapid ice-divide shifts between positions of relative stability, inducing the development of superimposed forms.
- *Flow stage E*: Represented by SW-NE aligned lineations over the Rhins of Galloway which cross-cut the less concentrated and distributed lineations of flow set 17 (stage C), and are therefore considered to post-date them.
- *Flow stage F*: Represented by lineations of flow sets 1-3 which reflect topographically (valley) constrained flow of more or less independent glaciers occupying the Ken (set 1) and Cree (set 3) valleys. It is uncertain whether both were synchronous, but they appear to relate to a period when local ice became influenced by the valley topography. Both flow sets may have been overridden by a later ice readvance (stage G) within each respective valley (explaining their low concentration).
- *Flow stage G*: Represented by topographically constrained lineations (flow sets 2-4-5-6-7) within the major valleys, inferring a similar generalised pattern of ice flow to the overall character of Late Devensian glacial retreat modelled by Charlesworth (1926a). It is believed that stage G equates with Charlesworth's (1926b) Southern Upland Readvance of local ice radiating along the main valleys from the upland massif: the Ken valley (set 2); Fleet valley (set 6); Cree valley (set 4); and Doon valley (set 7). Flow set 5, on the eastern periphery of the Rhins peninsula, reflects the constrained flow of ice along the Loch Ryan Basin which has also been correlated with the Southern Upland Readvance, but is thought to have developed from the readvance of Highland ice. The superimposition of flow sets of stage F by those assigned to stage G supports the conjecture of a Southern Upland Readvance. The congruous alignment of the most recent flow sets implies only one major readvance of local ice, prior to gradual retreat as represented by Charlesworth's (1926a) retreat stages.

4.4.4: The Highland-Southern Upland Readvance

- An objective assessment of the geomorphological criteria used by Charlesworth (1926b) to formulate a Highland Readvance episode has been addressed by applying a modern reappraisal of the landform distribution throughout the Loch Ryan Basin.
- The general character of the geomorphology observed by Charlesworth (1926b) has been detailed. The principal evidence used by Charlesworth to infer a readvance are: (i) conspicuous moraines marking the limit of Highland readvance ice whose interpretation is substantiated; (ii) morphology typical of ice stagnation; (iii) a great outwash plain beyond the readvance limit; (iv) high concentration of meltwater channels; and (v) a stratigraphic signature which invokes high levels of drainage. Charlesworth's suggestions of the occurrence of ice margin oscillations during the readvance phase are based on the following evidence: (i) intercalation of till with outwash sands and gravels; (ii) dissection by marginal streams of earlier deposits from the initial readvance phase; (iii) 'choking' of marginal channels at their mouths by moraines; and (iv) the entire obliteration of marginal channels by subsequent infilling.
- A map showing the distribution of the major glacial landforms has been produced, which is accompanied by a description of the landforms and their distribution. These include: (i) lateral moraines; (ii) depositional ridges; (iii) drift lineations; (iv) kames and kame-and-kettle topography; (v) glaciotectonic depressions; and (vi) meltwater channels.

4.4.5: Palaeo-depositional settings

- *Site 1 - McCulloch's Point and Hillhead of Craichmore, Rhins of Galloway*: Interpreted as a far-travelled massive diamict which has been subjected to glacitectonic disturbance. A contrast between macrofabrics from the upper and lower portions of the exposure strongly indicate that there has been a change in the palaeo-ice flow direction along the Loch Ryan Basin from a transverse flow (lower fabric) to one corresponding closely with the general trend of the axis of the basin. Two explanations are provided: (i) the change in fabric up-section corresponds to a readjustment of a single phase ice flow; or (ii) the upper fabric records a superimposed strain signal resulting from glacitectonic modification by a later flow stage.
- *Site 2 - Pinminnoch, Rhins of Galloway*: Differentiated into two lithofacies associations: (LFA1) an upper, coarse-grained lithofacies association comprising gravels and stratified granules; (LFA2) a lower, fine-grained lithofacies association comprising clay, and stratified and rippled sands. The depositional history has been divided into two phases: (1) deposition of subaqueous deposits (LFA2); and (2) deposition of glaciofluvial sediment (LFA1). The spatial relationship between the two lithofacies can be explained by the

following scenarios: (i) renewed proximity to an ice margin and sediment source; or (ii) release of a high, pulsed discharge event from a relatively stabilised ice margin depositing a high sediment concentration, subsequent to a quiescent phase in sediment discharge.

- *Site 3 - Aird, Stranraer*: Interpreted as the upper and middle components of a Gilbert-type delta complex (upper horizontal topsets and inclined foresets) connected to a subaerial feeder river situated in a sandur environment. The plateau surface sits at 27 m above sea-level, and the top of the foresets is estimated at between 15 and 20 m above sea-level. Beyond the delta slope, to the north-west, lies a flat-bottomed basin composed predominantly of two sediment units which are believed to reflect the deposition of marine tidal mud flats on top of lacustrine clays and silts. A mollusc found at a depth of 210 cm (within the marine muds), yielded a date of 6 075 ± 55 years B.P. which infers a minimum deglacial date for this area. This date closely supports the height of sea-level inferred by Lambeck's sea-level curve at 6 000 years B.P. for the west Solway Firth, thereby providing evidence to suggest the inundation of shallow marine waters (possibly only at high-tide) into the basin at this time. It is proposed that the initial lacustrine environment, responsible for the deposition of the distal deltaic clays and silts, was formed by ice in the Loch Ryan Basin blocking the north-west opening of the Aird basin, creating Glacial Lake Aird. Eventual recession of the ice allowed lake waters to drain out of the basin, and marine waters later entered forming a shallow marine embayment. The subaerial meltwater drainage system feeding the delta is thought to have exited from an east-west aligned lateral margin of a piedmont lobe situated within the lowland area to the south of Stranraer.
- *Site 4 - Arbrack, Isle of Whithorn*: Interpreted as two coalescent grounding-line fans which evolved from the deposition of masses of sediment built out into a water body by englacial and/or subglacial processes, and by a combination of gravitational mass movement and suspension settling below water level. The interpretation of a glacimarine setting is possible due to the absence of features common to terrestrial ice-contact fans which have been reported in the literature. This type of depositional environment would have considerable implications for the status of sea-level during the period when ice receded out of the Solway Firth onto this part of the coastline. Absence of topsets suggests: (i) the fans did not aggrade to water level, reflecting either a limited re-equilibration period of the retreating ice margin; or (ii) tidal erosion of the upper surface of the fans. However, a glaciolacustrine interpretation can also be entertained if the depression, across which the grounding-line fans are positioned, was blocked at the seaward end by ice situated within the Solway Firth during a time when water and sediment were issuing from a north-easterly retreating Southern Upland ice flow. A bivalve from the foresets yielded an date of 35 800 ± 770 years B.P. which suggests that the fauna has been reworked.

CHAPTER FIVE

Discussion

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Discussion

5.1: Introduction

The overriding objective of this investigation was to expand upon the current level of knowledge which exists regarding the characteristics of events which took place during Late Devensian glaciation and deglaciation in the north-east sector of the Irish Sea Basin. More specifically, in light of the recently recognised temporal changes in ice sheet configurations, the validity of traditional models delineating Late Devensian glacial events in south-west Scotland was re-evaluated, thereby more clearly defining the glacial and deglacial chronology. Additionally, the lack of consideration and clarification of episodes such as ice retreat onto the mainland of south-west Scotland in the context of Irish Sea glaciation, and the controversial status of ice-sheet readvance that purportedly interrupted overall retreat of the Late Devensian ice sheet was addressed. Finally, the intention was to consider the sedimentological and glaciological effects of isostatic submergence, and the retreat of the ice sheet onto a part of the Scottish mainland which has remained substantially unconsidered in connection with Irish Sea glaciation. In doing so, the palaeoenvironmental interpretation of glacigenic sequences brings into consideration the influence of a marine ice sheet on the sedimentary record, rather than from the outset trying to apply a glacio-terrestrial model, as many other studies around the Irish Sea Basin have done.

This chapter summarises the major findings discovered in relation to the issues raised above, and attempts to relate them back, where appropriate, to the themes which were introduced in Chapters 2 and 3. The format for the discussion will follow a chronological framework which relates directly to the objectives outlined in Section 1.2.2. Thus, Section 5.2 provides a proposal for a revised model of the Late Devensian palaeo-ice sheet dynamics in south-west Scotland (Objectives 1 and 4); Section 5.3 discusses the character and mode of deglaciation in south-west Scotland (Objective 2); and Section 5.4 attempts to reconcile the status of Lateglacial readvance(s) in south-west Scotland (Objective 3). Finally, Section 5.5 discusses the history of sea-level in the north-east sector of the Irish Sea Basin. Clearly the sea-level record cannot be entirely separated from the deglacial history and therefore both are addressed when necessary.

5.2: Model for Late Devensian Palaeo-Ice Sheet Flow in South-West Scotland

5.2.1: Background

In Section 3.2.3 a strategy for applying Landsat imagery to reconstruct flow geometries through identification of quite often fragmentary geomorphological and geological evidence was outlined in detail. This recent methodological advance in reconstructing the evolution of former ice sheets has enabled a greater analysis and concrescence of the geomorphological evidence which exists in south-west Scotland. This is in contrast to the classic ice sheet models of Geikie (1901) and Charlesworth (1926a) whose interpretation of the internal components of ice sheet geometry are derived from rudimentary field-based approaches. Section 3.2.1 highlighted the main problems associated with attempts to arrive at a coherent synthesis of the available geomorphological evidence. From this, it is evident that the application of a traditional type of field investigation for the purpose of this sort of investigation is somewhat restrictive, and presents a number of limitations for regional-scale palaeo-ice sheet reconstructions. The fundamental principle which has failed to be considered during the construction of traditional ice sheet models has been that of temporal changes that an ice sheet flow-line is likely to experience during its long history. Geomorphological signatures reflecting this occurrence have only recently been discovered from the North American continent by Boulton and Clark (1990) (Section 3.2.1.2), and therefore it is understandable that the early attempts to reconstruct ice flow dynamics on regional or continental scales conformed to what could be described as the first principles of glaciological theorem during that era. A more comprehensive discussion of this critical issue is addressed in a wider context in Section 5.2.3.

The recently acquired knowledge of cross-cutting relationships between drift lineations, which has been brought to the forefront of glaciology by advanced remote sensing techniques, argues strongly for the need to re-examine long-standing ice sheet models, particularly in regions where the main geomorphological evidence exists in the form of well-preserved flutings and drumlins. On the grounds of the discovery in south-west Scotland of an extensive coverage of subglacial drift lineations and suites of superimposed lineations, from the interpretation of satellite imagery, the author has presented in Section 4.2 a revision of the evolving palaeo-ice sheet dynamics in Carrick and Galloway regions, south-west Scotland. In order to consolidate the information detailed in this reconstruction, the flow stages and generalised palaeo-ice flow dynamics which are thought to have prevailed during the history of the Late Devensian glaciation in this region are summarised in Figure 5.1. The main points accompanying the revised model are discussed in Section 5.2.2, in the context of the most glaciologically plausible interpretation that the available evidence presents. The author is aware, however, that this reconstruction is partially confounded by (i) the difficulty in accurately assigning a particular drift lineation to a flow set/stage; (ii) the absence of dateable evidence for accurately correlating flow sets into flow stages; and (iii) the absence of ice-moulded and ice-abraded features on upland terrain. Nevertheless, the reconstruction is viewed as being the first attempt to apply the methods pioneered by Boulton and Clark (1990) to the re-modelling of part of the Late Devensian British ice sheet.

The evolving palaeoglaciological structure of the south-west sector of the Scottish ice sheet during the Late Devensian has been reconstructed by reference to bedform alignment, location, morphological attributes and overprinted or cross-cutting relationships (Section 3.2.3, 4.2.1). In doing so, the time-dependent dynamic evolution of the ice sheet and the pattern of ice stream development are reconstructed. Spatially, glacial lineations showing a consistent directional signature are grouped into ten major flow stages: A, pre-B, B, post-B, C, D-1, D-2, E, F and G (Figure 5.1).

5.2.2: Evolution of the ice sheet and the pattern of ice stream development

FLOW STAGE A

During the initial phase of ice flow over Carrick and Galloway regions two basic scenarios were outlined to explain the limited, though consistently orientated north-east to south-west drift bedforms which exist in four distinct pockets (Figure 4.5a). The most favoured hypothesis is that the residual lineations of this flow stage reflect the earliest expansion of ice from an ice divide situated over the north-west to south-east upland axis delineating the Merrick range. Other lineations pertaining to this initial ice flow phase are considered to have been substantially overprinted and modified by later, more powerful ice flow associated with maximum glaciation. From the pattern of lineations which have been observed, this would suggest that flow stage A advanced at least 40 km over the lowland plains of Glenluce district and the Rhins of Galloway into the Irish Sea Basin. It is anticipated, however, that this distance is an underestimate and that if it is accepted that the dispersal centre was situated over the Merrick range, then this would invoke a minimum ice flow distance of nearer 55 km.

