The geomorphology of Scotland, with particular reference to the Southeast

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D.Sc. abstract - J. B. Sissons

The geomorphology of Scotland with particular reference to the South-East

The submission comprises 29 publications (27 papers, a book, and a chapter from a book) together with a commentary thereon. The publications may be grouped under five main headings, although considerable overlap occurs.

(i) Tertiary landform evolution. It is suggested that slope retreat was much more important in Tertiary landform evolution than is generally believed, that the erosion surface remnants that now survive have been much modified since their initial formation, and that the apparently conflicting views on the origin of erosion surfaces are less divergent than is generally believed. Particular attention is given to the initiation of the Scottish drainage system and a principal initial watershed of simple form is proposed.

(ii) Fluvioglacial landforms. Using evidence obtained particularly by detailed mapping of such landforms in Scotland it is shown that interpretations of fluvioglacial landforms that require the former existence in upland Britain of numerous ice-dammed lakes are largely invalid. These interpretations, which prevailed for over half a century, are replaced by interpretations involving extensive subglacial drainage systems. In addition the concept of the englacial water-table is introduced.

(iii) Relative sea-level changes. Relative sea-level changes in Scotland during lateglacial and postglacial times were for eighty years almost always considered in terms of the 100-ft, 50-ft, 25-ft and possible 15-ft raised beaches. This approach is entirely rejected and a new pattern of

land/sea-level changes is developed in detail for South-East Scotland and tentatively outlined for the rest of the country. The need for accurate levelling of raised shorelines and related outwash deposits at closely-spaced intervals is stressed and the results of such work presented. The need to study buried morphology (as well as marine/estuarine stratigraphy) by means of numerous closely-spaced boreholes is emphasised, and the existence of buried raised beaches and a major buried marine erosional feature in the Forth valley demonstrated. It is also shown by very detailed work that the widely-adopted working hypothesis that glacio-isostatic uplift results in regular tilting of raised shorelines does not apply to part of the Forth valley. It is further suggested that important glacial readvances may result in renewed downwarping of the earth's crust.

(iv) Readvances. Three glacial readvances have been tentatively postulated by the writer as having occurred in Scotland. The first of these (the Aberdeen-Lammermuir Readvance) is now rejected. The second (the Perth Readvance) is currently the subject of discussion and some salient points relating to this discussion are given in the commentary. The third readvance (the Loch Lomond Readvance), originally mapped by the writer at a small scale in Scotland as a whole, is now firmly established and its limits have been mapped at a larger scale in certain parts of Scotland.

(v) Some practical applications. While drift borehole information obtained in the course of commercial site investigations, mineral prospecting, etc. can constitute valuable evidence relating to landform evolution in certain areas, it can also be used for practical purposes by geomorphologists. The latter should use their specialist knowledge to correlate and

plot on maps various aspects of surface geology relating to foundation conditions, such as rockhead relief, drift thickness and character of surface deposits. Examples of such maps, which can be of considerable value for planners, civil engineers, etc., are given for the Grangemouth area and for central Edinburgh.

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Publications

1. The deglaciation of part of East Lothian. Trans. Inst. Brit. Geogr., 25, 1958, 59-77.

2. Subglacial stream erosion in southern Northumberland. Scot. Geogr. Mag., 74, 1958, 163-174.

3. Supposed ice-dammed lakes in Britain with particular reference to the Eddleston valley, southern Scotland. Geogr. Annlr., 40, 1958, 159-187.

4. Subglacial, marginal, and other glacial drainage in the Syracuse-Oneida area, New York. Bull. Geol. Soc. America, 71, 1960, 1575-1588.

5. Erosion surfaces, cyclic slopes and drainage systems in southern Scotland and northern England. Trans. Inst. Brit. Geogr., 28, 1960, 23-38.

6. Some aspeets of glacial drainage channels in Britain, Part I. Scot. Geogr. Mag., 76, 1960, 131-146.

7. Some aspects of glacial drainage channels in Britain, Part II. Scot. Geogr. Mag., 77, 1961, 380-392.

8. A subglacial drainage system by the Tinto Hills, Lanarkshire. Trans. Edinb. Geol. Soc., 18, 1961, 175-193.

9. The central and eastern parts of the Lammermuir-Stranraer Moraine. Geol. Mag., 98, 1961, 380-392.

10. A re-interpretation of the literature on lateglacial shorelines in Scotland with particular reference to the Forth area. Trans. Edinb. Geol. Soc., 19, 1962, 83-99.

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12. The glacial drainage system around Carlops, Peeblesshire. Trans. Inst. Brit. Geogr., 32, 1963, 95-111.

13. The Perth Readvance in central Scotland, part I. Scot. Geogr. Mag., 79, 1963, 151-163.

14. The Perth Readvance in central Scotland, part II. Scot. Geogr. Mag., 80, 1964, 28-36.

15. Peat bogs in a postglacial sea and a buried raised beach in the western part of the Carse of Stirling. Scot. J. Geol., 1, 1965, 247-255. (with D. E. Smith)

16. Some pre-carse valleys in the Forth and Tay basins. Scot. Geogr. Mag., 81, 1965, 115-124. (with R. A. Cullingford and D. E. Smith)

17. Raised shorelines associated with the Perth Readvance in the Forth valley and their relation to glacial isostasy. Trans. Roy. Soc. Edinb., 66, 1965, 143-168. (with D. E. Smith)

18. Quaternary, in The geology of Scotland, ed. G. Y. Craig, 1965, 467-503.

19. Relative sea-level changes between 10,300 and 8,300 B.P. in part of the Carse of Stirling. Trans. Inst. Brit. Geogr., 39, 1966, 19-29.

20. Lateglacial and postglacial shorelines in South-East Scotland. Trans. Inst. Brit. Geogr., 39, 1966, 9-18. (with D. E. Smith and R. A. Cullingford)

21. Glacial stages and radiocarbon dates in Scotland. Scot. J. Geol., 3, 1967, 375-381.

22. The evolution of Scotland's scenery, 1967, 259 pp.

23. Drift stratigraphy and buried morphological features in the Grangemouth-Falkirk-Airth area, central Scotland. Trans. Inst. Brit. Geogr., 48, 1969, 19-50.

24. Isobases for the Main Perth Raised Shoreline in South-East Scotland as determined by trend-surface analysis. Trans. Inst. Brit. Geogr., 46, 1969, 45-52. (with D. E. Smith and R. A. Cullingford)

25. Drift stratigraphy and buried morphology beneath the Forth at Rosyth. Scot. J. Geol., 6, 1970, 272-284. (with D. W. Rhind)

26. Geomorphology and foundation conditions around Grangemouth. Quart. J. Eng. Geol., 3, 1970, 183-191.

27. The last glaciers in the Lochnagar area. Scot. J. Geol., in the press. (with A. J. M. Grant)

28. The geomorphology of central Edinburgh. Scot. Geogr. Mag., in the press.

29. Dislocation and non-uniform uplift of raised shorelines in the western part of the Forth valley. Trans. Inst. Brit. Geogr., in the press.

Introduction

The papers comprising this submission date back to 1958. Since that time the writer's knowledge of and experience in geomorphology have constantly increased, and he now no longer accepts certain interpretations given in some publications included in this submission. One purpose of the commentary that follows is to indicate and account for these changes in interpretation.

For convenience of presentation the commentary is divided into five sections, although there is considerable overlap between the sections. For example, the section on readvances involves papers on fluvioglacial landforms and on raised shorelines.

All but two of the publications submitted are entirely or largely concerned with Scotland. The two exceptions are papers 3 and 4, which are included since they are relevant to the development of the writer's ideas on fluvioglacial landforms.

The writer's book "The evolution of Scotland's scenery" was given a slightly popular title for an obvious reason. It could equally well have been called "The geomorphology of Scotland" or "The historical geomorphology of Scotland". Although certain items in the text (e.g. explanations of terminology) are included for the non-specialist reader, the book was written as a research contribution that sought to integrate the available evidence on the historical geomorphology of Scotland in terms of the writer's ideas.

Especially since 1962, when the writer began his fieldwork on relative sea-level changes, much of his research has been closely

linked with that of his research students. This has resulted in a number of joint papers. The writer has written the whole or a large part of each of these papers, sometimes re-writing the research student contributions, and for each paper did a considerable part of the research.

The writer wishes to acknowledge his great indebtedness to his research students. He has supervised or is still supervising twenty-one Ph.D. students and one M.Sc. student, all of them working on aspects of the Quaternary geomorphology of Scotland (excepting one whose area was situated just across the Border in the Cheviots). Of these students, ten have been awarded research degrees, one has withdrawn, and the work of the remaining eleven is still in progress. These students provide (and have provided) constant stimulation, and discussions and fieldwork with them are (and have been) the most rewarding part of the writer's academic life. The work of some of these students is particularly relevant to the commentary that follows and is mentioned where appropriate.

Commentary

Tertiary landform evolution

This section involves only two publications, namely paper 5 and chapter 1 of "The evolution of Scotland's scenery". The fact that the latter contains one chapter on Tertiary landform evolution compared with eleven on subsequent events has been criticised adversely by some reviewers. This imbalance in the book and in the writer's research as a whole has two main causes. Firstly, the writer's Ph.D. study, which was concerned mainly with Tertiary landform evolution in Southwest Yorkshire (summarised in Sissons, 1954), left the writer with a feeling of dissatisfaction owing to the impossibility of accumulating sufficient evidence to test his hypotheses thoroughly. When, immediately on concluding this study in 1953, he was appointed to a post in the University of Edinburgh and found himself surrounded by a super-abundance of Quaternary landforms, it was inevitable that he concentrated his attention on these features. Secondly, there is a large volume of published work on the Scottish Quaternary, which contains a fascinating and challenging accumulation of information and conflicting opinions. On the other hand, while divergent views certainly exist on the evolution of Scottish Tertiary landforms, the amount of published work on them is small. This discrepancy is constantly increasing for, during the last ten years or so, there has been a steadily-increasing output of publications on the Scottish Quaternary while very few (though usually important) publications have been devoted to Tertiary landforms.

The widespread and ever-increasing interest in Quaternary problems (which, of course, is not confined to Scotland) is reflected in the speed with which controversial points are discussed in the literature. For example, the writer's first contribution on raised shorelines (10) was immediately criticised (Earp et al., 1962). On the other hand, his discussion of Tertiary landforms in paper 5, published eleven years ago, has still not received serious consideration in the literature. Among the points of general application to the British Isles that are made in this paper and in 22 (chapter 1) are that slope retreat has been very important in the evolution of Tertiary landforms, that the widespread view that the erosion surface remnants that now survive have been little modified since their initial formation is not acceptable, and that the apparently-conflicting views as to whether erosion surfaces are of marine or subaerial origin are much less divergent than is generally believed. The writer is of the opinion that, if these points are taken into account and if the significance of the intense Tertiary erosion demonstrated in the Tertiary volcanic areas by T. N. George (1966; see also 22, p. 21) is recognised, considerable revision of current views on Tertiary landform evolution in the British Isles will result.

Fluvioglacial landforms

The first fieldwork by the writer in Scotland was the mapping of fluvioglacial landforms in a considerable part of East Lothian (1). At the time this fieldwork was carried out British interpretations of fluvioglacial landforms were based on the views of P. F. Kendall published in his classic paper of 1902, and workers envisaged backwasting

ice, generally-impenetrable ice margins, and frequent ice-dammed lakes. Kendall himself, along with E. B. Bailey, had studied the fluvioglacial landforms of East Lothian (1908). The writer found great difficulty in applying Kendall's views in the field, but it was not until he consulted Scandinavian work, especially that of C. M. Mannerfelt (1945), that he came to realise that ice-dammed lakes had been much less common than Kendall believed, that subglacial drainage had been important, and that the downwasting of an ice mass and its eventual stagnation are usually significant considerations in the interpretation of fluvioglacial landforms. In paper 1 interpretations of this type were incompletely developed. For example, many channels interpreted as having been formed along an ice margin could well have been formed submarginally, for (like other workers) the writer had not appreciated that, while a contour channel implies the former existence of an ice margin on its downslope side, it does not imply that ice did not exist on its upslope side also.

In paper 2 a much smaller area of fluvioglacial landforms was considered in more detail. This area was chosen because it contained two large meltwater channels with up-and-down long profiles (Peel, 1951) for which the lake overflow interpretation given by A. R. Dwerryhouse (1902) was clearly unsatisfactory. The writer attempted to show by detailed argument and by analogy that the large meltwater channels had been mainly formed beneath the ice. The distinction between large meltwater channels that functioned for a considerable time and small channels produced in a short time (by local meltwaters) was recognised, and it began to be appreciated that gorges in valley floors occupied by present streams and frequently loosely termed "postglacial" in the literature are often of

subglacial and/or proglacial origin.

On the other hand, in paper 2 and also in paper 3 the writer continued to interpret as of marginal origin many meltwater channels that may well have been formed submarginally, and on this basis continued to plot ice margins of questionable validity. This too-ready acceptance of marginal channels derived from the fact that the writer was so impressed with Mannerfelt's work that he failed to appreciate its limitations (cf. 2, p. 164). However, in paper 8 (pp. 180-184) this error was recognised through consideration of a series of more or less parallel channels that at first had seemed to be excellent examples of successively abandoned marginal features.