The alternative scenario (presented in Section 4.2.3.1), that the lineations of this stage reflect earliest Devensian ice sheet expansion, can be accommodated on the basis of the geomorphological evidence alone although it is considered to be unlikely based upon two important reasons outlined in Section 2.3:

(1) Evidence for the presence of glacier ice throughout the Irish Sea Basin is thought to be found only in the Late Devensian record (Bowen *et al.*, 1986). Within the British Isles, an Early Devensian glaciation has been demonstrated only in Ulster where till units contain marine molluscs of oxygen isotope sub-stage 5e as the youngest faunal element (Bowen *et al.*, 1986). It is not unreasonable, however, to infer that at this time the upland areas of Scotland were also glaciated, though evidence supporting this conjecture remains unfurnished. Nevertheless,

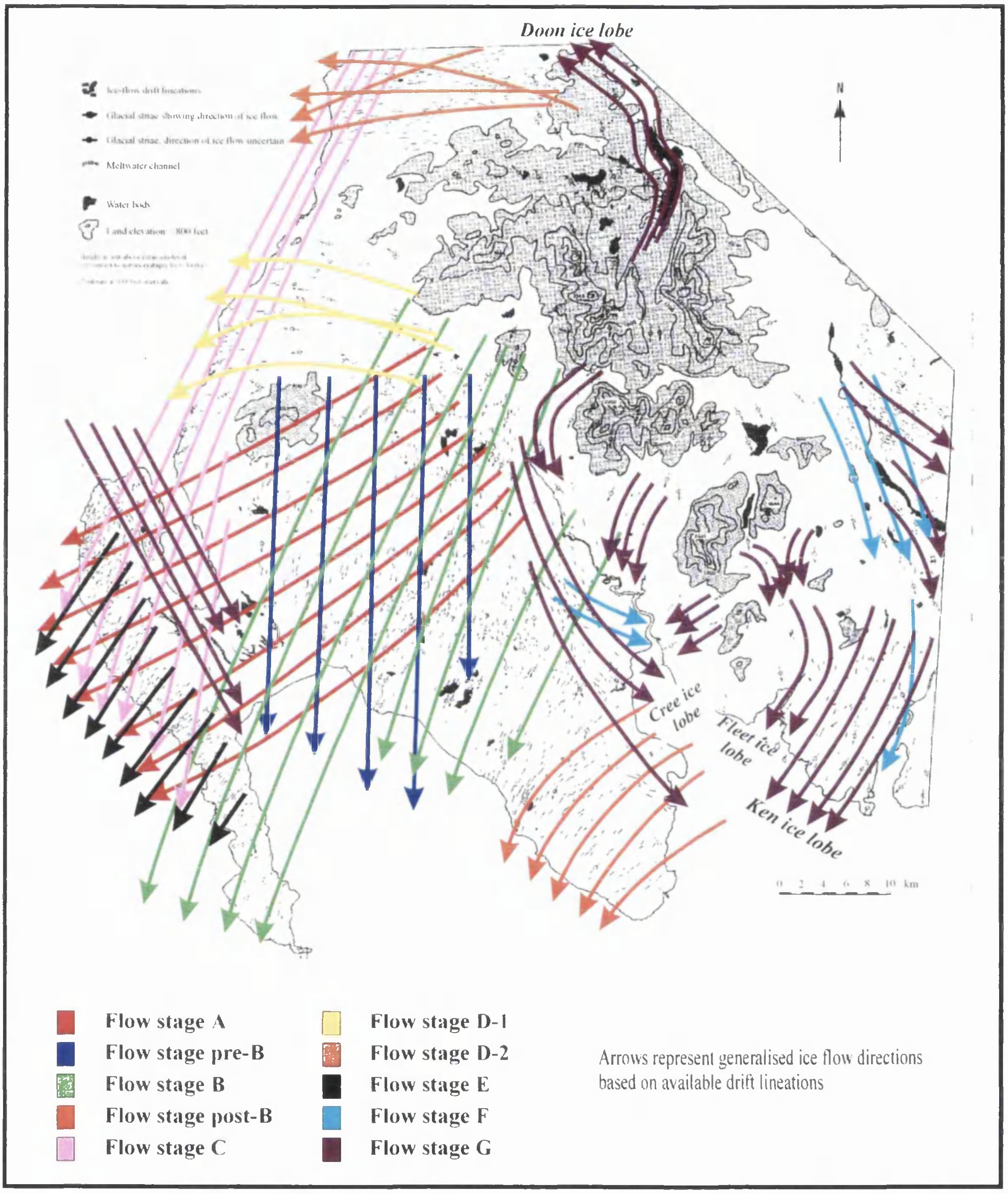

Figure 5.1 Model for Late Devensian palaeo-ice sheet flow in south-west Scotland.

stratigraphic evidence reflecting substantial glacio-isostatic depression of the crust during Early Devensian ice expansion over Scotland (Sutherland, 1981), further supported by implications of no major deglaciation after 75 000 years B.P., from an oxygen isotope curve of Shackleton and Opdyke (1973), does indicate the possibility of substantial glaciation during the Early Devensian, and therefore the likelihood that the residual lineations of flow stage A post-date this period cannot be confirmed (Section 2.3.1). However, the overall refusal of contemporary workers to adopt this latter theme leads the author to side in favour of the more popular viewpoint, until evidence is discovered over south-west Scotland to reject this.

(2) It is generally accepted that in Britain major expansion of ice from mountainous dispersal centres, of the magnitude which would be required to satisfy the apparent extent of ice flow during stage A, did not occur on a major scale until after 25 000 years B.P. (eg. Thomas, 1985).

FLOW STAGE PRE-B

Overprinting, and hence post-dating, the initial expansion of Late Devensian ice from the Southern Uplands, are more concentrated and spatially distributed lineations which together depict the southerly ice flow direction constituting stage pre-B (Figure 5.1). It is anticipated that the majority of lineations which developed during flow stage A have subsequently acted as protuberances to planar ice flow coming from an oblique angle to them, and have thereby served as an available sediment supply for the development of bedforms of stage pre-B and later stage B. Indeed, it is evident that stage pre-B has itself been subjected to considerable overprinting and deformation by flow stage B, with lineations pertaining to this stage suggesting initial high levels of distribution throughout Glenluce district prior to re-modification, due to their current confinement to only the eastern and western peripheries of this region, on the margin of the zone occupied by stage B lineation (Figure 4.5c).

Both cases of overprinting imply the process of 'pre-existing lineation deformation' (Clark, 1993), as there is no evidence to suggest the occurrence of superimposition of stage A lineations by those of stage pre-B, and likewise stage pre-B by those of stage B (Section 3.2.1.2). This hypothesis coincides with the thoughts of Prest (1984) and Boulton *et al.* (1985), who previously considered the ability of a dynamic ice sheet to reorganise sediment as being potentially high. It is suggested that such a relationship depicts a zone of high ice flow velocity which resulted in the strong deformation of pre-existing bedforms, and almost total reorganisation of sediment into an entirely new orientation during this period of glaciation (Figure 3.3, 3.4). Other parameters which must also be considered as possibly responsible are the time which has elapsed between consecutive stages, ice thickness, and intricacies of sediment rheology (Clark, 1993). These parameters have not been addressed in this investigation.

From the two scenarios presented in Section 4.2.3.2 the author is more in favour of a shift in the location of the ice-divide to that of an east-west alignment, which would account for the generalised ice flow direction depicted by lineations of stage pre-B. This hypothesis has been supported by Clark (1993, 1997) as being the most common explanation for the documentation of cross-cutting flow directions where superimposition is absent, and is therefore adopted here in light of no stronger, alternative theory. Although a small number of intermediate orientations would be expected if the ice flow direction gradually changed from NW-SE (stage A) to E-W (stage pre-B), the existence of some form of on/off mechanism of lineation generation under the emanating flowlines is called upon to satisfactorily explain the occurrence of the two discrete orientations (Clark, 1993; Section 3.2.1.2).

The alternative scenario, involving the early dominance of powerful Highland ice streaming from the Firth of Clyde over the districts of Girvan and Ballantrae, is currently thought to arise at a later stage in the glacial history. The reasons for suggesting this are:

(1) It is argued that the southerly positioning of Highland ice over the Glenluce area, implemented here for the purpose of discussion, would have arisen during the maximum period of Late Devensian glaciation in south-west Scotland. This period of glaciation is not thought to relate to flow stage pre-B, principally because of the extent to which at least one other ice flow stage overprints and re-modifies lineations of this stage. It is not envisaged that further ice flows of such magnitude (*c*. 50 km in length), as seen to cross-cut stage pre-B (stages B and possibly F), would have been initiated during the post-maximum glacial period.

(2) Alternative ice-flow directional evidence (erratics and striations) from earlier studies, which correspond with an inferred southerly passage of Highland ice at some stage during the Late Devensian glaciation, do not invoke the presence of Highland ice as far east over the lowlands of south-west Scotland as the generalised ice flow lines depict in Figure 5.1. The only previous evidence inferring the former flow of Highland ice over south-west Scotland has been discovered along the coastal margins of Loch Ryan, Girvan and Ballantrae districts, and over the Rhins of Galloway (Charlesworth, 1926a).

FLOW STAGE B

This flow stage represents a predominant south-west flow over the lowlands of Glenluce district, positioning an ice divide aligned along the main upland masses of Merrick, Lamachan

Hill and Cairnsmore of Fleet (Figure 5.1). As inferred above, the positioning of the lineations assigned to flow stage B mark out a distinct pathway through those lineations assigned to previous flow stages, in particular flow stage pre-B. This pattern can be clearly visualised in Figure 4.5c, and again the general lack of superimposition suggests high ice flow velocity during this stage. The direction of flow stage B can again be satisfactorily explained by a shift in the positioning of the main ice divide from the east-west alignment of previous stage pre-B. This presents the possibility of equating this flow stage to the earlier phase of ice flow which is claimed to have been responsible for the development of the residual elements depicting flow stage A. If this compelling theory is implemented, it requires a readjustment of the ice-divide to a similar position to that where it originally existed during the initial phase of ice sheet evolution. If this is such, the main implication to arrive from this hypothesis is that the ice sheet over this part of the Southern Uplands had a centre of mass and flow radiation which were relatively fixed, but which did experience an episode of dramatic reorientation.

The earlier suggestion in Section 4.2.3.3 that the ice flow of stage A experienced episodes of decay, during which time Highland ice of stage pre-B superimposed itself as the dominant southerly flow (prior to the reorientation major expansion of stage B), is rejected on the basis of the preferred interpretation provided above. The reorientation of the principal flow direction from south-west, near to the source area, to one demonstrating a south-south-west direction over the Rhins of Galloway is, however, considered to be strongly linked to the interaction of a more dominant ice stream flowing down the western seaboard (Flow stage C). Therefore the author is in favour of the second scenario outlined in Section 4.2.3.3, thereby discounting the impact of Highland ice on flow stage B until later on, when it is claimed that the two ice flows coalesced over the Rhins of Galloway, some 50 km from the source area of stage B as discussed below.

FLOW STAGE POST-B

Flow stage post-B clearly demonstrates an overprinting and realignment of drift lineations from previous stage B, predominantly in the area to the south-west of the Machars district (Figure 4.6, 5.1). The temporal period between this phase of ice flow and that of earlier stage B is indeterminate, therefore it can only be confidently ascertained that it reflects a period post-dating the events of stage B. The scenario outlined in Section 4.2.3.4 is maintained, and summarised here. It is strongly favoured that the flow convergence which has been observed on satellite imagery represents a zone of coalescence between local Southern Uplands ice of stage B (zone 1 of Figure 4.6) and Lake District ice of stage post-B (zone 2 of Figure 4.6). The inferred sequence of events relies upon the ice flowing from the Lake District massif in a westerly to north-westerly direction and coalescing, and being subsequently reoriented, by south-south-west flowing local Southern Upland ice of stage B over the Machars of south-west Scotland. The deviation in the generalised flow direction of stage post-B over this region by Lake District ice is confidently argued, as all other scenarios considered by the author have failed to provide a plausible explanation:

1. No north-east to south-west aligned topographic barrier exists over the Machars which could be responsible for directing the ice flow into a south-south-westerly orientation.
2. No obvious ice-dispersal centre has been identified which would provide a dominant westerly flowing ice stream. The single isolated mass of Criffel to the east-north-east is not favoured as a viable source of ice dispersal over the far more dominant Lake District ice, which reportedly congested the Solway Firth during the Late Devensian (Charlesworth, 1926a).

The extent to which the Lake District ice might have encroached onto the Galloway region is indeterminate. It is believed by the author that evidence of its existence around Wigtown Bay and Kirkcudbright is depleted due to total reorganisation of sediment by later stages of ice flow, in particular by late-stage topographically constrained ice which is depicted by flow stages F and G within the Cree, Fleet and Ken valleys (Figure 5.1). This hypothesis is, however, conjectural and requires further evidence in order to certify its viability. The discovery of Lake District ice encroaching upon the existing mainland of south-west Scotland has, until now, been unreported by previous workers though requires further compelling evidence in order to affirm its pre-existence.

FLOW STAGE C

This stage denotes the infiltration of an extensive southerly-flowing ice sheet from a source area outwith the investigation area to the north of Carrick region (Figure 5.1). The orientation of the lineations which reflect stage C are strikingly compatible to previous evidence from alternative ice-flow directional records used by Charlesworth (1926a) to invoke the theory of south-south-westerly flowing Highland ice along the western seaboard. This evidence has been detailed in Section 4.2.3.5. The southerly expansion of Highland ice along the coastline of south-west Scotland is also favoured by the author in this palaeo-ice sheet dynamics model largely because of the discovery of erratics of Ailsa Craig microgranite (paisanite) specific to the island of Ailsa Craig on the western shoreline of Loch Ryan. The author cannot conceive any other physical mechanism, apart from transport by northern ice, which could be responsible for the deposition of these erratics over the Rhins of Galloway.

It has been reported in Section 4.2.3.5 that along the western seaboard of Girvan and Ballantrae district, there are only a few isolated drift lineations which relate to the Highland ice sheet (Figure 4.8). The author strongly suggests that the overlying reason for the scarcity of bedforms pertaining to flow stage C is the subsequent supersedence by lineations of later flow sets (stage D-1 and D-2) originating from local source areas to the east. Despite the evidence for interaction between competing regional and local ice streams, there appears to exist two distinct areas along the western seaboard which are devoid of bedforms representing either of the two flow patterns (Figure 4.8). It is claimed that bedrock topography is the main parameter inhibiting bedform development in these areas for reasons summarised below (a more detailed explanation is provided in Section 4.2.3.5):

1. In areas displaying a scarcity of bedforms of Highland origin, and virtually no lineations reflecting Southern Upland ice, it is visible that the predominant upland masses of Hadyard Hill (323 m), Saugh Hill (296 m), Troweir Hill (296 m) and Grey Hill (297 m) played a critical role in constraining the flow of the lateral margin of the Highland ice sheet (Figure 4.8). To the south of Ballantrae district, it is suggested that the upland areas of Beneraird (439 m), Carlock Hill (319 m) and Milljoan Hill (403 m) also dictated the course of flow stage C (Figure 4.8). An estimate of the thickness of the Highland ice sheet along the south-west Scotland coastline at this time can be deduced from the heights of these summits, which are suggested not to have been overtopped. It is important to acknowledge that this finding conflicts considerably with the Late Devensian maximum ice thickness suggested by Boulton *et al.* (1977) over this area of south-west Scotland (*c.* 1550 m), based upon the modelling of an equilibrium rigid-bed ice sheet under non-streaming ice flow conditions. However, this fact may be compensated for if consideration is given to the presence of other alternative parameters which the current British ice sheet model fails to accommodate, such as high velocity ice flow which has been identified earlier in this section. The general impact that this might have upon the mean surface elevation and morphology of the British ice sheet is fully discussed in Section 5.2.4.
2. The areas containing intensive overprinting of Highland ice bedforms by Southern Upland bedforms are in two distinct regions where westerly flowing local ice (flow stages D-1 and D-2) has not been obstructed from flowing directly into the Irish Sea Basin. It is suggested by the author that the directions of flow of stages D-1 and D-2 have been partially dictated by the positioning of the upland areas along the western margin, indicated in point 1, which have acted as a 'screen' to preserve lineations of Highland origin on the seaward margin of these topographic masses.

Critical in validating Point 2 above is the proviso that the Highland ice sheet retreated back to the north or north-west (or at least away from the seaboard) from the position which is portrayed in Figure 4.8, in order to allow the cross-cutting of lineations. It otherwise cannot be imagined how the smaller localised ice streams could have overridden the drift bedforms of the Highland ice sheet, given that it is unlikely that the local ice flow would deviate the more powerful Highland ice sheet away from its preferred flow path. The cleaving of the confluent Southern Upland and Highland ice sheets is thought to have taken place during a phase of retreat of the Highland ice sheet. However, during the period when Highland ice attained its maximum southerly position, during the Late Devensian maximum, it is expected that the Southern Upland ice streams acted as tributaries to the Highland ice, which incorporated them into a south-south-west flow over the Rhins of Galloway.

FLOW STAGES D-1 AND D-2

As briefly discussed in the context of Highland ice flow, stages D-1 and D-2 indicate a period of local ice expansion (Figure 5.1). The author has suggested that this was synchronous with the early stages of flow stage C, but for reasons presented above the majority of the drift lineations pertaining to these two ice lobes must post-date the later recession of Highland ice from the seaboard. Flow stage D-1 and D-2 both conform to a similar generalised flow direction changing from west-north-west, through west, to west-south-west at the coastline, and in part this is thought to reflect the trend of the valleys of the Stinchar and Water of Girvan. Although no evidence currently exists to confirm the synchronicity of these two flow stages, it is conceivable from the cross-cutting of drift lineations of flow stage C, that they occurred during a period sometime after the Late Devensian maximum, but one which still maintained a level of intense glacial activity throughout south-west Scotland.

A notable characteristic of the drift lineations belonging to flow stage D-2 is the apparent degree of cross-cutting. Of the three scenarios presented in Section 4.2.3.6, the author considers that the most plausible explanation for this overprinting is by the reorientation of the principal flow direction over the Girvan district by migration of the main ice-divide, which is likely to have been situated over the Loch Doon/Cairnsmore of Carsphairn area. This conforms with the apparent variation in discrete flow directions which appear to be evident throughout the history of the Late Devensian glacial period in south-west Scotland. Based upon the general trend of the generalised ice flow direction arrows in Figure 5.1, the source area responsible for the expansion of flow stage D-1 is most likely centred over Merrick and Lamachan Hill to the south, but the precise location is indeterminate.

The favoured scenario brings into question why a whole array of intermediate orientations are not preserved in this area (as opposed to a number of distinct cross-cutting flow sets), or indeed just the final orientation following readjustment of former lineations which would undoubtedly have acted as protuberances to the most recent ice flow direction. In such cases where it is necessary to account for such 'jumps' between the alignment of lineation sets within a particular area, Clark (1993) generally invokes the requirement of a rapid ice divide shift between positions of relative stability. It remains to be answered, however, whether such a sudden re-positioning of an ice divide is glaciologically plausible, given that mechanical changes within glacial systems are rarely perceived as taking place over a short time-span. Clark's (1993) alternative suggestion of a steadily migrating ice divide with some form of on/off mechanism of lineation generation is certainly more attractive, however, it is currently uncertain as to what is/are the major component(s) responsible for determining lineation production.

FLOW STAGE E

The lineations from this stage clearly superimpose the lineations of stage C over the Rhins of Galloway, and so are considered to post-date this stage (Figure 5.1). They also cross-cut those lineations relating to initial ice expansion from the Southern Uplands during flow stage A. The apparent restriction of the lineations to the western part of the peninsula is undoubtedly a consequence of later overriding by flow stage G. Only in the southern part of the peninsula do the lineations exist throughout the whole area, which further implies that flow stage G did not encroach upon these parts of the Rhins of Galloway. It is unclear why lineations pertaining to this flow stage are not identified to the north-east of the Loch Ryan Basin. Normally, under such circumstances the absence of drift bedforms could be explained by this flow event pre-dating flow stages pre-B and B, though the author is confident that flow stage E represents the most recent phase of ice flow over the western area of the Rhins of Galloway. This interpretation is based on the high concentration and spatial distribution of the lineations, as well as observed overprinting relationships.

FLOW STAGES F AND G

The location of the drift lineations which reflect these two independent flow stages are found within the major valleys of the Ken and Cree alone. Elsewhere, ice flow belonging to flow stage G is also suggested to be represented within the valleys of Fleet and Doon. These troughs all radiate out to the north and south from the central upland massif depicted in Figure 5.1. In addition, a flow set identified within the confinements of the Loch Ryan Basin to the west has been correlated with flow stage G on the basis of its glaciological compatibility. In general, flow stages F and G are confidently thought to represent topographically (valley) constrained flow of more or less independent glaciers occupying these troughs. They infer a similar generalised pattern of ice flow to the overall character of Late Devensian glacial retreat modelled by Charlesworth (1926a), and the apparent overprinting and re-modification of those elements related to flow stage F by later flow stage G provides strong evidence for the support of a readvance episode as highlighted in Sections 4.2.3.8 and 4.2.3.9. This subject, along with a more detailed discussion of the character of the deglacial period depicted by the constrained nature of ice flow within these major troughs, is addressed in Section 5.3.

5.2.3: Ice flows and the dynamics of a time-dependent ice sheet through the last glacial cycle

It has been suggested by Boulton *et al.* (2001) that palaeo-ice flows are intrinsically related to the dynamics of contemporary ice sheets, although the knowledge of whether they represent stable or transient features of an ice sheet during its evolution remains a key problem. The inability to solve this issue stems from the fact that historical observations of modern ice sheet dynamics over the last few centuries can only address changes over a short-time period, and consequently fail to reveal any time-dependent changes which may occur within the slow variability of a glacial system. In particular, Boulton *et al.* (2001) recognise that erosional or depositional palimpsests of earlier ice flows are not revealed in contemporary ice sheets, despite modern observations such as those from the west Antarctic ice sheet (Whillans *et al*, 1987) suggesting that ice streams have stagnated only recently in the context of the geological time-scale (e.g. ice stream C in the Ross Sea sector stagnated only 250 years ago). This is principally because the catchment from stagnant ice flows becomes immediately reoccupied by adjacent ice streams, and thereby covers any earlier history (Boulton *et al.*, 2001).

In contrast, regional observations of palaeo-ice sheets have been shown to provide a wealth of information which is preserved in the geological record, especially when interpreted using modern analytical methods (e.g. Punkari, 1985, 1993; Boulton and Clark, 1990; Clark, 1993; McCabe *et al.*, 1999; Boulton *et al.*, 2001) (Section 3.2). Specifically, recognition of cross-cutting patterns (e.g. Rose and Letzer, 1977; Riley, 1987; Rose, 1987; Dyke and Morris, 1988), is of valuable importance as it has been shown how these relationships provide a detailed insight into time-dependent variations within palaeo-ice sheets. This not only helps us to extend the understanding of ice stream mobility beyond that derived from modern ice sheet studies (Boulton *et al.*, 2001), but also suggests the necessity to reconsider the implications of former data used in the initial reconstructions of former ice sheets at the beginning of the twentieth century.

The reassessment of Late Devensian palaeo-ice sheet flow in south-west Scotland (Section 5.2), using remotely sensed analysis, has undoubtedly demonstrated that previous reconstructions by Geikie (1901; Figure 2.12) and Charlesworth (1926a; Figure 2.13) (Section 2.4.2) have unaccountably failed to simulate the complexity and dynamism of this former ice sheet and are therefore no longer tenable. Additionally, the general character of glaciation, originally described by Charlesworth (1926a), has fallen short of considering the time-dependent variations that occurred throughout the history of the ice sheet, as proven by this investigation (Figure 5.1). Instead, it appears that the initial views on the concept of a Southern Uplands ice sheet have failed to expand much beyond the recognition of an equilibrium steady-state ice sheet model generating radial flow to its margins, despite the fact that the geological information in this region does not lend itself to this concept. Consequently, the true character of glaciation in south-west Scotland during the Late Devensian has, until now, been over-simplified because it has not detected the transient nature of former ice flows. At the extreme, the only time-dependent variation which has been suggested (Charlesworth, 1926a, b) is that of the occurrence of ice flow readvance and margin oscillations during late-stage flow (Section 2.5.4; 5.4).

Further to this, current generalised ice flow models of the British ice sheet (e.g. Boulton *et al.*, 1977; Bowen *et al.*, 1986) appear also to have masked the major changes in ice-mass dynamics which are evident in this region. The determination of the surface shape and velocity distribution of the Late Devensian ice sheet over Britain by Boulton *et al.* (1977) has relied upon the assumption that the glacier flow field over the Southern Uplands was constant through time. However results from this investigation have shown that steady-state conditions were unlikely to have been attained (Section 2.4.1).

An account of the development history relating to continental-scale reconstructions of the Laurentide ice sheet has similarly highlighted the inability of many early investigations to recognise time-dependent changes within the glacial system. Despite the initial non-equilibrium ice sheet theory of Tyrrell (1898a), based upon three distinct centres of dispersal over the North American continent (Keewatin district, Labrador district and Patricia district, northern Ontario), the continually favoured model which existed until the 1970s was restricted to an equilibrium model of radial flow from a single-dome dispersal area over Hudson Bay (Flint, 1943). The primary reason for Flint (1943) rejecting Tyrrell's (1898a) conclusions was that they did not conform to the conceived idea of a single-domed ice sheet, which Flint (1943) based instead on topographic and climatologic data, despite field evidence contradicting the hypothesis as similarly paralleled in south-west Scotland.

The adoption of the equilibrium model in the reconstruction of the Laurentide ice sheet further exemplifies the apparent failure of many early workers at the beginning of the last century to consider a non-equilibrium dynamic ice sheet when reconstructing the glacial history of mid-latitude ice sheets. It is only through the development of concepts relating to ice flow features, till composition, distance of erratic transport, and refinements to patterns of glacioisostatic recovery that has prompted a return to the former hypothesis of a transient multi-dome ice model (e.g. Dyke, 1978; Shilts *et al.*, 1979; Andrews and Miller, 1979; Hillaire-Marcel *et al.*, 1980; Dredge, 1982; Dyke *et al.*, 1982). In particular, more recent reconstructions (e.g. Dyke *et al.*, 1982) have provided independent geological evidence that suggests a time-dependent evolution of the Laurentide ice sheet. With the application of remotely sensed analysis the notion has recently been affirmed that the Laurentide ice sheet's behaviour was indeed much more dynamic than previous views imply (Boulton and Clark, 1990).