In studying fluvioglacial landforms in the Eddleston valley a clear altitudinal relationship between the lower ends of certain northward-sloping meltwater channels and the upper limit of the main area of fluvioglacial deposition was found (3). In order to explain this relationship, which was not explicable by the ice-dammed lake hypothesis, the concept of the englacial water-table was introduced. It was also shown that the water-table ceased to exist with the development of southward drainage beneath the ice. The same concept enabled fluvioglacial features in the Cowaselon valley, New York State, to be explained (4, pp. 1579-1580), and was especially valuable in helping to explain the relationship between kames and kettles on the one hand and a "parallel road" and related depositional flats on the other in the Lochaber area. In this area the anticipated relationship between an ice-dammed lake and an englacial water-table was demonstrated (22, pp. 117-122).

In paper 4 the ideas developed in Britain are shown to be applicable

in part of New York State. Here there is much evidence of subglacial drainage, while there is a great contrast between the huge channels that carried waters from the Great Lakes area and those produced by locally-derived waters. Furthermore, owing to the great size of some of the channels and depositional masses, it could be shown that certain channels have a complicated history. It may be added that one result of the New York investigation was to convince the writer that considerable revision of the story of Great Lakes drainage through the area would be needed, for this story was (and is still) based on old concepts of the character of glacier wastage and drainage.

While working as a research student under the writer's supervision R. J. Price (1960) put forward the idea that certain series of meltwater channels in the upper Tweed valley were formed by superimposition of englacial streams on to the ground buried beneath the ice. This idea is developed further in paper 12 in the context of the complicated channel system around Carlops. This theme has also been taken up by C. M. Clapperton (1968) in the Cheviot Hills and is also considered in 22, chapter 6.

The wider applications of the writer's work on fluvioglacial landforms are considered in paper 3 (pp. 177-185) and especially in papers 6 and 7. In papers 3 and 6 the unsatisfactory nature of Kendall's ice-dammed lake hypothesis is discussed. It should be emphasised, however, that Kendall was himself well aware of the limitations of his interpretation (6, p. 133). By an analysis of the British literature on meltwater channels it is shown that many workers found difficulties in applying Kendall's hypothesis, but they regarded these difficulties as local anomalies of no general significance. Taken together, however,

these anomalies constitute a strong body of evidence. In paper 7 this consideration of the literature is continued and in addition the types of evidence that permit the origins of different types of channels to be inferred are discussed. The influence of the writer's work in this field is indicated in Embleton and King's book (1968, chapter 12).

Relative sea-level changes

The writer's first paper on this topic (10) is entirely theoretical and resulted from an analysis of the literature on Scottish raised shorelines and related evidence. Innumerable inconsistencies in the then widely-accepted "100-foot", "50-foot", "25-foot" and possible "15-foot" raised beach sequence were revealed by comparison of different publications and an alternative type of interpretation was presented. This re-interpretation of the literature was criticised by officers of the Geological Survey (Earp et al., 1962) and the writer replied to their criticisms (Sissons, 1962). The Geological Survey criticisms were understandable, for the old interpretation was almost entirely based on Survey work, which had been carried out over a period of eighty years. Furthermore, the writer's re-interpretation was not based on any fieldwork of his own.

Immediately on finishing paper 10 the writer commenced fieldwork on the south side of the River Forth and Firth of Forth. Raised beaches and related features were mapped on the 1:10,560 scale and shoreline measurements were made with an aneroid barometer. The latter was soon found to be unsatisfactory (22, pp. 167-168) and accurate levelling from Ordnance Survey bench marks was subsequently carried out instead. Meanwhile, D. E.

Smith commenced research under the writer's supervision on the raised marine evidence along the northern side of the River Forth and Firth of Forth. Shortly after R. A. Cullingford began a similar study taking in both sides of the Firth of Tay, the lower Earn and Tay valleys, and the Fife coast as far as Fife Ness, where he linked up with Smith. This work by three people resulted in the rapid accumulation of a large body of evidence, including more than 10,000 accurately levelled altitudes (4,000 by the writer) mostly located on raised shorelines, but including measurements along outwash spreads, on kame summits and on the rims of kettle holes. During the course of this work it soon became apparent that the writer's rather premature criticisms of the then widely-accepted raised shoreline sequence were justified and that a completely new story of relative sea-level changes was developing. Some early results of the measurements of shorelines on the south side of the Forth are given in paper 11, while detailed results for the Perth raised shorelines in the Forth valley west of Burntisland and South Queensferry are given in paper 17.

During the course of his research D. E. Smith put down boreholes using a small Hiller borer and found that the sequence long ago described by T. F. Jamieson (1865) as comprising carse clay resting on a peat layer, which in turn overlay marine/estuarine sediments is widespread. This made the writer realise that much information relevant to sea-level changes could be obtained by hand boring and since 1964, by when the visible shorelines and related features south of the Forth had been mapped and levelled, almost all the writer's work on relative sea-level changes has been concerned with borehole evidence. More than 1850 hand bores have now been put down by the writer, while north of the Forth D. D.

Kemp has put down some 700 bores. R. A. Cullingford has also put down a limited number of bores in the Tay area. In addition a large amount of commercial bore data has been assembled, such bore records being especially numerous for Grangemouth and its vicinity.

The first problem investigated by boring arose from consideration of an unpublished map produced by the Scottish Peat Survey showing surface peat thickness in East Flanders Moss. This map shows part of this extensive surface peat accumulation as occupying a steep-sided flat-floored depression in the carselands, the depression being several metres deep and almost 2 km across at its widest part. Since it is almost impossible for a depression of this form to exist in a raised mud flat in an area not subject to subsidence, an alternative explanation was sought, and it was inferred that the surface peat and buried peat merged together owing to the carse clay not having been deposited here. This inference was subsequently verified by boring in and near East Flanders Moss and also in and near West Flanders Moss, where a similar anomaly had been revealed by the Peat Survey evidence. It was found that the carse clay gradually thinned out as the area of deep peat was approached, implying that peat accumulated contemporaneously with the surrounding mud flat (15).

In paper 15 the extensive flattish surface of the fine sediments underlying the carse clay and peat was interpreted as a buried raised beach. It was therefore decided to try to map this raised beach in the vicinity of the Menteith Moraine and to locate and determine the altitude of its shoreline. This investigation (19) in fact demonstrated the existence of three distinct buried raised beaches (referred to as the High, Main and Low buried beaches), each composed of slightly

different deposits, as well as the occurrence of buried valleys and buried outwash deposits. The beaches were dated by pollen analysis (Newey, 1966), radiocarbon assay, and morphological relationships. These three beaches have since been identified by D. D. Kemp over a considerable area on the north side of the River Forth. The High Buried Beach has been found by the writer to extend along the south side of the Forth valley intermittently to Stirling and he has traced the Main Beach almost continuously to near Airth, where the Low Beach has been mapped also (23, 29). R. A. Cullingford has identified buried beaches in the Tay area, one of which, on the evidence of pollen analysis and radiocarbon dating, is probably the Main Buried Beach.

The main contribution in paper 23 concerns the buried gravel layer, located by hand bores and commercial bores over an area of 28 sq km. This feature is remarkable in that it proves a period of extensive marine erosion in lateglacial times in an area that is (and was) located a considerable distance from the open sea. The feature is also clearly revealed at Rosyth by a large number of old bore records that the writer and D. W. Rhind were fortunately able to obtain (25) and it has also been identified by D. D. Kemp in the Kincardine area. The clear development of buried beaches and the buried gravel layer in various parts of South-east Scotland strongly suggests that they exist in other parts of the Scottish coastlands, but no investigations of them have yet been published.

Valleys cut into the buried raised beaches are quite frequently encountered. Numerous small valleys in the High Buried Beach were found near the Menteith Moraine (19), while larger ones, often continuing

gullies cut into lateglacial raised beach deposits, occur in various parts of Southeast Scotland. In paper 16 these gullies were shown to have been formed before 8000-8500 B.P., but in paper 23 it is demonstrated that they were formed before about 9500 B.P. and their periglacial origin, tentatively mentioned in the earlier paper, is favoured. In addition, a second series of buried valleys related to the low sea-level that occurred about 8500 B.P. is referred to in paper 23.

In order to construct isobases for the Main Perth Raised Shoreline, which is the most conspicuous lateglacial shoreline in Southeast Scotland, and for which over 500 altitudes have been obtained, trend surface analysis was used (24). The computer work for this paper was done by D. E. Smith. Trend surface analysis is currently being employed to establish isobases for the Main Postglacial Shoreline in Southeast Scotland using more than 1000 altitudes.

Linear trend surface analysis and linear regression analysis (the latter having been used in papers 11, 17, 19 and 29) reproduce shorelines as plane surfaces and straight lines respectively. It has long been recognised, however, that raised shorelines in areas affected by glacial rebound, are very unlikely to be of such simple form: it is reasonable to expect that gradient diminishes near the centre of the dome of uplift and also near the margin of the isostatically-uplifted area. In addition, the possibility of irregular uplift due, for example, to movement along fault lines, has long been appreciated. In raised shoreline investigations such as those being considered here, however, it is necessary initially to adopt the simple hypothesis of regular tilting unless there is clear evidence to the contrary, and the fact that the

linear trend surface for the Perth Main Shoreline accounts for 99.26% of the sums of squares (24) justifies this approach in the present case. Yet, despite this and other statistical evidence, the writer has suspected throughout his shoreline investigations that uplift was not of this simple form. A very detailed study of the altitude of the Main Buried Shoreline was therefore undertaken for comparison with the altitudinal evidence available for the Main Postglacial Shoreline. The results of this study are described in paper 29 and it is suggested that they show conclusively that in the western part of the Forth valley isostatic uplift has not occurred as uniform tilting. It should be added that the several sections of the Main Buried Beach have been shown to be parts of one and the same feature by pollen analysis (C. L. Brooks, in the press).

The shoreline sequence in Southeast Scotland as a whole is discussed and illustrated in paper 20 and again in 22, chapter 10. The six lateglacial shorelines in eastern Fife are based entirely on the work of Cullingford and Smith (1966), while the Main Perth Shoreline is based on their work and on that of the writer. The remaining shorelines are largely derived from the writer's investigations.

In "The evolution of Scotland's scenery" relative sea-level changes in Scotland as a whole are considered. A great deal of information in the literature, comprising several hundred publications, relates to the form, composition and fossil content of the lateglacial and postglacial marine deposits. On the other hand, very little accurate altitudinal information on visible raised shorelines was (or yet is) available outside Southeast Scotland and none has been published on buried shorelines. Furthermore, no detailed shoreline sequences have been published

for other parts of Scotland, excepting that by F. M. Synge and N. Stephens (1966), which the writer cannot accept (Sissons, 1967). Consequently the pattern of sea-level changes in Scotland as a whole (22, chapter 11) had to be based on the pattern established in the Southeast. While the evidence from the Southeast can be reconciled with that from the rest of Scotland, it is obvious that detailed studies throughout the Scottish coastlands are needed before a reliable story for the country as a whole can be established. Fortunately, several such studies are now in progress, two of them under the writer's supervision.

Readvances.

On the European mainland and in the Great Lakes area of North America various positions of the margin of the last ice-sheet of the respective areas have been identified. Often these marginal limits correspond with important moraines and it is presumed (or demonstrated beyond reasonable doubt) that the moraines mark the limits of readvances that interrupted the general decay of the last ice-sheets. One might therefore expect a similar succession of readvances to have occurred in the British Isles, some of them at least, being recorded in Scotland. Yet the only unequivocal evidence for a glacial readvance in Scotland is that for the Zone III or Loch Lomond Readvance. Outside this readvance limit end moraines are poorly developed or absent. This lack of well-defined end moraines is one factor contributing to the present uncertainty regarding the various limits of the last ice-sheet in Scotland.

The writer's first contribution on the topic of readvances (9) was of a negative nature in that it sought to show that the eastern and

central parts of J. K. Charlesworth's Lammermuir-Stranraer moraine (1926b) do not record a readvance or an ice limit. This contribution followed from studies of fluvioglacial landforms in parts of East Lothian, Midlothian, Peeblesshire and Lanarkshire. These studies (1, 3, 8) showed that the fluvioglacial mounds were deposited in association with dead ice and that their frequent occurrence together in large numbers often reflects topographic controls. The detailed mapping of meltwater channels and eskers also provided indirect evidence of the direction of former ice movement. Thus in and near the Eddleston valley the channels decline towards the north and northeast, implying formation in association with Southern Uplands ice, in accord with evidence from erratics but contrary to J. K. Charlesworth's view that Highland ice was the last to occupy this area. In the Clyde valley the Thankerton and Carstairs systems of ridges, variously interpreted by different writers and considered by Charlesworth to be ice-marginal features, are esker systems and possess the typical features of such systems. (G. A. Goodlet (1964) subsequently interpreted the Carstairs ridges as of marginal origin but A. G. McLellan reaffirmed the esker origin (1969).)

In paper 9 the writer's first argument against J. K. Charlesworth's interpretation of the Lammermuir-Stranraer features as an end moraine was that they are composed almost entirely of fluvioglacial material, yet the writer subsequently made much use of assemblages of fluvioglacial depositional forms in attempting to delimit the Perth Readvance (13, 14). This change of view reflects the writer's undue concern, when paper 9 was written, with glacier stagnation and the development of dead ice. He did not sufficiently appreciate the possibility that a glacial readvance, brought to an end by rapid climatic amelioration, could result in the

development of a stagnant ice-marginal zone and the formation of belts of fluvioglacial forms which, since the active ice of the readvance would have been considerably influenced by relief, would often be related to the major relief features. It is also relevant to note, with reference to the first paragraph of this section, that, since no clear evidence of major end moraines with a significant till content has yet been presented in Britain for the area between the maximal limits of the late Weichselian ice-sheet and the Zone III limit, it may well be that the British equivalents of Continental and North American readvance limits are belts of fluvioglacial deposits, a possibility likely to be enhanced by a more rapid response to climatic change of the relatively small British ice-sheet situated in a maritime location as compared with the Continental and North American ice-sheets.