It is therefore argued that simulated ice sheets from mid-latitude regions always appear to adopt the notion that they were never far from a steady-state, with the ice sheet only ever changing configuration at a slow rate. Results from the author's model of palaeo-ice sheet flow in south-west Scotland suggest, however, that in reality, ice sheets do display high levels of dynamism which tend to vary considerably over short timescales, thereby being far from a steady-state.

5.2.4: High velocity ice flows and their implications for ice sheet topography and margin retreat of the Irish Sea glacier

In Section 5.2.2 consideration was given to the high degree of sediment re-organisation by transient ice flows which have produced entirely new bedform orientations reflecting the most recent flow. It is considered by the author that zones in which intensive modification has taken place, throughout the Galloway lowlands, are caused by high ice velocities during sustained periods of flow. The occurrence of periods of rapid ice flow, if accountable, lends itself as an important disclosure which has failed to be included in the earlier descriptions of ice flow dynamics in south-west Scotland (cf. Geikie, 1901; Charlesworth, 1926a). This in turn appears to have led to the failure by Boulton *et al.* (1977) to consider the impact of high velocity ice flow when determining the surface configuration of the British ice sheet model (linked to the steady-state assumption discussed in Section 5.2.3). Consequently, it is argued that this must undoubtedly have had an impact on the inferences drawn from this modelling, particularly with respect to the morphology and kinetics of the Irish Sea glacier (Section 2.4.1).

The overall claim that the Southern Uplands ice sheet was subject to streaming instability leads to the implication that the mean surface elevation of the ice sheet could be significantly lower than that inferred from the rigid-bed ice sheet model (Boulton *et al.*, 1977), which does not incorporate fast ice flow. This claim is based upon the general acceptance that streaming ice has a lower surface profile than non-streaming ice, because relatively low basal friction permits it to discharge a fast ice flow at low values of basal shear stress (Boulton *et al.* 2001). The constituent allowing low basal friction within the Irish Sea Basin is favoured to be the sequences of pre-Late Devensian fine-grained, unconsolidated glacial and interglacial marine sediment which have the potential to deform (Thomas, 1985; Section 2.1.3). In agreement with the recent discussion by Boulton *et al.* (2001) with reference to the European ice sheet, the lowering of the mean surface elevation of the ice sheet over the north-east sector of the Irish Sea Basin may help to explain conflicts between ice sheet surface elevations derived from glaciological models, and those derived from models of Earth rheology which inverts sea-level data to infer mantle response to ice sheet loads (eg. Lambeck, 1996; Lambeck *et al.*, 1998). Generalised ice flow models of the British ice sheet (e.g. Boulton *et al.*, 1977) which do not include streaming suggest that the elevation of ice over the Southern Uplands during the glacial maximum was 1600 m (Figure 2.11), while the reconstruction by Lambeck (1996), based on numerical modelling, suggests a lower elevation of about 1100 m (Figure 2.30).

It is suggested that the recognition of ice streaming in south-west Scotland may, therefore, help to reconcile the differences between glacio-sedimentary models and numerical models which have been developed to explain the chronology of deglacial events in the Irish Sea Basin, with particular emphasis placed on the controversy which exists over the role of sea-level during the recession of the Irish Sea glacier (Section 2.7). Numerical models, for testing crustal response to glacial unloading, require a detailed knowledge of the ice-sheet configuration, including the thickness through time (Lambeck *et al.*, 1998). A compensation in ice thickness estimations over Ireland, northern Scotland, and Wales may therefore be necessary if streaming ice is confirmed to have existed also in these regions, as is claimed in south-west Scotland during the Late Devensian.

In order for the discrete flow stages outlined in Section 5.2.2 to have advanced into the Irish Sea Basin it is agreed with Eyles and McCabe (1989a) that the sea-level, at that time, must have been at least consistent with the present level if not lower. Although it cannot be proven, the author is of the opinion that initial advance of ice from the Southern Uplands into the Irish Sea Basin was more than likely across a terrestrial floor containing the deformable layer inferred above. These views stem primarily from the fact that the flow stages (e.g. stage A, stage B, stage post-B, stage E) would otherwise have had to enter into areas of deep water associated with the structurally-deepened troughs along the North Channel which would undoubtedly have halted the advance (Figure 2.1). It is also believed that they would have been subjected to high energy tide-dominated conditions which would have further inhibited the advance of these ice flows. Additionally, if some of these flow stages took the form of low gradient ice streams then this would again render their margins incapable of advancing into areas of deep water.

An alternative viewpoint which can be considered is that episodes of ice streaming may have become established during later deglacial stages, but may still rely upon the deforming-layer hypothesis. This has been explained by Eyles and McCabe (1989a) as being the direct result of destabilising ice dynamics at a calving front of the Irish Sea glacier margin, thereby relying upon marine inundation soon after the Late Devensian maximum (Eyles and McCabe, 1989a). Evidence for ice streaming in south-west Scotland cannot be used as unequivocal evidence to support this glaciomarine model, but nevertheless does lend some credence to the scenario of an unstable marine margin uncoupling from the underlying bed, and thereby progressively drawing down ice in the form of stream flows from source areas. The glaciomarine model also suggests that enhanced ice streaming is explained by the rapid retreat of the marine ice sheet downslope into an isostatically-depressed moat which included a substantial portion of southern Ireland, southern England, the south-western approaches and the English Channel (1989a). This event is claimed to have hastened the instability and rapid retreat of the Irish Sea glacier. Eyles and McCabe (1989a) stress the importance of zones of fast ice flow in accounting for enhanced subglacial erosion represented by the multitude of drumlin fields in regions surrounding the Irish Sea Basin.

5.3: Character and Mode of Deglaciation in South-West Scotland

5.3.1: Background

In Section 2.5.2 is was presented that the consensus view, with regard to the mode of Irish Sea ice sheet deglaciation, is that most sediment sequences found in depositional settings around the Irish Sea Basin's seaboard were produced by deposition from retreating land-based (terrestrial) glacial systems during eustatically depressed sea-levels. Sediment sequences and lithofacies associations from localised depositional environments described in Section 2.5.2 which favour this model, suggest one or more of a subglacial, glaciolacustrine or proglacial environment as being typical. It was also pointed out that some investigations have described Quaternary lithostratigraphic formations which mark a transition from one palaeo-environment to another (e.g. Macamore member (Irish Sea till) underlying the Screen member (pro-glacial outwash) in County Wexford, Ireland; Section 2.5.2.1). In account of these findings, the Irish Sea ice sheet has been pictured in Section 2.5.2 as a fully terrestrial ice margin with discrete lakes, possibly only of short duration, dammed along its margins (e.g. Thomas and Summers, 1981, 1982, 1983; Thomas *et al.*, 1985; Thomas and Kerr, 1987; Thomas, 1989). Section 2.5.2.2 revealed that the nature of the stratigraphic successions invoking a terrestrial-based ice sheet have been, for the most part, closely related to two landsystems models which characterise deposition by large terrestrial ice sheets in lowland areas: the subglacial/proglacial sediment association and landsystem model (Figure 2.14), and the supraglacial sediment association and landsystem model (Figure 2.15). These landsystems models are related to the sequence of events that account for the nature of the stratigraphic successions typified in the coastal region of County Wexford, Southern Ireland.

The alternative model of Irish Sea ice sheet deglaciation was subsequently described in Section 2.5.3 which is reported to have been derived from a number of recognised similarities between the sedimentological signature of Irish Sea Basin deposits and those drawn by workers from contemporary glaciomarine deposits of Spitsbergen and Alaska. Consequently, Section 2.5.3 focussed upon the validation of the terrestrial model which has been questioned by Eyles and McCabe (1989a) who themselves favour the reintroduction of the marine deposition concept. Instead of representing deposition at a time of eustatically depressed sea-levels, they have argued that most of the deposits of Late Devensian age in the Irish Sea Basin instead reflect glaciomarine deposition with isostatically raised sea-levels. The evidence used to support the glaciomarine model is concentrated on detailed facies and biofacies examination of drift exposures along the coastal zone of the Irish Sea Basin which have been suggested to reveal a complex record of subglacial and glaciomarine sedimentation at and beyond a tidewater margin. This evidence has subsequently been adopted to invoke that Irish Sea deglaciation was controlled primarily by high relative sea-levels as a result of flooding of a glacio-isostatically-depressed foreland basin. The sequence of events envisaged by Eyles and McCabe (1989a) have been described in Section 2.5.3 (Figure 2.16, 2.17).

Section 2.6 developed the encompassing theme associated with the glaciomarine model by presenting the transitional stratigraphic relationships which are claimed to exist between distinct depositional settings along the Irish Sea Basin. This correlation has been achieved by adopting a depositional systems approach which stresses the importance of lateral accretion on a basin-wide context during deposition within the complex glacial environment. The depositional settings which have been recognised in the Irish Sea Basin, detailed in Section 2.6.2 (Figure 2.21), are thought to define a distinct event stratigraphy that describes the sequence of deglaciation and records changes of relative sea-level caused by ice sheet loading and crustal downwarping. However, in Section 2.6.1 it was stressed that inherent in this approach is the recognition of the major depositional controls that have dictated the history of sedimentation (e.g. sea-level, tectonics, climate). In Section 2.7, a shortcoming in the glaciomarine model, which has been identified by some workers (e.g. Austin and McCarroll, 1992; Harris, 1991; McCarroll and Harris, 1992), was addressed concerning the misrepresentation of some glaciomarine deposits at high elevations which are fundamental components to the glaciomarine concept. The misinterpretation of such deposits is also believed by Lambeck (1996) to explain the differences in the relative sea-level between this glacio-sedimentary approach and that derived through numerical modelling (Section 2.7).

Despite the multitude of field investigations which have occurred along the peripheral margin of the Irish Sea Basin in eastern Ireland, west Wales and north-west England, the debate over whether or not the receding Irish Sea ice sheet was terrestrially- or marine-influenced has not been addressed along the lowland zone of south-west Scotland. On the grounds of the identification of two distinct flow stages depicting the retreat of ice towards the Southern Uplands (Section 5.2.2), the author presents a broad interpretation of the discrete ice lobes which are suggested to have existed in south-west Scotland following the period when local ice finally cleaved from the Irish Sea ice sheet and retreated onto the current onshore region (Section 5.3.2). The favoured interpretation of the deglacial dynamics is reconciled where appropriate with the geomorphology (Section 1.3) and depositional settings (Section 4.3) which have been described in Carrick and Galloway in order to allow a justifiable assessment of whether the mode of ice margin retreat over south-west Scotland was influenced by a marine or terrestrial depositional setting (Section 5.3.3). As highlighted in Section 4.3.1 the assessment of the above theme has been substantially restricted by the distinct absence of stratigraphic exposures throughout the investigation area. The foregoing interpretation and discussion are therefore derived from the available evidence which has been presented in Chapter 4, and are presented as a first approximation of deglacial events during late-stage deglaciation in south-west Scotland.

5.3.2: Character of late-stage deglaciation in south-west Scotland

In Section 5.2.2, flow stages F and G have been classified as representing two independent phases of ice flow within the confinements of some of the major valley systems in south-west Scotland. No evidence has been discovered to place a control on the age of the flow stages. However both clearly relate to a period when ice was receding towards the upland massif from the Irish Sea Basin, following detachment from the main Irish Sea ice sheet. At this late stage in the retreat of the Irish Sea ice sheet, the dynamics of ice flow over south-west Scotland were, for the first time, largely dictated by the general characteristics of the underlying rigid-bed relief. The depicted pattern portrayed by the Late Devensian palaeo-ice sheet dynamics model (Figure 5.1) further implies that there is clear evidence for considerable overprinting of an earlier topographically-constrained flow within the Ken and Cree valleys (stage F) by a reactivated ice flow (stage G) (Section 5.3.3).

Similar cross-cutting evidence within the confinements of the valleys of Fleet (Galloway region) and Doon (Carrick region) have not been observed on satellite imagery. However, the recording of topographically-confined drift bedforms within each valley, which parallel the trend of the respective valley axis, do nevertheless depict a discrete flow stage during late stages of glaciation. Whether or not the lineations in the Fleet and Doon valleys reflect similar reactivated flow of stage G or reflect the initial development of valley-constrained flow (stage F) is unclear. However, for the purpose of reconstruction the flow set within both the Fleet and Doon valley has been tentatively assigned to the last phase of ice advance in south-west Scotland (stage G), based upon the correlation by Charlesworth (1926a) of morphological evidence in these valleys with what are claimed to be similar features in the Ken and Cree valleys to the south (Section 5.2.3). This allocation may require a degree of revision if further evidence is found in the future to dispute this.

It is postulated that one or more of the flow sets assigned to flow stage D-1, and likewise stage D-2, may be compatible with late-stage deglacial flow represented by stage F and G. However as mentioned in Section 5.2.2, some of the flow sets assigned to these two stages only in part reflect the general alignment of the valleys in these areas (Stinchar valley (D-1) and Water of Girvan valley (D-2)), and therefore are not considered to be completely constrained by the topography. Nevertheless, independent deglacial flow along these two valleys is probably likely, and was inferred by Charlesworth (1926a) on the grounds of observed moraines spanning the valley floor near Colmonell. Evidence has not, however, been found by the author to support this earlier observation. Finally, the ice flow represented in Loch Ryan has also been correlated with the late-stage reactivation of ice flow represented by flow stage G, but is distinct from the others as it is linked to the reactivation of Highland ice. The implications of flow stage G are addressed in Section 5.4.

In order to promote the initial valley containment of flow stage F the author favours a scenario whereby the retreating Irish Sea ice sheet cleaved within the northern sector of the Irish Sea Basin allowing the separate recession of ice towards the ice dispersal centres situated in the Sperrin mountains, Northern Ireland; the Cumbrian mountains, north-west England; and the Southern Uplands, south-west Scotland. On the basis of the discussion in Section 5.2, the

Highland ice sheet is suggested to have disconnected from local Southern Upland ice somewhere off the western coast of the present mainland allowing the local ice to retreat back towards the source area in the upland massif to the east.

In consolidating the discussion above, the constrained configuration of drift bedforms of flow stage F and G within the major valleys of the investigation area, represented by the palaeo-ice sheet model (Figure 5.1), support the suggestion of discrete valley glaciers during late-stage deglaciation. The development of this phase is claimed by the author to have taken place following the detachment of Southern Upland ice from the main ice sheet in the Irish Sea Basin, which then was followed by gradual recession onto the current onshore area alongside further fragmentation creating independent ice lobes. On the basis of this model of deglaciation the following ice lobes are suggested to have existed in south-west Scotland following this episode: (1) Ken ice lobe; (2) Fleet ice lobe; (3) Cree ice lobe; and (4) Doon ice lobe (Figure 5.1). Other independent glaciers are tentatively suggested to have developed within the valleys of the Stinchar and Water of Girvan, however they cannot be confirmed on the basis of the current available evidence.

5.3.3: Mode of deglaciation of the Irish Sea ice sheet onto south-west Scotland

In providing an assessment of the mode of ice margin retreat onto south-west Scotland the evidence which can help to define whether the margin was terrestrially-or marine-influenced is presented by those depocentres which are situated within the vicinity of the current south-west Scotland coastline. Of the field sites reported in Section 4.3 the only stratigraphic sequence falling into this category, and which also presents evidence to entertain both a glacioterrestrial and glaciomarine scenario, is the grounding-line fan complex at Arbrack (Section 4.3.2.4). The development of this depositional setting is also closely associated with the suggested recession of Lake District ice across the Scottish mainland, representing an ice-marginal stand-still position, and so forms an important component of the record left by the Late Devensian ice sheet. Also significantly related to this depocentre, throughout the southern district of Whithorn, are a number of other similar sedimentological observations and morphological features which also depict subaqueous depositional environments, such as flat-topped outwash fans of which relics occur on the coast of Cairndoon [NX 373392], at Burrow Head [NX 451342] and at Port Allen [NX 477409] (Barnes, 1989; Section 1.3.3.2). Evidence reported by Barnes (1989) from the Burrow Head deposits also invokes a similar southerly direction of deposition currents, as recorded at Arbrack. Small lacustrine environments are reported in Section 1.3.3.2 which occur at Whithorn Moss [NX 409422] and Glasserton [NX 424390], close to the depocentre at Arbrack. Finally, the preservation of longitudinal P-forms close to the southern promontory of the Machars region may be of importance with respect to the deglacial history of the Southern Upland ice sheet over this district.

There are two basic scenarios which can account for the evidence observed at Arbrack. The first is that the development of the grounding-line fan is associated with episodic retreat of flow stage post-B (Lake District ice), in a north-easterly direction, whose marine-influenced ice-margin equilibrated, allowing meltwater and sediment to be discharged from subglacial and/or englacial portals into the subaqueous environment. This formed the basis for allowing the development, in a south-westerly direction, of the grounding-line fan complex by a combination of gravitational processes and suspension settling. High magnitude discharge events, creating buoyant plumes of suspended sediment carried the fine-grained sediment away to lateral, distal positions where it eventually settled, leaving a concentration of coarse-grained sediment near to the fan apex. Sea-level at this time was higher than the height of the surface of the fan apex. Based on available evidence, this would have important consequences for sea-level history in this area as the minimum height which the sea-level would have had to attain is 40 m above present sea-level. The alternative hypothesis, which has already been developed in Section 4.3.2.4, is that the grounding-line fan complex, situated within an elongate depression 30 m above sea-level and surrounded for the most part by rises in the topography of up to 70 m above sea-level, merely represents a prograding fan complex within a glaciolacustrine setting, having been temporarily dammed by ice still situated in the Solway Firth to the south.

Clearly the consequences of the alternative glaciolacustrine hypothesis fail to explain why much of the evidence is indicative of a glaciomarine grounding-line fan. However, this does not necessarily preclude the glaciolacustrine scenario, as the evidence for a glaciomarine setting can also be explained in the context of a lacustrine subaqueous environment. The evidence which could be accepted as indicative of a glaciomarine setting, such as the distinct absence of fine-grained sediment and high concentrations of coarsest-grained material at the ice-contact point, has been accounted for within a glaciolacustrine environment in Section 4.3.2.4. The presence of comminuted shells, although anticipated upon discovery to fix a minimum deglacial date for marine-influenced ice recession onto this part of the Scottish mainland, yield a date of 35, 800 ± 770 years B.P., and therefore present too great an age to be related to an episode of marine transgression during late-stage deglaciation.

An assessment of the relation of the P-forms to the ice-grounded fan to the north also led to the initial assumption that the northern margin of these features, marked by a south to north transition from bedrock troughs to surface drift, represented a marine washing limit which tied in with the elevation of the depocentre. However a reappraisal of this conjecture, upon analysis of aerial photographs, revealed that the proposed washing limit did not maintain a standard height above present sea-level, and therefore failed to meet the criteria of a marine limit.

The necessity of an unusually high sea-level in order to account for the development of the ice-grounded fan within a marine setting does, however, favour the acceptance of a glaciolacustrine setting. Clearly, the only way in which the possible marine limit can be attained is by maintaining a glacio-isostatically depressed shoreline, however whether this is likely at such a late stage in deglaciation of the Irish Sea ice sheet is doubtful. Indeed, the sea-level observations documented in Section 1.3.3 are substantially lower than the required 40 m above sea-level (*c*. 5 m above sea-level around the district of Whithorn), and therefore it has always been generally assumed that sea-level, during late deglacial times (*c* 14 000 years B.P.) must have been very similar, if not below, the current level. Furthermore, the recording by Bishop and Coope (1977) of a mean sea-level at -3 to -4 m at 12 290 ± 250 years B.P. requires a substantial fall in relative sea-level to accommodate this evidence.

Overall, although it may be argued that the conflict in information does not necessarily require the rejection of the glaciomarine hypothesis, it appears that the glaciolacustrine model accommodates all of the current data (lacustrine environments reported in Section 1.3.3.2; sea-level predictions, and sedimentological and stratigraphic observations) to a more satisfactory extent. With regard to the mode of Irish Sea deglaciation it is concluded by the author, as a first approximation, that the ice receding onto the Scottish mainland during the late-stage deglacial period had a fully terrestrial margin.

5.4: Ice Readvance in South-West Scotland

5.4.1: Background

In Section 2.5.4 the theme of Irish Sea deglaciation was further developed to incorporate the status of marginal ice readvances which have been claimed by various workers to have arisen over the northern half of the Irish Sea Basin on a wide scale. Those who dispute the readvance hypothesis either argue for short-distance, endemic oscillations of the ice margin as a means of explaining the stratigraphic evidence which supports these claims, or dismiss the need to invoke readvance events at all. Despite the many documented readvances and/or oscillations which are reputed to have occurred in the Irish Sea Basin, the main readvance episodes over the lowlands of south-west Scotland which have been proposed are the Scottish readvance (from southern Scotland onto the Cumbrian lowland) (Section 2.5.4.1), the Highland readvance (into

Loch Ryan and onto the northern base of the Southern Uplands) (Section 2.5.4.3), and the Southern Upland readvance (along the main valleys in south-west Scotland) (Section 2.5.4.4).

From an assimilation of the criteria used to propose readvance phases throughout Section 2.5.4, the two main forms of diagnostic evidence which appear to have mainly been adopted to support these conjectures are (1) ice-marginal landforms, e.g. moraines (on occasions marking the limits of drumlin belts) (cf. Charlesworth, 1926a, b, 1939; Synge, 1952; Mitchell, 1960, 1963; Gresswell, 1967) and (2) vertical lithofacies transitions (i.e. advance–retreat–advance cycles) from fluvioglacial deposits (kames, eskers, proglacial sandur, deltaic sands and gravels, glaciolacustrine silts and clays) to till (e.g. Scottish readvance). These are claimed to be indicative of the overriding of proglacial deposits by reactivation of deglacial flow (especially in Cumbria) (cf. Trotter, 1929; Hollingworth, 1931; Trotter and Hollingworth, 1932; Huddart *et al.*, 1977; Huddart, 1991, 1994). Attempts have also been made to relate such evidence in Cumbria to broadly correlative readvance limits in Ireland and the Isle of Man (e.g. 'north-east Ireland – Isle of Man – Cumberland' moraine; Charlesworth, 1939). Likewise in south-west Scotland, Charlesworth (1926b) has correlated readvance evidence across much of the northern head of the Southern Uplands, thereby inferring the Lammermuir–Stranraer moraine, marking the postulated readvance of combined Highland and Southern Upland ice. In Section 2.5.4.1 it has been described how recent recognition of evidence for glaciotectonic deformation and glacial overriding to the south of Whitehaven district, Cumbria has renewed interest over the concept of a Scottish readvance in particular.

It has been highlighted throughout Section 2.5, however, that much of the evidence used to invoke glacial readvances during deglaciation of the Irish Sea ice sheet is unreliable (cf. Richmond, 1959; Flint, 1971). Likewise, the fundamental sedimentological criteria described to invoke readvances is claimed by some workers not to automatically indicate late-stage reactivation of ice flow (cf. Boulton, 1972), particularly in light of the recognition of multiple till sequences arising from a single glacial episode (cf. Boulton, 1977). A detailed sedimentological investigation by Huddart (1999; Section 2.5.4.4) showed how the evidence invoked by Charlesworth (1926b) for a combined readvance of Highland and Southern Upland ice could be interpreted as supraglacial trough fills associated with the decay of ice-cored ridges in a marginal environment.

The discussion on the status of readvance episodes in south-west Scotland begins by revisiting the theme outlined in Section 5.3.2 where the author claimed that flow stage G represented the reactivation of ice flow during deglaciation. Section 5.4.3 then introduces to the discussion the comprehensive findings from the Loch Ryan Basin which are believed to provide particularly important evidence to support what the author has termed the Highland–Southern Upland readvance, for reasons detailed in Section 4.2.4. This commentary incorporates the evidence used by Charlesworth (1926b) to originally propose this readvance event (Section 4.2.4.1, 4.2.4.2) and relates this to the recent reappraisal of the landform distribution as illustrated in the glacial geomorphology map in Figure 4.12 (Section 4.2.4.3). Section 5.4.4 outlines the proposal of a new model for a palaeo-surge event along the Loch Ryan Basin, representing the final passage of ice along this trough. Additional information is sought from the geomorphology detailed in Section 1.3 and particular palaeo-depositional settings detailed in Section 4.3. Finally, Section 5.4.5 assesses the mechanism which may be responsible for driving readvance episodes within the Irish Sea Basin sheet during deglaciation.

5.4.2: Implications for a Southern Upland readvance

Evidence to suggest the overprinting of flow stage F by flow stage G, within the confinements of the Ken and Cree valleys, has been explained in Sections 4.2.3.9 and 5.3.2 as the result of the reactivation of topographically constrained deglacial ice flow. The author suggests that Stage G can be equated with Charlesworth's Southern Upland readvance whereby local ice was pictured within the main valleys radiating out from what the recent palaeo-ice flow model now claims was a bogus north-east to south-west ice-divide situated along the main watershed of the Southern Uplands. Disparate from the readvance model proposed by Charlesworth (1926a) is that the congruous alignment of the most recent flow sets of stage G, throughout the entire length of the Cree, Ken and Doon valleys, argues for only one readvance episode of Southern Upland ice. This differs from the early suggestion by Charlesworth (1926a) of at least two readvances, based mainly upon the positioning of moraines along the Cree valley (Section 4.2.3.9; Figure 2.13). It can also be inferred from Figure 5.1 that the readvance of flow stage G must have occurred at a time after the retreat of Lake District ice (flow stage post-B) from Wigtown Bay, and therefore post-dates the ice-grounded fan at Arbrack (Section 4.3.2.4, 5.3.3). This conjecture is based upon the evidence of substantial pre-lineation deformation of bedforms of flow stage post-B by the Cree ice lobe, along the eastern margin of the Machars.