Despite these considerations the main case presented in paper 9 still stands and additional evidence for Southern Uplands ice having extended north of the Southern Uplands Fault in Peeblesshire and Midlothian has been presented in paper 12 as well as by R. P. Kirby (1968, 1969).

The writer's first attempt to define the limit of a readvance is represented by papers 13 and 14. The evidence given for this readvance is considered to be strong but by no means conclusive. New evidence has become available since these papers were published and some of the salient features relating to the validity of the readvance in the light of current knowledge may be summarised as follows.

(i) Part of a mammoth tusk found in Ayrshire beneath thick till presumed to have been laid down during the Perth Readvance has been radiocarbon-dated as 13,700 (+1300, -1700) B.P. (21). This date, if

valid, proves a late readvance of ice in this part of central Scotland. The date does not accord with a date of 27,550 (+1370, -1680) B.P. reported by W. D. I. Rolfe (1966) for a bone of woolly rhinoceros found near Bishopbriggs in deposits associated with the Perth Readvance. Since the woolly rhinoceros bone was water-worn and was obtained from fluvio-glacial deposits it can be argued that it was derived from an earlier phase and need not be inconsistent with the Ayrshire date. However, a reindeer antler from the same site as the mammoth tusk has recently been dated as greater than 40,000 B.P. (F. W. Shotton et al., 1970). These three very different dates for materials apparently associated with the presumed readvance present a problem that has yet to be solved.

(ii) The date of 13,500 to 13,000 B.P. suggested for the culmination of the Perth Readvance in paper 21 is consistent with radio-carbon dates now available for shells from the glacio-marine deposits in the Glasgow area. The oldest dated shells are in the range 13,000 to 12,500 B.P. (W. W. Bishop and J. H. Dickson, 1970; J. D. Peacock, 1971).

(iii) In both the Forth and Tay valleys important outwash deposits, which merge into the most extensive lateglacial raised beaches in the whole of Southeast Scotland, have been correlated with the Perth Readvance. In both valleys the relation of beaches to outwash deposits indicates rapid relative emergence following the culmination of the presumed readvance. The inference in paper 13 that these phenomena in the Forth and Tay valleys were produced contemporaneously is strongly supported by trend surface analysis of the Main Perth Raised Shoreline (24).

(iv) I. B. Paterson (personal communication) has shown that in Strathmore beyond the Perth Readvance limit mapped by the writer the fluvioglacial landforms have the same general trend as within the mapped limit, suggesting that the latter may be invalid. While it is possible that ice movement (and hence the principal fluvioglacial forms) had a similar direction in Strathmore during both the Perth Readvance and the previous more extensive ice cover, the critical evidence is at Almondbank (J. B. Simpson, 1933). Here part of an extensive outwash plain containing large kettles overlies varved sediments 12 m thick, thus proving forward movement of ice over ground it had previously vacated. Furthermore, the varves (whose authenticity is supported by the authority of G. de Geer (1935)) indicate about 640 years of ice-free conditions at Almondbank, implying a significant time interval before the ice readvanced to this point.

(v) In paper 13 the writer suggested that the gravel layer with boulders that overlies the lateglacial clay and underlies the carse clay in the Forth valley (e.g. around Grangemouth) represents ice-berg deposits associated with the Perth Readvance. This interpretation was subsequently shown to be erroneous (23). The removal of this "evidence" leaves no clear stratigraphic evidence for the readvance in the Forth valley.

(vi) In the Clyde coalfield basin the stratigraphic evidence for a readvance is considerable (14, pp. 32-33) and the morphologic evidence described in paper 14 has been accepted, with modifications, by A. G. McLellan (1969). The critical evidence that is required, however, in this area and elsewhere, is a series of radiocarbon dates. When suitable materials for this purpose are discovered the problem of whether or not there was a Perth Readvance should be finally solved. It should

be added, however, that even if it should be shown that a significant readvance did not occur, it is still probable that the limit mapped from morphological evidence from the Perth area to Lanarkshire marks a definite stage in the decay of the ice-sheet (a halt or minor readvance). One reason for this statement is that the limit near Perth has been shown to correlate with the limit in the Forth valley by the independent method of trend surface analysis of the Main Perth Raised Shoreline and it is therefore reasonable to infer that the limit mapped between these two localities has some validity.

In 18 the writer was required to discuss the Quaternary of Scotland as a whole and an attempt was made to identify readvance limits. Much use was made of published work, especially the outstanding contributions of J. K. Charlesworth (1926a, 1926b, 1955). The map of readvance limits was reproduced with slight modifications in 21 and 22 (Fig. 59), but it was emphasised that even though the then available radiocarbon dates could be fitted to the scheme presented this was by no means evidence of its validity. The present uncertainty regarding the Perth Readvance has been discussed sufficiently above. The Aberdeen-Lammermuir Readvance is now considered not to have occurred. Detailed field mapping by the writer of ice-moulded landforms over 1400 sq km of the Tweed basin (not included in this submission) shows that these landforms cross the supposed readvance limit without any change of direction or character. The pattern of fluvioglacial landforms in the Tweed area is also opposed to the readvance concept. D. E. Sugden and C. M. Clapperton (personal communications) have evidence indicating that the limit identified by F. M. Synge (1956) in the Aberdeen area is also incorrect.

The Loch Lomond or Zone III Readvance is firmly established, although

much detailed work has yet to be done to determine its limits exactly. Evidence of local valley glaciation was recognised in the 1840's by the early glacialists and during the 1860's it came to be widely (though not universally) accepted that the mer de glace was followed by a valley glaciation, although the limits of the latter were not defined. In the present century this simple pattern was lost sight of but the writer has tried to emphasise the importance of this pattern, for the Loch Lomond Readvance in many upland valleys of Scotland is recorded by the abundant fresh hummocky moraines, sometimes limited downvalley by clear end moraines, that were frequently described around the middle of last century.

Since paper 21 was written radiocarbon dating has confirmed that the limits at Loch Creran and at Loch Spelve are of Zone III age (J. D. Peacock, 1971; J. M. Gray and C. L. Brooks, in the press). K. S. R. Thompson, working under the writer's supervision, has mapped the presumed Zone III limit in the field in an area of the central Grampians extending from Callander to a point northwest of Blair Atholl and has shown that the ice in some valleys occupied a larger area than depicted in paper 21. On the other hand, D. E. Sugden (1970) has presented a different interpretation for the western Cairngorms, arguing that there was not a readvance of valley glaciers but instead a minor oscillation of a downwasting ice-sheet. This interpretation is discussed in paper 27, in which it is shown that in the adjacent Lochnagar area there is very good evidence for an advance of corrie and valley glaciers. Furthermore, in the Lochnagar area there is a clear relationship between large solifluction lobes and the former glacier limits. This is in agreement with the relationship between large-scale periglacial phenomena and

presumed Zone III limits in the central Southern Uplands and the Applecross peninsula (22, p. 224).

Some practical applications

For a considerable time the writer has been interested in the possible practical applications of the information and ideas gained from the study of geomorphology, and he has sought to develop such applications where reasonably possible. Two relevant publications (26, 28) are included as part of this submission, but most of the writer's work in this field is not included for various reasons (e.g. confidential).

Paper 26 seeks to show that the analysis of drift and rockhead borehole data by a geomorphologist can result in the provision of information of value in site investigations and in local and regional planning. The writer was particularly concerned about the considerable sums of money that are often wasted in site investigations (two specific examples are given on p. 190) and was also concerned that a very large amount of drift borehole information is rapidly "lost" and cannot therefore be used to guide future investigations (or used in research). This concern led the writer to ask the Institute of Geological Sciences for a grant to enable himself and D. W. Rhind to undertake a study of the feasibility of establishing a computer-based drift borehole data bank in the Institute. A report on this project has been submitted to the Institute but is not included since it is mainly the work of Rhind.

The text of paper 28 is mainly concerned with the geomorphological history of central Edinburgh but the very detailed maps were compiled primarily to assist future site investigations. The writer was

surprised that such a large amount of borehole data existed for such a small area (about 1400 bores in an area measuring 2.0 by 2.8 km) and, as pointed out on page 1 of the paper, believes that similar maps should be constructed for the central areas of other cities. The practical value of the Edinburgh maps is not known at present, since copies have only recently been sent to the City Engineer, the City Architect and various consultant engineering firms, but it is known that they are already being used.

The Grangemouth maps were compiled for an area that has a relatively simple drift stratigraphy related largely to past changes of relative sea-level. The work on central Edinburgh shows that, with good borehole cover, meaningful generalisations can be expressed in map form even in areas of very variable drift stratigraphy and rockhead relief.

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THE DEGLACIATION OF PART OF EAST LOTHIAN

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IT has long been recognized that the landforms of East Lothian owe a great deal to glaciation. The most comprehensive treatment of the subject is that of P. F. Kendall and E. B. Bailey,[1] the results of whose work were subsequently incorporated in the second edition of the Geological Survey Memoir on East Lothian[2] along with material contributed by others. Despite the abundance of marginal phenomena over large parts of East Lothian, however, a map showing stages of the ice-retreat during the final deglaciation has not been published, and the construction of such a map was, therefore, the principal object of the present study. In this paper an explanation of the method of constructing this map is preceded by brief descriptions of the principal types of marginal and sub-marginal features present in East Lothian, together with short explanations of their probable modes of origin, and is followed by a discussion of some of the major conclusions of Kendall and Bailey and of J. K. Charlesworth.[3]

The limits of the area considered are shown in Figure 6. Along, or a short distance south-eastward of the Southern Upland Fault, which runs north-east–south-west through the area, the Lammermuirs rise steeply to heights varying between 1100 and 1750 feet. Their principal watershed lies only one to three miles south-east of the fault, so that short swift streams flow northward and north-westward to join the Tyne, the largest river of East Lothian. At the foot of the Lammermuir scarp from Fala to near Garvald is a dissected drift-covered rock bench, situated between heights of about 550 and 700 feet, with a maximum width of 1½ miles near Gifford. South-westward from Fala and north-eastward from Garvald the country immediately in front of the main mass of the Lammermuirs is more irregular and higher than the Garvald-Fala bench, several hills, with Deuchrie Dod the most notable, rising to between 800 and 970 feet. North-westward towards the Tyne there is, on the whole, a decline in summit heights, the most important exception being isolated, steep-sided Traprain Law. Beyond the Tyne is a ridge of variable width and altitude, with culminating points in the Garleton Hills (600 feet) and the Camp Ridge (850 feet).

The main movement of ice was generally towards a point slightly north of east. This is indicated with varying degrees of precision by indicator stones, crag-and-tail features, a few striae, and large grooves cut by the ice in bedrock (particularly well developed north and east of Haddington)[4].

The bedrock of the greater part of the area is concealed by a mantle of till or stratified drift or both, although steep slopes, even in low-lying areas, are frequently drift-free. The till is probably of considerable thickness where it fills former valleys, but is in general fairly thin as evidenced by sections revealed in

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numerous quarries and by the exposure of bedrock along the beds and lower valley sides of many streams. In detail the surface of the till is gently undulating or featureless and its larger forms generally reflect those of the bedrock beneath.

In contrast the stratified drift is associated with a variety of surface forms including kames, kame terraces, eskers and kettles, and valley fills that have been dissected to form river terraces. Although accumulations of stratified drift occur in many parts of East Lothian, they are particularly abundant in two long belts. One of these lies along the northern edge of the Lammermuirs between 550 and 1000 feet, especially between 550 and 750 feet. The other occurs on the floor of the Aberlady-Tynemouth depression and continues along the coast south-east of Dunbar (Fig. 1).

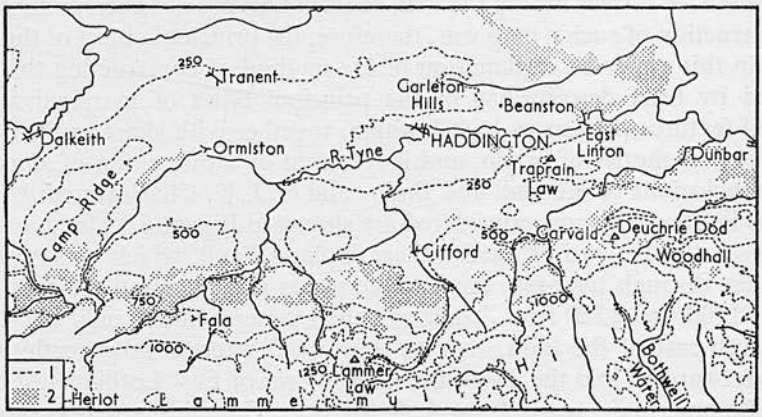


FIGURE 1—Relief, drainage and principal sand and gravel areas
1, Contours: interval, 250 feet. 2, Principal areas of glacial sand and gravel according to Kendall and Bailey and Sheet 33 of the one-inch Geological Map of Scotland. Innumerable small areas of sand and gravel cannot be shown on this small-scale map.

Besides laying down extensive deposits of stratified drift the waters of the streams draining the ice-mass and of those draining the ice-free areas combined to erode a large number of glacial drainage channels[5] along or near the edge of the ice-mass and beneath the ice itself as the latter retreated towards the north-west and north. Most of the larger channels and a few of the smaller ones in the area south of the Tyne were mapped by Kendall and Bailey, but these authors did not show on their published map the numerous ice-marginal benches and kame terraces (although they recognized their existence). Some of the channels and ice-marginal benches of the Garleton Hills, to the north of the Tyne, were described by G. Barrow.[6]

The glacial retreat features were mapped in the field on a scale of 1:25,000, and the results are summarized in Figures 6 and 7b, except that owing to cartographical considerations the kames and many minor features have had to

be omitted. A surveying aneroid was used for altitude determinations and slope measurements were made with an Abney level. Since good exposures in stratified drift forms are usually lacking where they would be most useful a number of sections were dug, but it was practicable to study only a small number of features in this way. In conjunction with the fieldwork the whole area was studied stereoscopically on vertical aerial photographs on a scale of approximately 1:10,000. Figures 2 and 4 were drawn initially from photographs and subsequently modified and completed in the field.

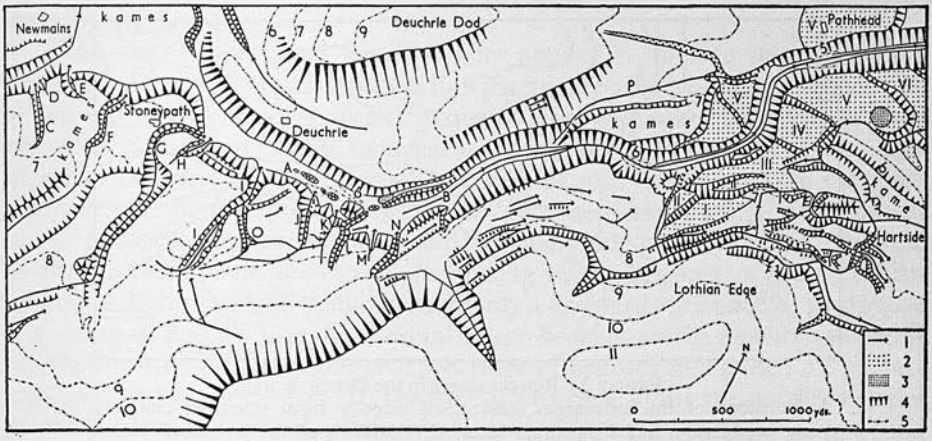


FIGURE 2—Retreat forms in the Deuchrie area.
1, Glacial drainage channels (including sub-glacial chutes) and ice-marginal benches. 2, Surfaces of kame terraces. 3, Kettles. 4, Steep slopes. 5, Contours: interval, 100 feet. National Grid reference of centre of map 36/633714.

Principal Retreat Features

Glacial Drainage Channels

These vary greatly in size, ranging from the gorge 200 feet deep south of Deuchrie Dod (Fig. 2) to faint depressions only a foot or two deep and a few yards broad. In long profile the whole or a large part of some channels appears flat to the naked eye, and instrumental measurements reveal only a very slight 'downstream' slope. Many channels with a gentle gradient for a large part of their length, however, become markedly steeper at their 'downstream' end. This increased gradient has often resulted from rejuvenation caused by retreat of the ice-margin or by the diversion of the meltwaters flowing along the channel into a tunnel or crevasse in the ice-margin, and the retreat of the resultant knick-point only part way along the channel before this was abandoned by meltwaters. In such cases the gently inclined part of the channel profile when projected 'downstream' usually intersects a kame terrace, ice-marginal bench or the floor of another channel.[7]

Marginal channels and depositional features are absent or feebly developed

in some localities where one might expect to find them. One therefore suspects that in such areas there was little or no marginal drainage, and that meltwaters here flowed over, into or beneath the ice, instead of flowing along its edge. That this third possibility, at least, was in fact a common occurrence is demonstrated by numerous channels that were clearly formed in the ice-marginal zone by streams flowing in tunnels beneath the ice or at the bottom of crevasses. Some of these channels are similar to the numerous examples described in parts of Sweden by C. M. Mannerfelt[8] and called by him 'sub-glacial chutes'.

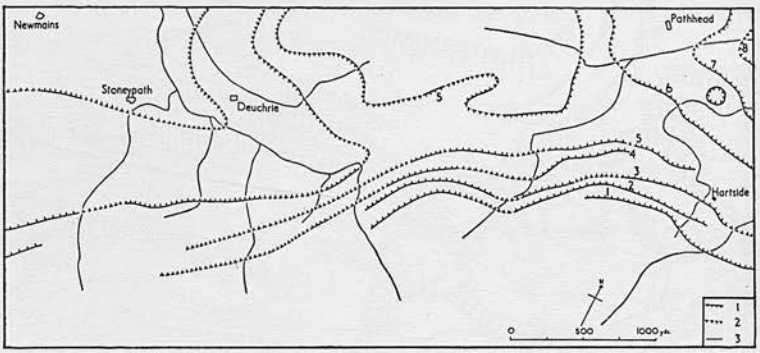


FIGURE 3—Retreat stages in the Deuchrie area.
1, Positions of the ice-margin determined directly from marginal channels, marginal benches and ice-contact faces. 2, Positions of the ice-margin where marginal features are absent or indefinite. 3, Present streams.

The sub-glacial chutes of East Lothian range from a few feet to about 100 feet deep, are usually a few hundred yards long, and where cut in bedrock form narrow, steep-sided valleys. They are generally easily distinguishable from marginal channels, for whereas these run almost parallel with the contours the sub-glacial chutes cross the contours approximately at right angles. Commonly a sub-glacial chute and a marginal channel form one continuous feature, the point at which the meltwaters flowed beneath the ice-margin being indicated by a sharp change in direction of the feature.

Whereas the origin of sub-glacial chutes continuing marginal channels and of streamless chutes cut in bedrock is immediately apparent in the field, that of chutes that contain streams and do not connect with marginal channels is less obvious. In such cases various considerations may point to a sub-glacial origin. For example, in Figure 2 thirteen chutes (C-O) have been mapped on the north- to north-west-facing slope around Stoneypath. The origin of six of these features (F, G, I, J, K and O) is indicated by their relation to marginal channels, but several of the remainder appear at first sight to be normal stream valleys at their lower ends. When followed up the slope, however, the streams are seen to be misfits, for several of them become mere trickles of water occupying steep-sided valleys cut in bedrock. A further consideration is that if these thirteen

valleys are regarded as normal stream valleys it is difficult to explain why they are concentrated along a 1½-mile stretch of slope when only three valleys occur on the remaining 5½ miles of slope forming the two sides of the deep valley running past Deuchrie and Pathhead. When, however, their origin as sub-glacial chutes is recognized, this concentration is readily explained, for large volumes of meltwater flowing from the south-west along and near the ice-margin were available for their excavation. The relation of these and other sub-glacial chutes to marginal channels may be seen in Figures 2, 6 and 7b.

Ice-marginal benches

The term ice-marginal bench is here applied to hillside steps that have resulted from erosion by meltwaters flowing between the ice-margin and ice-free slopes.[9] Some of these steps are probably shallow channels that have been converted into steps owing to mass-movement from the slopes above. Ice-marginal benches frequently occur as continuations of well-marked marginal channels, the flattish floor of the bench and the often steep slope behind it merging respectively into the floor and ice-facing side of the channel. On some of the steeper north- and north-west-facing bedrock slopes of East Lothian are found flights of small benches, each only a few yards to about 20 yards wide (e.g. south of Kidlaw, Fig. 4). Several of the benches on the northern slopes of the Garleton Hills have a long, narrow ridge of bedrock along their ice-contact side, as noted by Barrow.[10] To the east of Deuchrie along the northern slopes of the Lammermuirs all intermediate forms between narrow ice-marginal benches and deep channels with sides of approximately equal height occur (Fig. 2).

Kame terraces

Numerous kame terraces occur in East Lothian. Most of them are several hundred yards long and a few tens of yards wide, but the largest are a mile long and up to 200 yards wide. Although most of the terraces are approximately flat in cross-profile, some rise slightly towards the ice-contact side owing to greater deposition near the ice-margin by streams draining the ice-mass. The slope of the terraces along their lengths exceeds a few tens of feet per mile in only a few cases, and some terraces are approximately horizontal. Terrace faces vary from gentle slopes of a few degrees to steep slopes exceeding 20°. The composition of the terraces is very variable, both as between different terraces and different parts of the same terrace, and ranges from coarse gravel to fine sand, silt and clay, but the principal constituents are sand and gravel. The composition, bedding and slope of some terraces show that they were mainly built by streams diverted along the ice-margin and by streams draining from the ice, while the horizontality and fine grade of others suggest that they accumulated in narrow lakes impounded between the ice-margin and ice-free slopes.

The kame terraces frequently pass into the floors of glacial drainage

channels at their 'downstream' ends. In some parts of East Lothian, as around and east of Stenton, several terraces occur one above the other on a hillside; each can be followed 'downstream' into a channel, the channel passing at its lower end into another terrace, and this in turn into another channel, and so on (Figs. 6 and 7b).

Since during a considerable part of the retreat the ice-margin lay approximately at right angles to the pre-existing valleys of the streams flowing north and north-west from the Lammermuirs, thick deposits of sand and gravel were frequently laid down where these streams were blocked by the ice. Where conditions were suitable the streams continued to aggrade after being diverted along the ice-margin, thus producing kame terraces whose surfaces continued those of the valley fills. Under favourable conditions successive withdrawals of the ice-margin resulted in the production of successively lower kame terraces, but the down-valley extension of the valley fill permitted by the retreating ice-margin was accompanied by dissection of the valley fill farther upstream into paired river terraces, so that each kame terrace was continuous with a river terrace. Many such river terraces have doubtless been formed and subsequently destroyed, but their former existence can be deduced in some instances from adjacent ice-marginal features. Included among those that have survived are the examples around Stenton, where flattish-surfaced deposits of sand and gravel occur in pairs at two different levels on either side of the northward-flowing Sauchet Water; the eastern member of each pair passes imperceptibly into a kame terrace which, when traced along its length, gradually diverges from the course of the stream.

A particularly good series of four kame terraces occurs about three miles south of Gifford between heights of 800 and 1050 feet (Fig. 4). The terraces were built mainly by streams flowing eastward at successively lower levels between the ice-margin and steeply rising ground to the south. The highest terrace (I, Fig. 4) is the most interesting since it differs from others mapped in East Lothian. It consists of a step built against the bedrock hillside and several flattish-topped mounds separated from one another and from the step by hollows up to 25 feet deep. The tops of the mounds and the surface of the step fall into a single plane inclined east-north-eastward at about $2\frac{1}{2}°$. The terrace consists of very coarse gravel, the coarseness of this material accounting for the relatively steep long profile of the terrace and its exceptionally steep face (generally between 25 and 29°). Most of the gravel was brought by the stream flowing northward from the summit of Lammer Law towards the ice-margin with a fall of 600 feet in a mile (A, Fig. 4). A rapid reduction in the rate of flow of this stream on being diverted along the ice-margin resulted in the deposition of large quantities of coarse gravel between the ice-margin and the hillside as well as around and possibly over small blocks of ice separated by crevasses from the main ice-mass. Melting of the detached blocks was completed after the supply of gravel had ceased and thus gave rise to the hollows in the terrace.

Kames

Following the practice of Geikie and most other Scottish geologists and in accordance with the suggestions of R. F. Flint,[11] C. D. Holmes[12] and W. D. Thornbury,[13] the term kame will be applied to mounds 'composed chiefly of gravel or sand, whose form has resulted from original deposition modified by any slumping incident to later melting of glacial ice against or upon which the deposit accumulated'.[14] As so used the term includes features formed in several

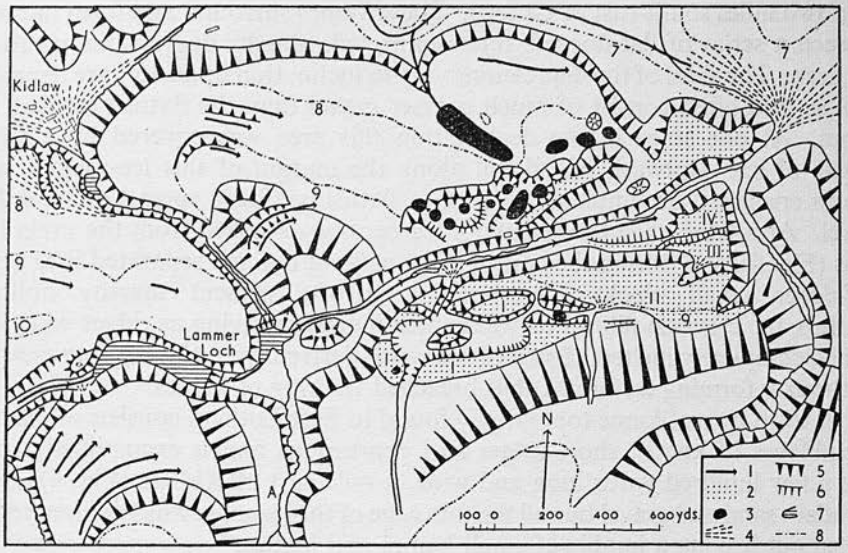


FIGURE 4—Retreat forms of the area east and south of Kidlaw.
1, Glacial drainage channels and ice-marginal benches. 2, Surfaces of kame terraces and related flattish areas of sand and gravel. 3, Kettles. 4, Fans. 5, Steep slopes. 6, Moderate and gentle slopes of kames and kame terraces. 7, Natural and artificial lochs. 8, Contours at 700, 800, 900, 1000 and 1250 feet. National Grid reference of centre of map 36/522638.

different ways, including the crevasse fillings and perforation deposits of some authors. This broad usage must be adopted, however, since, although the form and relation to other features of many of the kames of East Lothian (as of other regions) indicates their probable mode of origin, there remain others whose mode of origin must remain obscure in the absence of sections.