It is indicated from the palaeo-ice flow model that the readvance flow along the Ken, Fleet and Cree valleys (stage G), to the south of the Southern Uplands, entered into the Solway Firth Basin, although to what southerly position is indeterminate. From the pattern of convergence displayed by the flow sets depicting readvance flow out of the Ken, Fleet and Cree valleys, it is hypothetically suggested that the ice flows coalesced in the area of Wigtown Bay, prior to final recession within their independent valleys. Any suggestion that the magnitude of this readvance phase relates flow stage G to the concept of a Scottish readvance is disputed because of the valley-containment of the independent readvance flows. Although the pattern of ice flow over south-west Scotland during the proposed Scottish readvance has not been clearly documented, it is assumed that this would have been represented by unconstrained flow, in order to transgress as far as the various postulated limits in north-west England (Figure 2.19).

5.4.3: Implications for a Highland readvance

Within the Loch Ryan Basin the cross-cutting of flow stages A and C, and stage E representing the last unconstrained flow (Section 5.2.2), by a set of lineations correlated with flow stage G argues strongly for a reactivation of ice into, and along, the Loch Ryan Basin. The distribution of the lineations assigned to this phase (Figure 4.4, 4.5j) additionally invokes the transgression of this flow event over the isthmus connecting the Rhins of Galloway to the Glenluce district at least as far as the northern head of Luce Bay, but to what position is indeterminate. The assignment of this flow stage to a readvance episode is substantially based upon the fact that in order for this discrete flow to have entered into the confinements of the Loch Ryan Basin, unconstrained flow must have receded to the north or north-west, as implied by the palaeo-ice flow model (Figure 5.1). Correspondingly it is argued that flow stage G, along Loch Ryan, represents the readvance of Highland ice which, prior to the triggering of this event, was situated somewhere currently offshore from the northern promontory of the Rhins peninsula and the seaboard of Ballantrae district. Evidence to corroborate this hypothesis has been found in the form of Ailsa Craig microgranite erratics on the western shoreline of Loch Ryan and within the massive diamict reported at McCulloch's Point (Section 4.3.2.1). Furthermore, the classification of the massive diamict from the two sites along the western shore of Loch Ryan suggest the deposition of a far-travelled glacitectonite, which also contains a macrofabric matching the inferred flow direction (Section 4.3.2.1). On the basis of this evidence the author favours the proposal by Charlesworth (1926b) of a Highland readvance, principally based on the relationship between drift bedforms in this region. Thus, the lineations depicting this flow event have been tentatively correlated with the Southern Upland readvance phase described above (flow stage G) as synchronised by Charlesworth.

The primary evidence used by Charlesworth (1926b) to identify the readvance of Highland ice within the Loch Ryan Basin itself was based on a sediment-landform assemblage which was thought to reflect an ice-marginal kame-moraine, situated at the head of Loch Ryan and demarcating the maximum extent of Highland readvance ice. The morphological description and characteristics have been described in Sections 2.5.4.3 and 4.2.4.1. From the assessment of the magnitude of ice readvance discussed above, however, it is argued that this feature does not represent a maximum position. This can be additionally verified by the presence of the extensive lateral moraine, described in Section 4.2.4.3, which appears to extend in a south-easterly direction beyond the inferred position of the terminal kame-moraine. Charlesworth (1926b) does identify this moraine, but fails to distinguish its lateral position, instead claiming that it is a continuous feature across the width of the basin. However, the characteristics of the geomorphology within the isthmus clearly reflect a kame-and-kettle topography (Figure 4.12), and so portray an episode of ice stagnation at a marginal position. This interpretation agrees with the suggestion of Huddart (1999), though still accommodates the Highland readvance episode.

Within the Loch Ryan Basin, there has also been recorded a number of discrete sources of evidence based on morphological forms and sedimentological characteristics which suggest that the advance and retreat phase of the Highland readvance was not as straight forward as portrayed above. The overriding theme, particularly invoked by certain landforms, is for a independent phase of rapid ice flow, which was disconnected from the initial readvance of Highland ice. The impact of this evidence is discussed in detail below (Section 5.4.4).

5.4.4: Implications for a Loch Ryan palaeosurge

5.4.4.1: Background

It has been proposed that during the Quaternary many of the large ice sheets moved over a deforming bed which resulted in fast ice flow (e.g. Alley, 1991), but neither the temporal nor spatial scale of this behaviour is known. Recent studies of contemporary glaciers have also suggested that fast ice flow may indeed be a very common feature. Of the three different types of fast ice flow known to exist (tidewater glaciers, deforming-bed glaciers, and surging glaciers), it is the term 'surging glacier' which is used to relate to a small-scale, short-term velocity increase within a topographically-constrained ice flow. There have been many theories to account for this phenomenon, however recent research can be summarised into two models, based on either changes in the subglacial drainage system (Kamb *et al.*, 1985), or the subglacial deforming bed (Clarke *et al.*, 1984).

Numerous authors have proposed surging of Pleistocene ice sheets based on distinctive geomorphological and sedimentological criteria. Clayton *et al* (1985) suggested that the most characteristic feature of surging margins of the south-western part of the Laurentide ice sheet was extensive tracts of hummocky moraine. Dredge and Cowan (1989) suggested that inset fluting fields terminating at major moraines provided evidence of palaeosurges on the Canadian Shield. Other diagnostic landforms associated with contemporary surging glaciers have also been identified, including thrust moraines, crevasse-fill ridges, flutings, hummocky moraine, sediment gravity-flow, deposits from supraglacial melt-out tills and concertina eskers (Sharp, 1985a, b; Croot, 1988a, b; Knudsen, 1995).

Although many individual landforms have been identified as being associated with surging glaciers, almost all the features have also been similarly associated with non-surging glaciers (with the exception of concertina eskers). In particular, push moraines are almost ubiquitous to the proglacial environment and, on their own, are not indicative of surging glaciers. Therefore, such studies have failed to identify suites of landforms and sediments which can provide unequivocal evidence for palaeosurging. Evans and Rea (1999) have recently presented evidence from contemporary surging glaciers in Iceland, Svalbard, U.S.A. and Canada suggesting that there is no single preservable landform or stratigraphy which can be used to characterise the surging behaviour of a glacier. Using these observations, Evans and Rea (1999) have subsequently constructed a landsystems model (Section 3.3.1.4) which can be used for the identification of former surging-glacier margins. The application of this model in the assessment of the landform assemblage within the Loch Ryan Basin, south-west Scotland is presented below (Section 5.4.4.3).

5.4.4.2: Landsystems model for surging glaciers

Based upon a combination of observations from contemporary surging glacier margins, specifically Brúarjökull and Eyjabakkajökull, Iceland (cf. Evans *et al.*, 1999; Evans and Rea, 1999) and the published literature, a landsystems model has been developed which includes the likely geomorphic and sedimentological signature for glacier surging (Figure 5.2). Within the model, Evans *et al.* (1999) present a geomorphology which is arranged in three overlapping zones: an outer zone (A) of thrust-block moraines grading up-ice into patchy hummocky moraine (zone B), and then into flutings, crevasse-fill ridges and concertina eskers with an area of pitted, channelled and/or hummocky outwash and occasional overridden thrust and push moraines (zone C). Further elaboration of the landsystems model, based upon the signature of a single surging event described by Evans and Rea (in press) (cf. Evans *et al.*, 1999; Evans and Rea, 1999), is presented below.

The outer zone -A- reflects the limit of the surging event and is composed of weakly consolidated pre-surge sediments, proglacially thrust or pushed by rapid ice advance. Structurally and sedimentologically these thrust block and push moraines comprise interbedded thrust slices of folded and sheared glacitectonites (Figure 5.2). The development of major thrust block moraines (e.g. composite ridges and hill-hole pairs) is thought to be restricted to

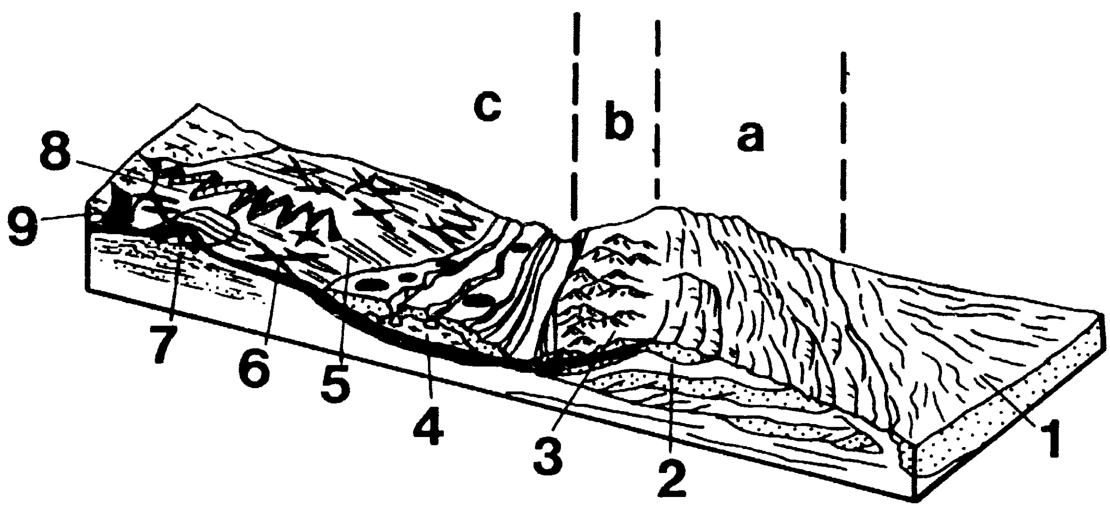

Figure 5.2 A landsystems model for surging-glacier margins: a) outer zone of proglacially thrust pre-surge sediment; b) zone of weakly developed hummocky moraine; c) zone of flutings, crevasse-fill ridges and concertina eskers; 1) proglacial outwash fan; 2) thrust-block moraine; 3) hummocky moraine; 4) crevasse-fill ridge; 5) concertina esker; 6) fluting; 7) glacier. (From Evans and Rea, in press).

topographic depressions that are large enough to collect sufficient sediment during quiescent phases.

The intermediate zone -B- is characterised by poorly-developed or patchy hummocky moraine located on the down-glacier sides of topographic depressions, and often draped on the ice-proximal slopes of the thrust-block and push moraines of zone A (Figure 5.2). The hummocks are thought to represent the combined product of (1) supraglacial melt-out, (2) flowage of debris derived from the incorporation of stagnant ice into the glacier; and (3) material transported from the glacier bed along shear planes and via crevasses during and immediately after the surge event. Kame-and-kettle topography can occur if proglacial outwash is deposited over the stagnating glacier snout (e.g. Brúarjökull; Evans and Rea, 1999).

The inner zone -C- consists of subglacial deformation tills and long, low amplitude flutings, produced by subsole deformation during the surging event, and crevasse-fill ridges, denoting the filling of basal crevasses during the final stage of the surge, prior to the quiescent phase (Figure 5.2). The formation of concertina eskers has also been attributed to this zone where they are draped over the flutings and crevasse-fill ridges, although the preservation of these features is believed to be poor. Discontinuous gravel mounds within palaeoenvironments may represent concertina eskers.

5.4.4.3: Application of the surging glacier landsystem model

Evans and Rea (in press) suggest that the surging glacier landsystem outlined above can be utilised when identifying possible palaeosurges in the Quaternary glacial record, based upon the principle that reconstructions of palaeo-ice dynamics rely heavily on suitable analogues from contemporary glacierised landscapes where form is fundamentally linked to process. The landsystems model is applied here in the assessment of a palaeosurge event within the Loch Ryan Basin, south-west Scotland.

The mapped area (Figure 4.12) comprises the following landform/sediment assemblages which exhibit the characteristics of the surging-glacier landsystems model (Figure 5.2; Section 4.2.4.3):

1. Megaflutes, of the order of 1800 m in length, which are superimposed on top of the lateral moraine and document the passage of fast-flowing ice from the north-north-west. Although flutings have been traditionally related to ice streaming (Jones, 1982), their association in

this context with other landforms of the surging glacier landsystem links them with a palaeosurge event.

2. Large-relief ridges of suspected proglacially thrusted pre-surge sediment (cf. thrust-block moraines) constructed at the margin of ice moving from the north-north-west.
3. Extensive tracts of kame-and-kettle topography occupying the present isthmus.
4. Transverse (crevasse-fill?) ridges to the east of Loch Ryan.
5. Possible glaciotectonic depressions (e.g. Black Loch and White Loch).

It is evident that the landform assemblages shown in Figure 4.12 exhibit many of the characteristics of the surging-glacier landsystem model. As the origin of some of the landforms is conjectural (e.g. crevasse-fill ridges and thrust-block moraines), the author admits that the evidence within the Loch Ryan Basin for a palaeosurge event is not unequivocal, though strongly suggestive. Additional information which strongly supports a palaeosurge claim can be sought from the fabric signature recorded in the deposits at McCulloch's Point and Hillhead of Craichmore (Section 4.3.2.1). It was noted from these two sites that although the dip direction remained fairly similar in the upper and lower sections of the diamict, the Schmidt nets for both sites displayed a shift from the north-east to north quadrants upsection. This is believed to clearly support a change in the palaeo-flow direction along the Loch Ryan Basin and is suggested to reflect a readjustment in the flow between the initial Highland readvance (Section 5.4.3) and the later palaeosurge.