The kames are mostly found in a belt up to two miles wide running along the edge of the Lammermuirs between heights of about 550 and 750 feet, although isolated kames and small groups of kames occur in many other parts of the county. The kames tend to occur in groups in each of which the great majority of the individual members display a similarity of form that strongly suggests a related origin.

Some groups of kames, for example, consist of flattish-topped, roughly parallel mounds separated by glacial drainage channels. In some instances such

kames appear to be the remnants of kame terraces that were dissected by melt-waters shortly after they were formed. This origin may be suggested by the accordance in height of the flat tops of kames situated on opposite sides of a channel or by the relation of kames to surviving kame terraces. An example of a kame whose truncated, almost horizontal bedding forms the side of a channel was mentioned by Kendall and Bailey.[15]

An example of another type of kame topography consisting of flattish-topped mounds of approximately accordant height covers an area of 0.4 square miles two miles south-east of Gifford. These kames surround and form mounds between a series of flat-floored, inter-connected, marshy depressions up to 40 feet deep. The sides of the depressions vary in inclination up to 15°, are irregular in plan, and often consist of much coarser gravel than the flattish tops of the kames. At one stage in the deglaciation this area was covered by a small tongue of ice. Outwash laid down along the margin of this ice-tongue, and also in crevasses, accumulated to form a flattish-surfaced spread of sand and gravel. At a later stage the tongue of ice became detached from the main ice-mass (Fig. 6), and during the course of its melting became separated into some half-dozen small blocks corresponding to the present marshy hollows (kettles), the crevasse-fillings between these blocks surviving as ridges when the ice had completely melted. Some of the water derived from this melting escaped northward, forming a channel that breached the ring of kames.

A third type of kame topography found in East Lothian consists of chaotic assemblages of knolls, short ridges and depressions and is exemplified by an area a few hundred yards long and wide, a mile east of Kidlaw (Fig. 4). Here abundant sand and gravel buried the thin edge of the ice-mass which subsequently melted out leaving a jumble of small kames and kettles.

Kames occurring singly or in very small groups all lie parallel with former positions of the ice-margin as deduced from other evidence, and it is, therefore, presumed that they accumulated against the ice-margin or in crevasses paralleling it. An example of the latter type is provided by the mounds whose flattish tops form part of terrace I in Figure 4. Examples of the former are the small mounds of sand and angular to sub-rounded gravel that occur intermittently along the ice-contact side of a number of kame terraces in the county, rising a few feet above the flattish terrace surfaces. Such kames appear to indicate the points at which individual streams emerged from the ice or flowed off its surface.

One of the few good sections available across the greater part of a kame is shown in Figure 5. This kame is the largest of a small group lying three miles south of Gifford between altitudes of 900 and 950 feet (National Grid reference 538637). The section is exposed in a pit at the eastern end of the kame and is approximately at right angles to the length of the feature, which extends for about 300 yards. The steep northern slope (up to 20°), clearly an ice-contact face, lies some 20 yards beyond the right-hand end of the section. The kame appears to have accumulated in a lake occupying a broad opening in the ice that ran parallel with the general direction of the ice-margin. There is no

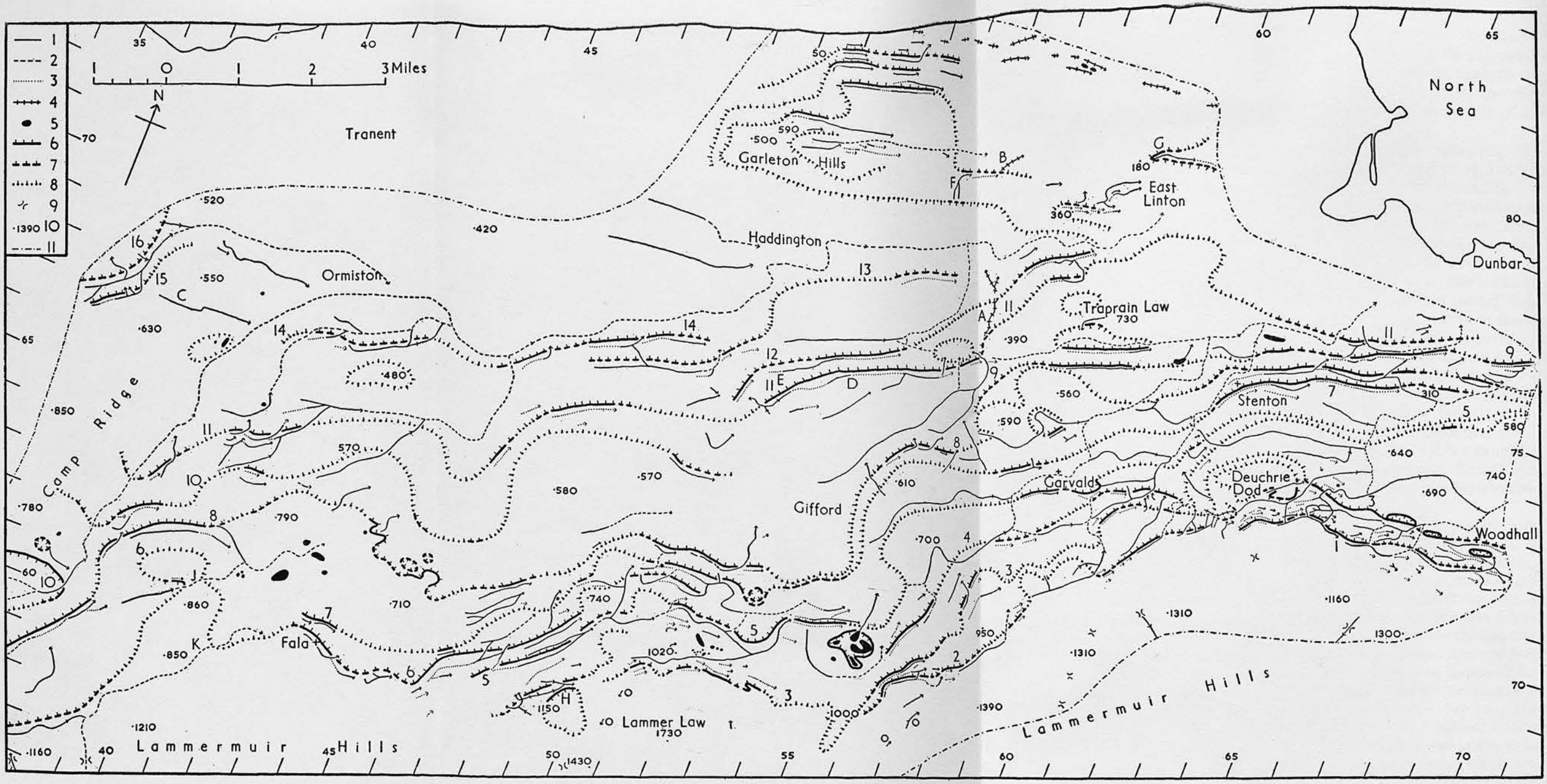


FIGURE 6—Marginal, near-marginal and sub-marginal features and retreat stages.

1, Streamless glacial drainage channels and channels containing very small streams. 2, Glacial drainage channels containing larger streams. 3, Kame terraces and ice-marginal benches. 4, Eskers. 5, Kettles. 6, Positions of ice-margin, accurate to within 50 yards. 7, Positions of ice-margin, accurate to within 200 yards. 8, Other positions of ice-margin. 9, Major cols in Lammermuir crest. 10, Summits with heights in feet. 11, Limits of the area studied. O, Channels believed to have been formed during an earlier phase of the glaciation. National Grid reference lines marked at intervals of one kilometre along borders of map; the map lies within 100-kilometre square No. 36. Ice-margin positions 9 and 10 may represent the same stage. Most of the sub-glacial chutes and other sub-glacial channels can be recognized from their relationship to the ice-margin positions.

channel or other feature marking the lake outlet which is, therefore, presumed to have been over or through the ice. Following the deposition of the sands at the bottom of the section (A, Fig. 5) considerable erosion occurred, and a bed of coarse gravel (B) was laid down. This was followed by the deposition of rhythmites consisting of thin layers of reddish clay alternating with thicker layers of brown sandy silt or yellow sand (C). The thick bed of sand (D)

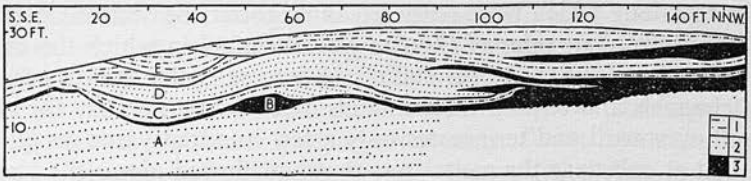


FIGURE 5—Section across a kame.
1, Alternations of sand, silt and clay. 2, Sand. 3, Gravel.

is of uniform composition and, along with the gravels into which it merges laterally, was apparently deposited by a small but powerful stream flowing into the lake from the ice. Gravels of similar origin occur at two higher levels and interdigitate with alternations of sand and silt (E). In places the gravels consist of fore-set beds dipping at about 20°. The broad flattish top of the kame suggests that the lake had been completely filled at, and west of, the line of the section by the time it was drained and deposition ceased.

Eskers

There are only a few eskers in East Lothian. Several of them occur below the 125-foot contour in the broad depression to the north of the Garleton Hills, where they form slightly sinuous lines of elongated mounds up to 40 feet high (Fig. 6). Small excavations made in the side of a railway cutting through one mound revealed current-bedded sands with thin beds of gravel. These eskers appear to represent the deposits of streams that flowed in tunnels beneath the ice as it stagnated in the depression.

An isolated esker occurs near the farm of Deuchrie on the floor of a deep valley and consists of a line of short mounds varying from a few feet to 50 feet high (A-B, Fig. 2). A small section in the easternmost mound revealed cobbles and coarse gravel, while the steepness (30°) of the northern slope of the highest mound suggests a similar composition throughout this mound. The esker appears to mark the course of a sub-glacial stream that was supplied with much of its abundant coarse load by streams rushing down the adjacent sub-glacial chutes.

The most interesting esker (A, Fig. 6), situated three miles east of Haddington, forms a ridge slightly over a mile long, 40 to 80 yards broad and up to 20 feet high winding down a north-west-facing slope. The crest is rounded and

the sides smooth with a maximum slope of 6 to 9°. The esker starts at an altitude of about 275 feet and forms a continuous even-crested ridge to about 170 feet. After a gap of about 100 yards it continues with an irregular crest and near its lower end splits into two diverging ridges, each of which terminates at about 150 feet. Sections dug to a depth of eight feet on the crest of the feature near its lower end showed that it here consists of current-bedded sand with thin beds of well-rounded fine gravel, the bedding indicating deposition by a stream flowing in a northward direction. Angular fragments of volcanic rock up to 3 inches long which were scattered throughout the sections presumably fell in from the roof or sides of the sub-glacial tunnel in which the esker was formed. The altitude of the upper end of the feature is the same as that of the series of channels and terraces that extends almost continuously for about 3½ miles south-westward and terminates only a few hundred yards away. It thus appears that at one stage the meltwaters that had flowed along the ice-margin for a considerable distance from the south-west entered a tunnel in the ice-margin and flowed down the hillside to form the esker. Subsequently other weaknesses in the ice-margin farther to the south-west were exploited and the meltwaters entered these, cutting sub-glacial chutes (Fig. 6). A much smaller example of an esker of this type, similarly related to a kame terrace, occurs three miles north-east of Haddington (B, Fig. 6).

Lake deposits

As would be expected, numerous small lakes were formed during the deglaciation of East Lothian. Most of these lakes were filled with sand and gravel (together with some silt and clay) soon after they came into existence. Many of the kames and kame terraces were deposited in part or in whole in these lakes. In some cases lake deposits are over 50 feet thick as, for example, 1½ miles north of Fala and again a similar distance south-east of Gifford; in each of these areas the approximate level of an infilled lake is represented by a partly dissected flattish surface about a mile long and several hundred yards wide that, along part of its length, terminates northward above an ice-contact slope marking the position of the ice-margin during the final stages of the accumulation of the lake deposits.

Mapping of Retreat Stages

Various stages in the retreat of the ice-margin as deduced from the features already discussed are shown in Figure 6. Since more than 900 marginal and sub-marginal features were taken into consideration in deducing these stages it is not possible to discuss them individually. Instead the method of determining the stages will be explained and certain significant aspects of the deglaciation described.

The most accurate evidence of former positions of the ice-margin is provided by well-defined ice-contact faces of kame terraces and kames; the edges of

narrow kame terraces not possessing a well-defined ice-contact face; the edges of marginal benches cut in bedrock; and small marginal channels so situated that very slight retreat of the ice-margin resulted in their being deserted by meltwaters. Positions of the ice-margin determined directly from such features are believed to be accurate to within 50 yards on the ground: in many instances the possible error is much less than this.

Many of the ice-margin positions of the second order of accuracy (those believed to be accurate to within 200 yards on the ground) are related to marginal channels that ceased to carry meltwaters before the edge of the ice had retreated more than 400 yards downslope from them, the ice-margin usually having been drawn along the middle of this zone of uncertainty. Ice-margins of this class also include those determined from the edges of the larger lake flats and those drawn between contemporaneously formed marginal features separated by distances of not more than a few hundred yards.

Some channels, especially most of the larger ones, continued to function after the ice-margin had withdrawn more than 400 yards from them, and the ice-margin positions deduced from them are shown in Figure 6 as being of the third order of accuracy. Also falling into this category are ice-margins drawn between features formed, or believed to have been formed, contemporaneously but separated by distances of more than a few hundred yards. When drawing ice-margins of this third category it was frequently necessary to determine the most probable position of the ice-margin at a given stage within a zone of considerable width (occasionally as much as a mile). In order to do this the slope of the ice-margin along its length was determined where the marginal evidence was sufficiently abundant and this slope was then extrapolated to adjacent areas where marginal phenomena are poorly developed or absent.

In making these extrapolations the fact that the slope of the ice-margin varied significantly in different localities and at different stages had to be taken into account. Although more than half the deduced ice-marginal slopes lie between 22 and 30 feet per mile, the remainder vary between approximate horizontality and more than 50 feet per mile. The great majority of the marginal features indicating gradients of the ice-margin exceeding 30 feet per mile occur where the latter lay against fairly steep slopes, especially when it lay along the scarp face of the Lammermuirs. Marginal features also indicate that in parts of East Lothian the ice-marginal slopes became less steep as the ice-mass wasted down, the evidence of this being particularly clear around Stenton and Deuchrie. Here the slopes of several well-preserved series of marginal features extending over distances of three to five miles indicate ice-marginal slopes at stages 1, 3, 7, 8, 9 and 11 of about 58, 55, 34, 32, 28 and 26 feet per mile respectively.

In other parts of the area studied, however, the tendency of the ice-marginal slopes to diminish as the ice-mass thinned was often complicated by local influences of relief. Thus when the ice-margin lay slightly above the 500-foot contour along the western slope of the Camp Ridge between Ormiston and Dalkeith a deep channel (C, Fig. 6) was cut by meltwaters draining away from

the ice into the Tyne valley. This channel descends to an altitude of 300 feet in 1½ miles and, together with related features, indicates that when it was formed the level of the ice-margin was considerably higher on the western side of the ridge than on the eastern: this suggests that the ice on the western side of the ridge was still moving whereas that on the eastern side, partly cut off from its source of supply, had become stagnant or almost so and was thinning more rapidly. In contrast, there is evidence of almost horizontal ice-margins in some areas. For example, a short distance south and south-east of Haddington two aligned series of marginal channels and kame terraces extend almost continuously for distances of 3½ and 2 miles respectively, each series being almost horizontal throughout its length (D and E, Fig. 6).

In order to take into account such variations in the slope of the ice-margin when plotting the positions of the ice-margin of the third order of accuracy Figures 7a and 7b were constructed. The latter shows most of the kame terraces, ice-marginal benches and glacial drainage channels situated between the Tyne and the Lammermuirs, together with several marginal features situated to the north of the Tyne west of Pencaitland. The construction of these diagrams was facilitated by the fact that almost all the major landforms, as well as a large proportion of the marginal features of the area are aligned between west-east and south-west–north-east. The most common direction is about S. 60° W.-N. 60° E., and this was selected as the direction of the imaginary vertical plane on to which the marginal features, the River Tyne, and the Lammermuir crest-line were projected. Minor disadvantages of the diagram are that a few marginal features running at a considerable angle to the plane of projection are greatly shortened in relation to the remainder, and that it may falsely give the impression that all the features occur on a single unidirectional slope between the Lammermuirs and the Tyne.

All ice-marginal slopes deduced directly from series of marginal features or from long marginal (or in part near-marginal) channels were plotted on the diagram. To avoid confusion these ice-marginal slopes are here reproduced separately in Figure 7a, where they are represented by the solid lines. In joining these lines together to form continuous ice-marginal slopes such additional evidence as heights of lake flats and kames was plotted on the diagram and utilized. Where such evidence was scanty or absent adjacent ice-margins whose slopes had been deduced directly from marginal and near-marginal features were taken into account. Thus, in the Stenton-Woodhall area ice-margins 3 and 7, determined from clearly preserved marginal features, slope at about 55 and 34 feet per mile respectively. On the assumption that when the ice-margin occupied intermediate positions between stages 3 and 7 it had an intermediate slope, stages 4 and 5 were continued across this area on Figure 7a (broken lines). When Figure 7a had been completed the gaps in the map of the retreat stages were filled in by abstracting from the diagram the heights of the ice-margins at different points along their lengths and plotting them on the map against the contours.[16]

An obvious disadvantage of using the diagram for this purpose is that it gives no information on the slope of the surface of the ice at right angles to its plane of projection. It is considered, however, that this disadvantage is unlikely to be serious over much the greater part of the area south of the Tyne since the ice-margins run almost parallel with the plane of projection of the diagram. Furthermore, although there appears to be no field evidence that allows the

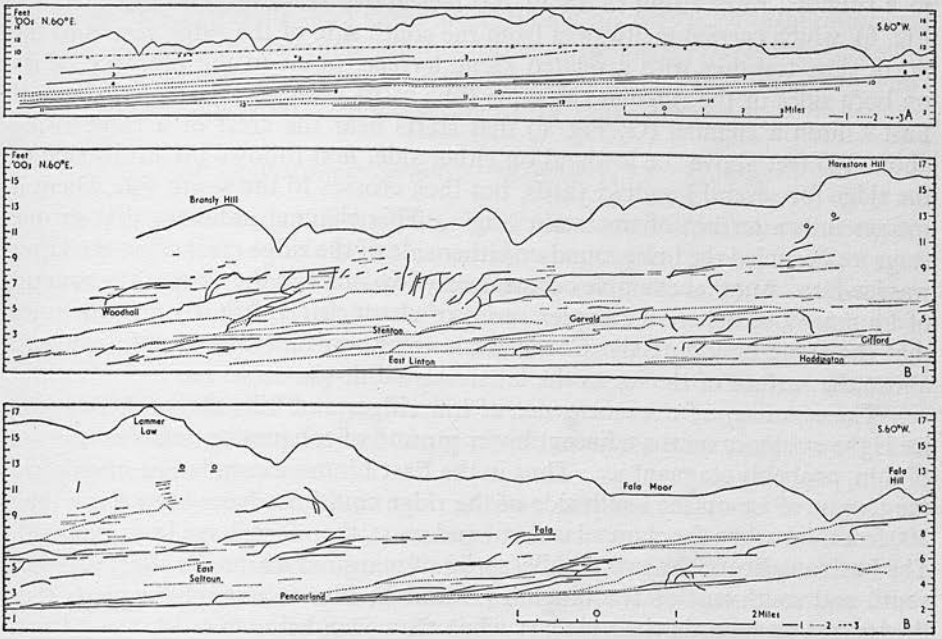


FIGURE 7—A. Slopes of the ice-margin at stages marked in Figure 6 projected on to an imaginary vertical plane aligned S. 60° W.-N. 60° E.
1, Ice-marginal slopes deduced from marginal (and in places in part near-marginal) features and from contemporaneously-formed marginal features separated by short distances. 2, Ice-marginal slopes deduced from adjacent higher and lower ice-margins as explained in text. 3, Upper limit of fresh marginal features (x) and smooth line drawn through them.
B. Marginal, near-marginal and sub-marginal channels, ice-marginal benches and kame terraces projected on to an imaginary vertical plane aligned S. 60° W.-N. 60° E.
1, Marginal channels that were abandoned by meltwaters before the ice had retreated more than 400 yards from them, and sub-glacial channels. 2, Kame terraces and ice-marginal benches. 3, Channels that continued to function after the ice-margin had retreated more than 400 yards from them. O, Channels believed to have been formed during an earlier phase of the glaciation.

slope of the ice-surface at right angles to the Lammermuirs to be estimated, the fact that there is clear evidence of the emergence of low hills through the ice as it wasted down implies that the slope was not steep, and hence the likelihood of significant errors having arisen from mapping parts of the retreat stages in this manner is reduced.

The emergence of hills through the ice as it wasted down is indicated by various lines of evidence. In the case of Deuchrie Dod abundant marginal

phenomena permit the stages of the deglaciation to be mapped in sufficient detail to prove that the higher parts of the hill became ice-free while ice was still present over its lower slopes and in the col to the south (Figs. 2 and 3). On the higher parts of the Garleton Hills several channels are so placed that they could have carried meltwaters only when the higher parts of these hills formed nunataks. About two miles to the east, where the Garleton Hills have declined to a rounded ridge rising to about 200 feet above sea-level, two channels (F, Fig. 6), which carried meltwaters from the south side of the ridge across to the north side, together with a related kame terrace, point to the presence of ice on both sides of the ridge at the same time as its crest was ice-free. North of East Linton a channel (G, Fig. 6) that starts near the crest of a ridge rising about 100 feet above the lowland on either side, first follows the north side of the ridge for several hundred yards, but then crosses to the south side where it merges into a terrace of uncertain origin. This channel indicates that at one stage ice occupied the low ground on either side of the ridge crest when the latter was ice-free. Another example occurs nearly two miles to the west of the summit of Lammer Law where a complex series of channels (H, Fig. 6) demonstrates that the ridge bounding the Cowie Burn valley on its western side emerged above the surface of the ice as the latter wasted down.[17]

The corollary of the emergence of low ridges and hills above the wasting ice is the existence on the adjacent lower ground of tongues or detached masses of thin, probably stagnant ice. Thus in the East Linton example just mentioned the tongue of ice on the south side of the ridge could not have been more than 100 feet thick when the channel was cut and must, therefore, have been stagnant. The horizontality of the two parallel series of marginal features a short distance south and south-east of Haddington (D and E, Fig. 6) strongly suggests that the ice was stagnant in their vicinity when they were being formed, a condition undoubtedly favoured by the topography, for the two series occur on the south-eastern side of a broad depression whose floor is only 100 feet below the level of the higher series. Stagnation is also indicated by the survival nearby of the esker (A, Fig. 6) winding down the hill-side from the level of the higher series of marginal features approximately at right angles to the direction of ice-movement. The ice in this depression (except perhaps during the final stages of its existence) was joined to a much larger mass occupying the basin in which Haddington lies. As noted above, the latter ice-mass, as it thinned, became separated from the main ice-mass to the north by the emerging Garleton Hills, but whether or not separation was completed by the emergence of the plateaux to the west of Haddington is uncertain. On the floor of the Tyne valley south-west of Ormiston the former existence of stagnant ice is indicated by a complex of irregular kames and depressions, while the eskers on the lowland to the north of the Garleton Hills imply thin, probably stagnant ice in their vicinity when they were being formed. These examples show that the emergence of hills and ridges above the ice and the wasting away of ice in valleys were of frequent occurrence during the deglaciation of East Lothian.

Local Glaciation of the Lammermuirs

Kendall and Bailey concluded that as the Highland ice retreated from the Lammermuirs the higher parts of these hills became an independent area of dispersal, while Charlesworth[18] refers to 'the corrie glaciers of the Lammermuirs'. The forms of these hills and their valleys, however, are not those normally associated with upland areas that were centres of ice radiation, even though on a small scale, and there are certainly no corries. Furthermore, no depositional features definitely attributable to these glaciers were recognized by Kendall and Bailey, and none has been observed since. It is, therefore, concluded that there was no local glaciation of the Lammermuirs.

The Lammermuir-Stranraer Moraine in East Lothian

Charlesworth suggested in 1926 that the Lammermuir-Stranraer moraine (or kame-moraine), which he had recently recognized, 'marks the limit of a re-advance of uncertain yet probably considerable magnitude'.[19] In East Lothian the 'moraine' includes the kames, kame terraces and related depositional features extending along the northern edge of the Lammermuirs between 550 and 1000 feet, but especially those situated between 550 and 750 feet. In agreement with Kendall and Bailey, Charlesworth regarded the features at any given altitude as having been formed more or less contemporaneously throughout the length of the Lammermuirs.

This last conclusion is seen to be incorrect when considered in relation to Figures 6 and 7a. For example, at stage 7 the ice-margin lay at an altitude of about 725 feet near Fala but only at about 225 feet south of Dunbar. *Viewed as a whole* the 'moraine' was formed successively from east-north-east to west-south-west: in the vicinity of Woodhall it was formed at stages 1 to 3, north of Lammer Law at stages 3 to 7, and south of the Camp Ridge at stages 8 to 10. The average direction of the receding ice-margin was slightly oblique to the line of the Lammermuir scarp, although in places it ran parallel with the scarp for short distances. During successive stages of the retreat abundant stratified drift was laid down in the depression between the ice-margin and the steep face of the hills, and often over the thin marginal zone of the ice itself. Melting of the buried ice and dissection by meltwater streams produced the irregular stratified drift topography that now characterizes the hill-foot zone.

Along a large part of the Lammermuir scarp there is a sharp change within a vertical interval of 100 feet or less from no marginal features at all to abundant, large and well-preserved features, the highest marginal feature of all in some places being a large kame terrace or deep marginal channel (e.g. Figs. 2 and 4). One can join most of the highest marginal features by a smooth line that, beginning at about 900 feet, at first rises gradually in a west-south-westerly direction and then levels out at about 1150 to 1200 feet (Fig. 7a).[20] Charlesworth concluded that the upper limit of his 'moraine', which approximates to this line, marks the limit of a re-advance, the ice having failed to extend beyond

the northern slopes of the Lammermuirs.[21] This view is attractive, for the fact that a smooth line can be drawn through most of the highest marginal features can then be explained, while the slope of the line can be accounted for in terms of the escape route for the ice at the south-western end of the Lammermuirs and of the much lower route around the north-eastern end. The possibility that the ice extended higher up the Lammermuir slopes (and perhaps even on to the Lammermuir plateaux in places) cannot be ruled out, however, for conditions here might have been unfavourable for the formation of recognizable marginal features. This might have been the result of one or more of several factors, including the climatic conditions prevailing at the time, the steepness of the upper slopes of the Lammermuir scarp, the limited drainage basins of streams from the ice-free area, and the existence of two deep gaps near the north-eastern end of the Lammermuirs through which meltwaters (and the ice itself) could escape to the south. It is possible that the relation between the floors of these two deep gaps and the line marking the upper limit of fresh marginal features is more than coincidental. One must, therefore, conclude that the limit of the re-advance is not yet definitely established.