In light of the discussion above this investigation has revealed what is believed to be convincing evidence to support a surge of Highland ice along the Loch Ryan Basin following an earlier readvance phase. The position from which the surge took place is indeterminate though is suspected to have been within the confinements of the Loch Ryan Basin. The palaeosurge clearly overrides the lateral moraine on the west side of Loch Ryan (as indicated by the superimposed lineations), although no evidence has been discovered which can suggest a southerly limit of the advance. Following the final rapid advance of Highland ice, the scenario is favoured whereby there was gradual recession of the ice out of Loch Ryan. The kame-and-kettle topography situated over the Stranraer isthmus therefore is likely to be related to stagnation of this final surge event, rather than to the previous retreat of the initial readvance phase. As described in Section 4.3.2.3, the formation of Glacial Lake Aird is believed to have formed beyond the margin of the retreating ice during a period of stillstand. The hypothesis described to account for the feeding of subaerial meltwater from the south-east into the north-westerly prograding Gilbert-type delta within the lake is maintained here (Section 4.3.2.3). This envisages the development of a laterally extensive palaeosurge piedmont lobe over the lowland plain to the south of the confined basin, allowing subaerial lateral runoff in a northerly direction towards Glacial Lake Aird.

5.4.5: Ice sheet variability during deglaciation

It has recently been suggested by McCabe and Clark (1998) that terrestrial records of millennial-scale ice sheet oscillations should be present in what they perceive to have been a climatically-sensitive north-eastern Atlantic region during the last deglaciation. From the results of an investigation carried out on the north-western margin of the Irish Sea Basin, McCabe and Clark (1998) have identified a deglacial readvance across eastern Ireland based on lithostratigraphic evidence which has been constrained by AMS ^{14}C dating of benthonic foraminifera *Elphidium clavatum* suggesting ice marginal readvance around 16 400 years B.P. It is claimed by McCabe and Clark (1998) that this episode corresponds with Heinrich (ice-rafting) event 1 (H1) (Heinrich, 1988), being preceded in the northern Irish Sea Basin by a short interstadial period, and subsequent to a major readvance experienced rapid retreat. This sequence of events is reported to closely match the records from the southern margin of the Laurentide ice sheet which also suggest that following a major retreat ending at 15.0-15 ^{14}C kyr B.P., the margin readvanced and then retreated rapidly during or near the termination of H1 (Hansel and Johnson, 1992; P.U. Clark, 1994).

Although this investigation has been unable to place a constraint on the age of the readvances discussed above, it is argued that there was at least one major phase of readvance in south-west Scotland which followed a general deglacial pattern as discussed in Section 5.3.2. The variability in the Southern Upland ice sheet record appears to suggest similar millennial-scale changes as identified for other ice sheets in the North Atlantic region, however the phasing relations between the British ice sheet and other ice sheets has not been widely established as of yet (cf. McCabe, 1996; McCabe and Clark, 1998; McCabe *et al.*, 1999). It may therefore be possible that the pattern of ice sheet variability over south-west Scotland was also consistent with the pattern of ice marginal movement displayed by some of the ice sheets in the North Atlantic region. It is suggested that the instability of the Laurentide ice sheet, by initiating H1, could be responsible for reversing the initial deglacial phase over south-west Scotland and instigating the Highland–Southern Upland readvance episode. However, whether or not the North Atlantic Heinrich events were in fact triggered by the behaviour of the Laurentide ice sheet is now being questioned (Grousset *et al.*, 2000). Evidence discovered by Grousset *et al.* (2000) in the Bay of Biscay and on the southern Portugal margin appears to suggest that the typical Heinrich layers observed in ocean cores were actually initiated by ice-rafted detritus of

European origin. Thus the initial readvance of the Laurentide ice sheet may in fact have responded to a mechanism created by events in the European ice sheet.

5.5: Sea-Level History

In Section 2.7, the complex pattern of sea-level change around the British Isles was described, being qualitatively understood in terms of two components: crustal rebound in response to the melting of the former local ice sheet, and the rise in sea-level produced by the melting of the more voluminous ice sheets over North America and northern Europe (Lambeck, 1993a). Within the Irish Sea Basin, a considerable body of evidence has been discovered which quantifies the positions of past sea-levels both above and below present level. Emergence curves, representing the history of sea-level at specific sites since deglaciation, have been constructed and clearly show the spatial variability in sea-level response which has been recognised at these individual localities.(Figure 2.28). The construction of such curves relies upon the radiocarbon dates obtained from organic material that can be ascribed to former sea-levels.

Unfortunately dateable material was particularly sparse in the investigation area and only two exposures were discovered which yielded dateable samples. One site (Arbrack) has been described in Section 4.3.2.3 where a complete bivalve was sampled from a grounding-line fan but was discovered, following AMS dating, to be of Middle Devensian age (35 800 ± 770 years B.P (AA-33829)) thereby indicating from its stratigraphic position (*c*. 40 m above sea-level) to have been previously reworked. The other site (Aird), whose depositional setting is outlined in Section 4.3.2.3, provided a complete mollusc which was sampled from a core at a depth of approximately 6 m above present sea-level. A date of 6 075 ± 55 years B.P. (AA-33830) was obtained by AMS dating which correspondingly provided a minimum sea-level height for this time, assuming the mollusc was discovered in its former growth position.

Within the context of the previously known sea-level history around the seaboard of south-west Scotland, this Middle Holocene date closely supports the height inferred by Lambeck's sea-level curve at 6 000 years B.P. for the west Solway Firth (Wigtown Bay and Luce Bay).

CHAPTER SIX

Conclusions and future research

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6.1: Conclusions

The conclusions of this investigation can be summarised as follows:

- **A reassessment of Late Devensian palaeo-ice sheet flow in south-west Scotland, based upon remotely sensed analysis, has demonstrated that past reconstructions have not simulated the complexity and dynamism of this former ice sheet.**

A model of palaeo-ice sheet dynamics representing ten major ice flow stages has been reconstructed by reference to bedform alignment, location, morphological attributes and cross-cutting relationships. This confirms the recent conjecture that bedform relationships are a powerful tool in mid-latitude ice mass reconstruction (cf. Punkari, 1985, 1993; Boulton and Clark, 1990; Clark, 1993; McCabe *et al.*, 1999; Boulton *et al.*, 2001). Unique to this model is the portrayal of time-dependent changes which have occurred within the slow variability of the glacial system. These flow stages suggest that ice flows, emanating from ice divides or dispersal centres in the Southern Upland massif, were transient features of the ice sheet, and therefore compliment contemporary studies of ice stream flows in Antarctica. This argues against the conventional equilibrium model of steady-state radial flow from a single ice divide situated along the main drainage divide of the Southern Uplands (cf. Geikie, 1901; Charlesworth, 1926a). Overall, traditional simulations of ice sheets in mid-latitude regions have failed to consider the temporal changes that an ice sheet is likely to experience during its history, and appear to adopt the notion that they were never far from a steady-state. However in reality, ice sheets do display high levels of dynamism, thereby being far from a steady-state.

- **A revised Late Devensian palaeo-ice flow model describes ten major transient flow stages which reflect the time-dependent, dynamic evolution of the Southern Upland ice sheet.**

Flow Stage A represents the initial Late Devensian expansion of ice from a source area situated over the NW-SE upland crestline of the Merrick range in a south-westerly direction over the Glenluce lowlands and central Rhins of Galloway. *Flow Stage Pre-B* of southerly flow direction overprints, and hence post-dates, the initial expansion of ice from the Southern Uplands. Extensive pre-lineation deformation of stage A flow sets possibly suggests dynamic, high ice flow velocity. Ice-divide migration, with some form of on/off mechanism, is thought to be responsible for the realignment of this flow event. *Flow Stage B* reflects a predominant SW flow from the upland masses of Merrick, Lamachan Hill and Cairnsmore of Fleet. A further shift in the ice divide is thought to be responsible for this flow direction, and the return to the position of stage A is favoured. *Flow stage Post-B,* representing the coalescence of Lake District ice with stage B flow, is concluded in a separate point below. *Flow Stage C* presents an alignment of lineations which support Charlesworth's (1926a) original conjecture of southerly flowing ice of Highland origin, and is further confirmed by observed Ailsa Craig microgranite erratics in the Loch Ryan Basin. An interchange of dominant flow between Highland and Southern Upland ice, and the positioning of bedrock topographic obstructions, are responsible for a) the spatial distribution of lineations pertaining to these flows, and b) the confinement of areas displaying overprinting relationships along the western seaboard. *Flow Stage D-1* depicts former flow changing from WNW, through W, to WSW at the coastline, which in part reflects the trend of the Stinchar valley. This flow stage is suggested to have been synchronous with *Flow Stage D-2* which itself conforms to a similar flow direction, in part dictated by the Water of Girvan valley. *Flow Stage E,* indicating SW ice flow over the Rhins peninsula, superimposes flow sets of stage C and A, thereby post-dating them. This flow stage represents the final unconstrained flow over this area. *Flow Stage F* represents the initial cleaving of ice from the main Irish Sea ice sheet within the Irish Sea Basin, and topographically-constrained recession back up into the major valley systems towards the upland massif. *Flow Set G,* confined within the major valleys, overprints flow stage F in the Ken and Cree valley and is assigned to the Southern readvance episode of Charlesworth (1926a) along the valleys radiating out from the Southern Uplands.

- **Spatial bedform configuration over the central Machars reveals a convergence zone between Southern Upland ice and Lake District ice.**

A clear convergence of lineations, developed from south-west Southern Upland ice flow (Flow Stage B), with a set of bedforms favouring flow from a Lake District source area (Flow Stage Post-B) documents unreported encroachment of Lake District ice onto this part of south-west Scotland. The divergence of Lake District ice into the maintained local flow direction suggests the latter to have been more highly sustained. This is thought to be directly related to the comparative closeness of its ice dispersal region. The extent to which Lake District ice encroached upon the Solway Firth coastline to the north-east and east of the convergence zone is indeterminate. No other explanation currently exists to explain the relationship between these two flow stages.

- **Cross-cutting zones throughout the lowlands of south-west Scotland, depicting intensive periods of former pre-lineation deformation, are suggested to indicate high ice velocities during sustained periods of flow.**

If ice flow velocity is the dominant factor responsible for the degree of bedform modification, it is believed that this may have a significant impact on previously modelled ice sheet surface elevation, ice surface morphology and internal ice kinetics which have been adopted in the reconstruction of the British ice sheet (Boulton *et al.*, 1977). This is based on the principle that most regional ice sheet models have applied a rigid-bed analogue, and therefore do not entertain the concept of ice streaming over a deformable layer. The general acceptance that streaming ice maintains a lower surface profile than non-streaming ice therefore implies an exaggeration of surface elevation, which correspondingly impacts on other components of the glacial system. The recognition of ice streaming in south-west Scotland is thought to perhaps provide a solution, at least in part, to the current conflict between the glacial geomorphic and numerical models.

- **During late-stage deglaciation of the Irish Sea ice sheet, the dynamics of ice flow in south-west Scotland were, for the first time, dictated by the characteristics of the underlying rigid-bed relief.**

The confinement of congruous bedforms within the major valleys reflect topographically constrained ice flow during deglaciation. The development of these independent valley glaciers (Ken ice lobe, Fleet ice lobe, Cree ice lobe and the Doon ice lobe) most probably resulted from the cleaving of Irish Sea ice sheet ice within the northern Irish Sea Basin and Solway Firth Basin, followed by discrete recession. Other independent glaciers are suspected to have developed within the valleys of the Stinchar and Water of Girvan.

- **Evidence for a marine-influenced ice margin retreat onto south-west Scotland is unattainable.**

A grounding-line fan near the southern promontory of the Machars, despite portraying many similarities to a glaciomarine complex, fails to provide unequivocal evidence for a marine-influenced ice margin retreat onto this part of the Scottish mainland. A more appropriate explanation is that the fan represents a former depocentre within a glaciolacustrine setting, which developed from the damming of meltwater in a NE-SW orientated natural depression at *c* 30 m a.s.l. in which the depositional landform prevails, surrounded to the east and west by topographic high-points (up to 70 m a.s.l.). This is thought to require the presence of ice within the Solway Firth, thereby providing an insight into the deglacial dynamics of the region. This scenario does not require an unusually high sea-level during deglaciation, as is a prerequisite of the glaciomarine theory (+40 m above present), and therefore relates more suitably to the documented mean sea-level of -3 to -4 m at 12 290 ± 250 years B.P. (Bishop and Coope, 1977).

- **Topographically constrained ice flow was subjected to a single readvance during the deglacial phase in south-west Scotland.**

Evidence which is disparate from that used by Charlesworth (1926a, b) to invoke a synchronous Highland and Southern upland readvance is provided by the congruous alignment of bedforms throughout the entire length of the Cree, Ken and Doon valleys which clearly overprint bedforms depicting the initial recession of ice back up towards the Southern Upland massif. The configuration of the lineations argues for one readvance episode of Southern Upland ice, thereby disputing the multi-readvance hypothesis of Charlesworth (1926a). The reactivated advance of the independent valley flows is believed to have extended into the Solway Firth Basin. The Scottish readvance hypothesis fails to account for the valley-confinement of this evidence.