The Character of the Retreat

Kendall and Bailey gave four reasons for believing that oscillations of the ice-margin occurred during the final deglaciation of East Lothian: (i) the trenching by glacial drainage channels of the abundant stratified drift deposits situated between 550 and 750 feet, which they considered to have accumulated in a lake 14 miles long overflowing by the Deuchrie-Pathhead valley; (ii) glacial drainage channels partly filled with drift; (iii) glacial drainage channels in parallel sequence whose height relationships appeared to them inexplicable unless oscillations of the ice-margin had occurred during the deglaciation; (iv) the existence of a few sections showing till interbedded with lacustrine or fluvio-glacial deposits. The best-exposed of these, described by Kendall and Bailey, shows from the surface downwards 50 feet of lacustrine sands, 10 feet of till, 20 feet of lacustrine sands, silts and clays, and finally over 30 feet of till.

The first reason need not be considered since it is invalidated by evidence already discussed (p. 70 and Fig. 6). Additional examples of the second type of evidence have been noted in this paper, including kames and deposits of till on some channel floors and small marginal channels and sub-glacial chutes on the sides of some large channels. Such features do not prove an oscillatory retreat, however, for the channels on whose floors and sides the deposits and other features occur may have been originally cut prior to the final deglaciation. In fact in most instances the width of such channels, coupled with the occurrence in various parts of East Lothian of large channels with a 'valley-in-valley' cross-profile, strongly suggests that during the final deglaciation meltwaters not infrequently took advantage of subaerially- and ice-modified channels formed during an earlier phase of the glaciation. In some instances the old channels

were only partly re-excavated, in some cut deeper, while in others stagnant ice prevented their full utilization by meltwaters.

In discussing their third type of evidence Kendall and Bailey cited as an 'excellent example' of a channel indicating an oscillatory retreat the channel lettered P in Figure 2. Detailed mapping of the retreat features shows, however, that it is unnecessary to postulate a re-advance to explain channel P (Figs. 2 and 3). In fact, all the evidence indicates that when it was formed the ice in its vicinity was completely stagnant. In other cases parallel channels that in terms of the old conceptions of ice retreat are difficult to explain without postulating oscillations of the ice-margin can be shown to have functioned, not alternately or successively, but largely simultaneously. In Figure 6 this is illustrated by the relation of channels J and K to ice-margin 6.

The sections comprising Kendall and Bailey's fourth type of evidence were carefully re-examined. No evidence inconsistent with their conclusion that these sections indicate slight oscillations of the ice-margin during the final deglaciation was found. On the other hand, the sections can equally well be interpreted in other ways: for example, while the upper till and the overlying lacustrine deposits were undoubtedly laid down during the final glaciation of the area, the lower lacustrine deposits and the till beneath may have been laid down during an earlier phase. In this context Charlesworth's remarks on the East Lothian sections and similar ones in other parts of central Scotland are significant: 'Though caution needs to be exercised in interpreting such intercalations, *since these do not necessarily imply oscillations*, in this region these structures *seem* to be of this origin'.[22] Thus Charlesworth was fully aware that Kendall and Bailey's interpretation of the East Lothian sections had not been proved. It is suggested that Kendall and Bailey's interpretation loses much of its weight when no longer supported by their other lines of evidence.

An alternative approach to the problem of the character of the retreat is to consider the extent of the evidence proving or strongly suggesting the existence of stagnant ice during the deglaciation. Not only is there clear evidence of stagnation in depressions and valleys, but such evidence also extends up the slopes of the Lammermuirs to heights of between 1100 and 1200 feet. For example, the features mapped in Figures 2 and 4, including kame terraces and kames with conspicuous (in places irregular) ice-contact faces, kettles, subglacial chutes, and parallel series of closely spaced marginal channels, together indicate that the marginal zone of the ice was stagnant when the ice lay against the Lammermuirs at an early stage of the deglaciation. Assemblages of these and other features associated with stagnant ice are found throughout East Lothian, and strongly suggest that extensive marginal stagnation occurred.[23] This conclusion is in accord with the absence of till moraines and is also supported by Figure 7a. Although caution must be exercised in interpreting this diagram, since it indicates the slope of the ice along its margin and not the slope of the free ice surface, it nevertheless strongly suggests (together with the evidence demonstrating the emergence of hills through the surface of the ice)

that, except during the earliest stages of the deglaciation, widespread thinning of the whole ice-surface occurred over East Lothian. The main body of the ice (as distinct from the marginal zone) was doubtless still moving, however, if only very slowly, and in places movement appears to have extended to the margin itself, as along the western slopes of the Camp Ridge (p. 70). The extensive and abundant evidence of thinning and marginal stagnation does not, of course, entirely exclude the possibility of occasional minor oscillations of the ice-margin but, as shown above (pp. 74-5), no proof that such oscillations occurred has yet been found. Thus the main picture that emerges is of an ice-mass that, after advancing up to an altitude of at least 1200 feet along part of the northern slopes of the Lammermuirs, became stagnant in its marginal zone. Thinning began first in the east and then gradually extended over the whole ice-mass, causing the ice-margin to retreat towards the north and north-west, hills appearing above the surface of the ice as nunataks, and detached masses of ice surviving for a time where buried beneath abundant stratified drift or where the topography was favourable.

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NOTES

[1] P. F. KENDALL and E. B. BAILEY, 'The glaciation of East Lothian south of the Garleton Hills', *Transactions of the Royal Society of Edinburgh*, 46 (1908), 1-31. Subsequent references to the work of Kendall and Bailey are to this paper.

[2] C. T. CLOUGH and others, 'The geology of East Lothian', *Memoir of the Geological Survey* (2nd edition, 1910).

[3] J. K. CHARLESWORTH, 'The re-advance, marginal kame-moraine of the south of Scotland, and some later stages of retreat', *Transactions of the Royal Society of Edinburgh*, 55 (1926), 25-50.

[4] A. GEIKIE and J. YOUNG, in 'The geology of East Lothian', *Memoir of the Geological Survey* (1866). J. GEIKIE, *Great Ice Age* (1894), 195. P. F. KENDALL and E. B. BAILEY, op. cit.; C. T. CLOUGH and others, op. cit.

[5] The terms overflow channel and spillway are not used since many of the channels were not overflows from lakes, or functioned as such only for part of the time they were in use.

[6] G. BARROW, in 'The geology of East Lothian', *Memoir of the Geological Survey* (2nd edition, 1910).

[7] The long profiles of a considerable number of channels do not decline in height continuously in a single direction, but show a distinct rise and fall.

[8] C. M. MANNERFELT, 'Några Glacialmorfologiska Formelement', *Geografiska Annaler*, 27 (1945), 1-239; 'Marginal drainage channels as indicators of the gradients of Quaternary ice caps', *Geografiska Annaler*, 31 (1949), 194-9.

[9] Cf. P. F. KENDALL, 'A system of glacier-lakes in the Cleveland Hills', *Quarterly Journal of the Geological Society of London*, 58 (1902), 471-571.

[10] G. BARROW, op. cit.

[11] R. F. FLINT, *Glacial geology and the Pleistocene epoch* (1947).

[12] C. D. HOLMES, 'Kames', *American Journal of Science*, 245 (1947), 240-9.

[13] W. D. THORNBURY, *Principles of geomorphology* (1954).

[14] C. D. HOLMES, op. cit., 248.

[15] P. F. KENDALL and E. B. BAILEY, op. cit., 14.

[16] The same methods were used in drawing ice-margins in the area to the north of the River Tyne as in drawing those to the south, including the construction of diagrams similar to Figures 7a and 7b. A tentative correlation between the two areas is suggested by ice-margin 12, Figure 6.

[17] Owing to insufficient appreciation of the significance of ice thinning Kendall and Bailey concluded that this series of channels provides definite evidence of the former existence in the Cowie Burn valley of a local glacier descending from Lammer Law.

[18] J. K. CHARLESWORTH, op. cit., 49.

[19] Ibid., 44.

[20] Four marginal channels, all of which occur in the vicinity of Lammer Law (Figs. 6 and 7b), were not used in drawing this line. These channels, which occur up to 100 feet above the line, all lack the freshness of features at lower levels and are, therefore, believed to be related to a considerably earlier phase of the glaciation than that under consideration.

[21] J. K. CHARLESWORTH, op. cit.

[22] Idem, p. 26. Present author's italics.

[23] R. F. FLINT, 'Glacier thinning during deglaciation: part II, Glacier thinning inferred from geologic data', *American Journal of Science*, 240 (1942), 113-36; W. D. THORNBURY, op. cit.; C. M. MANNERFELT (1945), op. cit.

SUB-GLACIAL STREAM EROSION IN SOUTHERN NORTHUMBERLAND

J. B. Sissons

A few years ago Peel [1] drew attention to the fact that the floors of two glacial drainage channels in southern Northumberland (Beldon Cleugh and the East Dipton channel) do not decline in height continuously in a single direction but exhibit a marked rise and fall. As yet no satisfactory explanation of the origin of these channels has been put forward. It is the purpose of this paper to suggest such an explanation. As will be seen the importance of this explanation is not only that it accounts for the anomalous features of the two channels but that it provides the key to an interpretation of the mode of deglaciation of southern Northumberland differing considerably from the orthodox interpretation of Dwerryhouse [2], Anderson [3] and Raistrick [4].

Only a brief description of the channels is necessary here since they have been described in detail and excellently illustrated already by Peel [5]. The Beldon channel (Fig. I) is over a mile long, steep-sided, has a maximum depth exceeding 150 feet, and is drained only by tiny streams. The northern part is floored with peat and has a winding course. The evidence from borings made by Peel leaves little doubt that the bedrock floor of the channel has a marked up-and-down profile: commencing at an altitude of about 1,070 feet at the northern end, the bedrock floor rises to at least 1,110 feet about half-way along the channel, before descending to about 1,000 feet at the point where the channel runs into the trench occupied by the Beldon Burn.

The East Dipton channel (Fig. 2) is about three miles long, steep-sided, has a maximum depth exceeding 150 feet, almost lacks peat, and is drained by small streams. The solid floor, which at the western end is at an altitude of 390 feet, rises gradually to a maximum altitude of about 410 feet after half a mile before descending to about 190 feet where the channel becomes continuous with the trench occupied by the March Burn [6].

Both channels cut across the spur to the east of the Devil's Water, a south-bank tributary of the middle Tyne. Numerous other channels occur on this and other spurs to the south of the Tyne. In Peel's words, "Many of these channels fall into aligned sequences at levels steadily declining eastwards, and all the evidence suggests that they were cut to link series of lakes impounded in the valleys of the northward-draining Tyne tributaries by the edge of the Tyne ice-sheet as it retreated north. This interpretation was advanced for the western part of the region in 1902 by Dwerryhouse. More recent work by Anderson in the eastern part of the area confirms and supports this view [7]". Yet despite this definite statement regarding the deglaciation of the area, which is further supported by Raistrick [8], the evidence is far from conclusive.

This evidence is best considered in relation to Dwerryhouse's map [9] of glacial drainage channels and former ice-dammed lakes in the area south of the middle Tyne reproduced in Figure 3. His is the only detailed

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published study of the area in which the Beldon and East Dipton channels actually occur. It may be noted that, although Dwerryhouse dealt with the whole of the Alston Block, this is the only detailed map of retreat stages and glacial drainage channels he provided, and it is thus logical to assume that he considered the evidence for deducing the existence of former ice-dammed lakes most satisfactory in this area. The following criticisms of Dwerryhouse's evidence as shown on this map may be made: (i) channel B, supposed to have carried water from lake A to lake C, does not descend to the level of lake C; (ii) the humped bedrock profile of channel E (Beldon Cleugh) is unexplained;

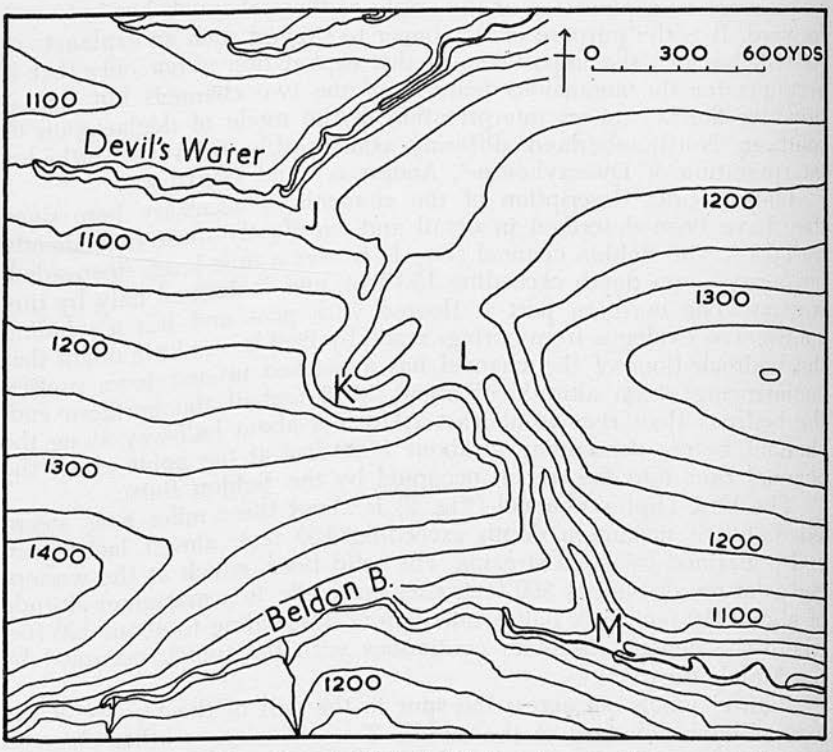


Fig. 1 The Beldon channel (J, K, L, M). Based on Peel, *op. cit.*, 1951.