- **Highland ice readvanced into Loch Ryan and transgressed over the area now occupied by Luce Bay, south of the Stranraer isthmus.**

A set of lineations cross-cutting all previous flow sets that are oblique to the orientation of the Loch Ryan Basin provides clear evidence for the readvance of ice into this topographically constrained trough. Both the distribution of the lineations and the positioning of a large lateral moraine on the western side of Loch Ryan Basin, argue for the transgression of this flow event beyond the Stranraer isthmus into the area occupied by Luce Bay. The assignment of a Highland origin to this flow event is based on the presence of Ailsa Craig microgranite erratics within a massive far-travelled glacitectonite exposed on the western banks of Loch Ryan. Correspondingly it is believed that Highland ice, initially receding northwards beyond the northern promontory of the Rhins peninsula and at which time was some distance away from the present coastline, readvanced into the confinements of Loch Ryan. This supports the early conjecture of Charlesworth (1926b), though the re-evaluated evidence firmly disputes the termination of the original reconstruction at the head of Loch Ryan.

- **A palaeosurge event was responsible for the final short-term advance of ice along the Loch Ryan Basin.**

A re-evaluation of the landform assemblage covering the isthmus between the Rhins of Galloway and the Glenluce lowlands provides a number of characteristics which are similar to those incorporated into a surging-glacier landsystems model. These include megaflutings of the order of 1800 m in length superimposed on top of the lateral moraine of Highland readvance age; large-relief ridges of suspected thrust pre-surge sediment; extensive tracts of kame-and-kettle topography; transverse (crevasse-fill?) ridges; and suspected glaciotectonic depressions. A change in palaeo-flow direction along the Loch Ryan Basin is recorded in the massive glacitectonite suggesting a slight readjustment in the flow between the initial Highland readvance and the later palaeosurge. The surge of ice along the confinements of the Loch Ryan Basin is though to have spilled out over the low-lying isthmus in the form of a piedmont lobe.

- **Relative sea-level at the head of Loch Ryan was +6 m above the present level at 6 075 ± 55 years B.P. (AA-33830).**

This date was yielded from a mollusc in marine muds and is based upon the assumption that the sample was discovered in its former growth position. The Middle Holocene date closely supports the height of sea-level inferred by Lambeck's (1993b) sea-level curve at 6 000 years B.P. for the west Solway Firth.

6.2: Recommendations For Further Research

There are several areas where further research could be profitably directed. These can be summarised as follows:

- Borehole records obtained from site investigations conducted by the British Geological Survey fail to deliver any detailed strata descriptions which could potentially allow the differentiation and correlation of subsurface drift units. The depth of the borehole information currently available for the investigation area only records shallow depths generally no greater than 5 m, which is adequate for the construction industry purposes. A renewed search for deeper and more informative borehole records might, in part, compensate for the current scarcity of exposed Quaternary sediment and aid a greater appreciation of the varying types of depositional setting.

- An age constraint on individual flow stages described by the palaeo-ice flow model would allow the accuracy of the reconstruction to be verified. In particular, the correlation of flow sets into discrete flow stages would be more accurately achieved if dateable material could be found in at least a few of the bedforms belonging to each recognised flow set. In order to achieve this, dateable material within the subglacial bedforms would need to be discovered, and this would subsequently depend on the internal constituents of these landforms being made available.

- The identification of erratic dispersal patterns throughout the investigation would undoubtedly provide valuable information allowing an overall assessment of the accuracy of the palaeo-ice sheet dynamics model for this region. Mineral magnetic analysis would also assist this process, but again relies on the availability of exposed glacial deposits.

- An age constraint on the age of the Southern Upland and Highland readvance episode would allow a time-frame to be placed on the chronology of these events. This would also assist in determining whether these readvances correlate with millennial-scale ice sheet oscillation elsewhere in the North Atlantic.

- Any future commercial extraction of sand and gravel in the investigation area would undoubtedly reveal more evidence to clarify the nature of depositional settings in south-west Scotland, and might clarify the character of deglacial events during the final stages of glaciation.

Appendix 1: Grain size scales

Phi (Ø) scale	*Size (mm)*	*British sieve sizes*	*US sieve no.*	*Wentworth scale*
				boulder
−8	256			
−6	64	63 mm		cobble
−4	16	20 mm		
−2	4	6.3 mm	5	pebble
−1.75	3.36	3.35 mm	6	
−1.5	2.83		7	granule
−1.25	2.38		8	
−1	2	2.00 mm	10	
−0.75	1.68		12	
−0.5	1.41		14	very coarse
−0.25	1.19	1.18 mm	16	sand
0	1		18	
0.25	0.84		20	
0.5	0.71		25	coarse sand
0.75	0.59	600 µm	30	
1	0.50		35	
		425 µm		
1.25	0.42		40	
1.5	0.35		45	medium sand
1.75	0.30	300 µm	50	
2	0.25	212 µm	60	
2.25	0.210		70	
2.5	0.177		80	fine sand
2.75	0.149	150 µm	100	
3	0.125		120	
3.25	0.105		140	
3.5	0.088		170	very fine
3.75	0.074		200	sand
4	0.0625	63 µm	230	
4.25	0.053		270	
4.5	0.044		325	coarse silt
4.75	0.037		400	
5	0.031			
6	0.0156			medium silt
7	0.0078			fine silt
8	0.0039			very fine silt
9	0.0020			
10	0.00098			
11	0.00049			
12	0.00024			clay
13	0.00012			
14	0.00006			

Source: Benn and Evans, 1998.

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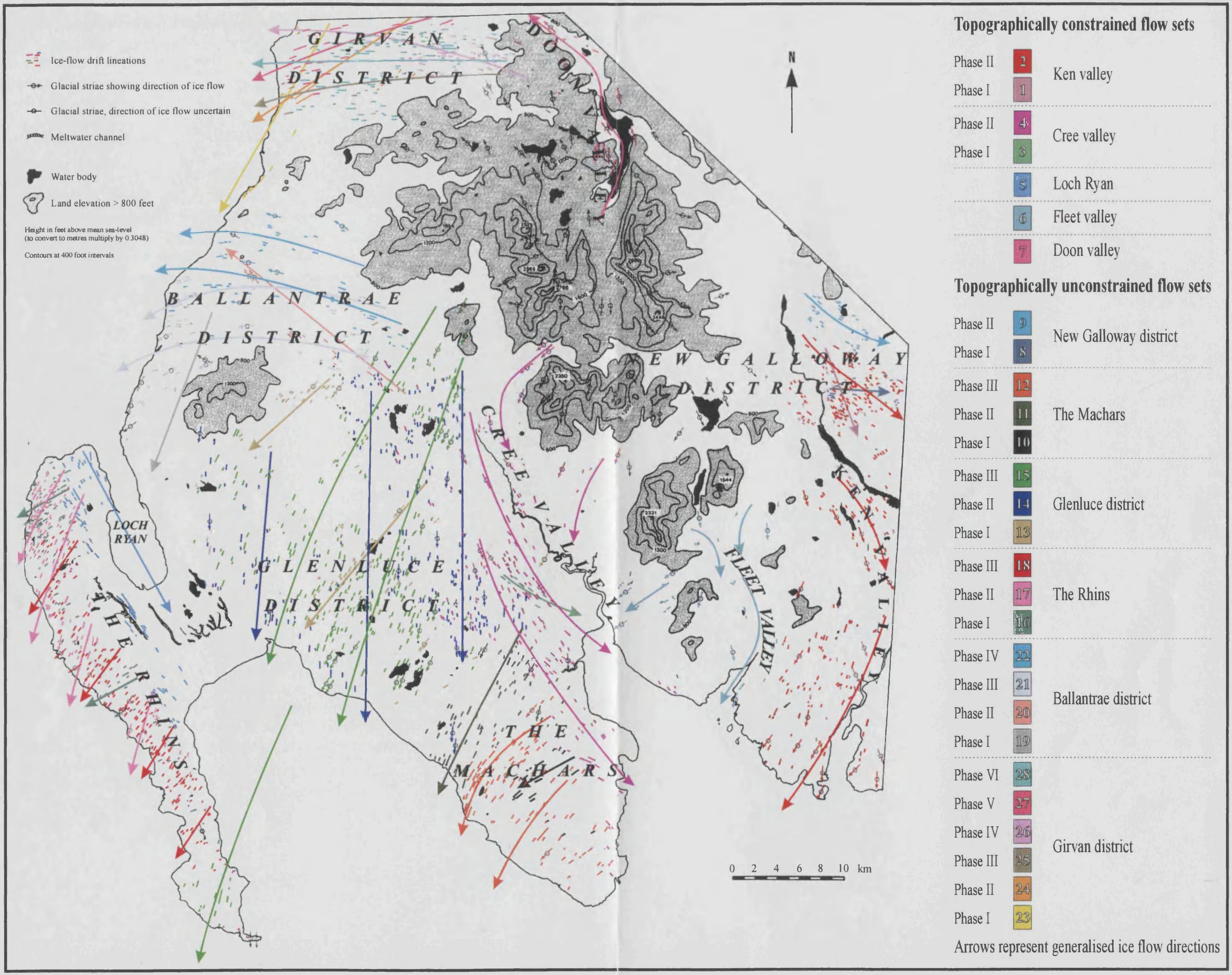

Figure 4.2 Classification of individual lineations into distinct flow sets (1-28) with inferred ice flow direction.

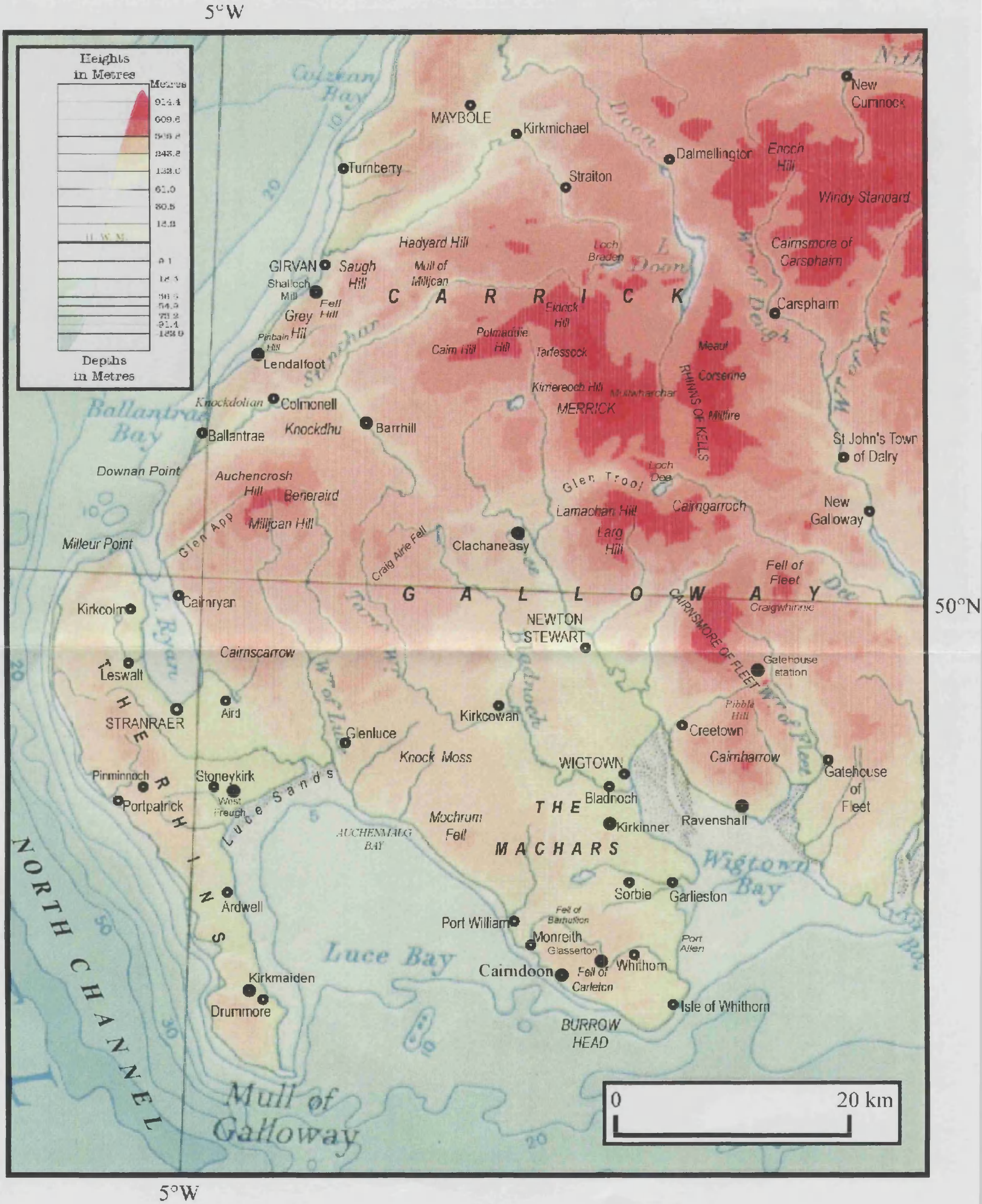

Figure 1.2 Topography and main physical features of south-west Scotland. Submarine contours given in fathoms. (Adapted from the 1: 1 000 000 O.S. map of XVII Century England, 1930).