(iii) no channel is shown draining into lake F; (iv) no continuous channel draining from lake F to H is shown; (v) channel G is shown as terminating well above the level of the lake into which it is supposed to drain. Thus the ice-margin positions drawn by Dwerryhouse do not appear to be consistent with his own evidence. The lakes do not constitute evidence, of course, since he deduced their existence and extent from the channels.

In addition to these detailed criticisms a more important difficulty is encountered when the relative rates of retreat of the ice-margin at different points along its length are considered. According to Dwerryhouse's map the ice must have been far thicker in the valleys than on

the intervening spurs. (For example at Stage I the ice would be some 600 feet thicker in the East Allendale valley than on the spur immediately to the west). It is extremely difficult to see how this very thick ice in the valleys melted back at approximately the same rate as the much thinner ice on the spurs in order to maintain the approximately-parallel retreat of fairly-straight ice-margins mapped by Dwerryhouse, even when allowance is made for more rapid melting of the ice where it was in contact with lake waters. Even if his Stage I position of the ice-margin is correctly mapped (and the evidence above suggests it is not) the ice, as it thinned, would inevitably become closely related to

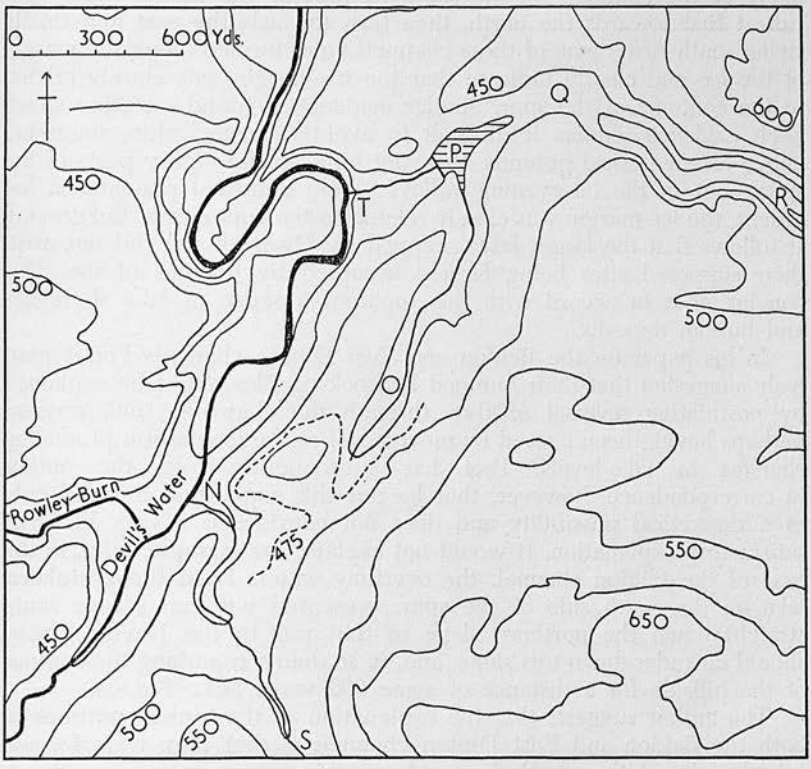


Fig. 2 The western part of the East Dipton channel (T, P, Q, R), its tributary channel (N, O, P) and the western part of the Strothers Dale channel (S). Based on Peel, *op. cit.*, 1951, and Ordnance Survey.

the contours, tongues of ice lying in the valleys and the intervening spurs rising above the surface of the ice.

The Stage I position of the ice-margin also can be criticised on theoretical grounds. Dwerryhouse stated that this stage marks the maximum extent of the ice. It follows that, on his interpretation, at least in the final stages of the advance to the maximum, the ice pushed up the valleys as a vertical or near-vertical wall hundreds of feet high while the much thinner ice over the intervening spurs advanced at approximately the same rate. This appears highly unlikely and is not

in accord with observations on present-day glaciers. One would expect tongues of ice to have advanced up the valleys. One must therefore conclude that the positions of the ice-margin mapped by Dwerryhouse are not justified by his own evidence, while they appear unlikely on theoretical grounds.

Partly in order to test the validity of Dwerryhouse's principal evidence, namely the glacial drainage channels, the author re-mapped some of these channels. This showed that many of the channels Dwerryhouse had mapped as more or less straight trenches across spurs [10] in fact run round the slopes of these spurs. Thus Figure 4 shows that many of the channels on the southern part of the Eshell Moor spur run at first towards the north, then turn towards the east and finally swing south-east. Some of these channels were formed along the margin of the ice and clearly indicate that the ice-margin was closely related to the contours of the spur. Similar evidence is found on other spurs. Such evidence renders it difficult to avoid the conclusion, suggested above on theoretical grounds, that, not only on the higher parts of the spurs, but in the intervening valleys where marginal phenomena are absent, the ice-margin was closely related to the contours of the ground. It follows that the larger lakes mapped by Dwerryhouse did not exist, their supposed sites being largely occupied by tongues of ice. This conclusion is in accord with the apparent absence of lake shorelines and bottom deposits.

In his paper on the Beldon and East Dipton channels Peel tentatively suggested that their humped bedrock profiles might be explained by postulating reversal of flow through the channels, such reversal perhaps having been caused by movements of the ice-margin producing changes in lake-levels. Peel has subsequently told the author in correspondence, however, that he put this suggestion forward only as a theoretical possibility and does not consider it a very likely or satisfactory explanation. It would not explain, for example, why, in the case of the Beldon channel, the overflow waters from the postulated lake on the south side of the spur, presented with an escape route straight down the northern slope of the spur to the Devil's Water, should meander down this slope, and, in so doing, run along the contour of the hillside for a distance of some 600 yards (KL, Fig. 1).

The author suggests that the explanation of the 'uphill' portions of both the Beldon and East Dipton channels is that they were formed sub-glacially. Although Peel considered this interpretation unsatisfactory in his 1951 paper [11] he appears more favourably disposed towards it in a more recent paper [12], and states that Mannerfelt "strongly advocated" it to him, although at the same time he (Peel) mentions several difficulties that it encounters.

Reasons have already been given for believing that ice, not a lake, occupied the valley of the upper Devil's Water at the time the Beldon channel was formed. The channel itself provides further evidence for this view, for the lake hypothesis, in addition to failing to explain (i) the humped bedrock floor of the channel, does not account for (ii) the 600-yard section of the channel running along the south side of the Devil's Water valley, (iii) the fact that the channel commences at the edge of the incised course of the Devil's Water, (iv) the fact

that the channel leaves the incised course of the Devil's Water exactly at the point where the present stream abruptly changes direction through 45 degrees, and (v) the feeble development of the northern side of the channel for the first 300 yards of its course.

Section KL of the channel (Fig. 1) could have been formed only when ice occupied the valley, the meltwaters being forced to flow along the contour of the hillside between the ice to the north and the rising ground to the south. Therefore KL was initiated either as a marginal or sub-glacial channel. The waters that cut KL crossed the pre-existing

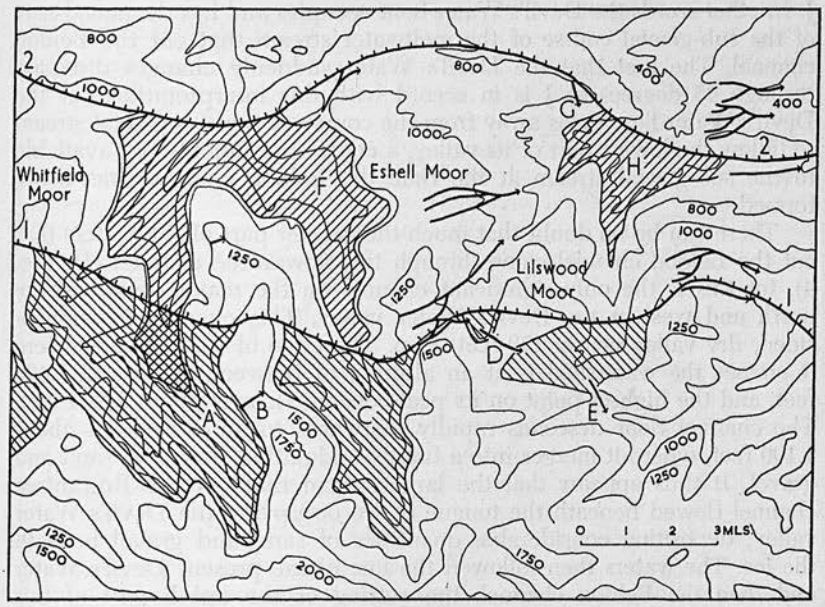


Fig. 3 Map of West Allendale (A), East Allendale (C) and the Devil's Water valley (H) showing ice-dammed lakes (shaded), glacial drainage channels and two positions of the ice-margin, according to Dwerryhouse *(op. cit.)*. D, Rowantree channel. E, Beldon Cleugh. Other letters explained in the text.

col at its lowest point and then flowed straight down the south side of the spur into the Beldon Burn valley, and thence along that valley and into the Derwent valley, cutting the deep trench described by Dwerryhouse and Peel. As the ice on the north side of the spur thinned and its edge retreated down the slope, KL functioned as a marginal channel (regardless of whether or not it was *initiated* as a marginal channel) and both KL and LM were cut deeper.

The waters that cut KLM must have flowed along the channel from J, for JM is one continuous feature, and at K, where one might expect to find an important marginal channel leading into and continuous with KL, no such channel exists. Therefore, whether KL functioned as a marginal channel throughout the time it was in use or was initiated as a sub-glacial feature, JK was certainly entirely formed sub-glacially. This is in accord with other evidence, for there appears to be no other

way in which the southward rise in the bedrock floor of the channel from J to K could be explained. The feeble development of the northern side of the channel for the first 300 yards of its course can also be explained: since there are no grounds for postulating the removal by 'post-glacial' erosion of much of this side of the channel one is forced to conclude that the missing upper part of the channel side was ice (*i.e.* the northern side of a sub-glacial tunnel).

The fact that the channel starts at the very edge of the incised course of the Devil's Water strongly suggests that the waters that cut the channel flowed along the line of the Devil's Water immediately above J. In other words the Devil's Water here occupies and has deepened part of the sub-glacial course of the meltwater stream that cut the Beldon channel. The fact that the Devil's Water suddenly changes direction through 45 degrees at J is in accord with this interpretation, for the Devil's Water here turns away from the course of the sub-glacial stream to follow the lowest part of its valley, a course presumably not available to the sub-glacial stream at the time the Beldon channel was being formed.

There can be no doubt that much the greater part of the waters that cut the Beldon channel came through the Rowantree channel (U, Fig. 4), for this is the only significant channel on the major divide to the north and west of the Devil's Water valley. This channel is a steep-sided, dry valley nearly 150 feet deep. The edge of the channel where it crosses the watershed is at an altitude of between 1,350 and 1,400 feet, and the highest point on its peat-covered floor is about 1,260 feet. The channel floor descends rapidly eastwards to an altitude of about 1,100 feet, where it merges into a faintly-undulating spread of sand and gravel. It thus appears that the large stream that cut the Rowantree channel flowed beneath the tongue of ice occupying the Devil's Water valley, depositing considerable quantities of sand and gravel beneath the ice. The waters then followed the line of the present Devil's Water and then the Beldon channel, the cutting of the uphill part of this channel being possible owing to the considerable hydrostatic pressure under which the waters were flowing.

The East Dipton channel, like the Beldon channel, commences at the edge of the incised course of the Devil's Water and leaves that stream at a sharp bend in the latter's course (Fig. 2). In fact in this case the connection between the course of the Devil's Water and that of the channel is even more obvious, for not only does the Devil's Water head straight for the channel entrance before twisting away in two incised meanders, but the channel "commences as a wide flat-floored trench cut down at least 60 feet *below* the general floor-level of the old valley, and opening out directly from the side of the.... trench of the Devil's Water [13]". There can thus be no doubt that, at least in the final stages of the cutting of the East Dipton channel, the meltwaters followed the present incised course of the Devil's Water before turning away at T (Fig. 2) to follow the channel. The solid floor of the channel rises eastwards some 20 feet between T and Q. Since a surface stream could not cut such an uphill channel one is forced to the conclusion that this part of the channel at least was formed sub-glacially. It follows that for some distance upstream from T the 100-feet-deep trench of the

Devil's Water also was initiated and cut to a considerable depth by sub-glacial stream erosion.

Unlike the Beldon channel the East Dipton channel is joined by a tributary channel (NOP, Fig. 2). This channel starts at the edge of the Devil's Water trench as a broad, shallow, dry hollow, some 10 to 20 feet deep, floored with sand and gravel. It deepens gradually northwards to its junction with the East Dipton channel, where it is some 40 feet deep and its floor is at an altitude of 395 feet. That a considerable part of the waters that cut this channel came down the Rowley

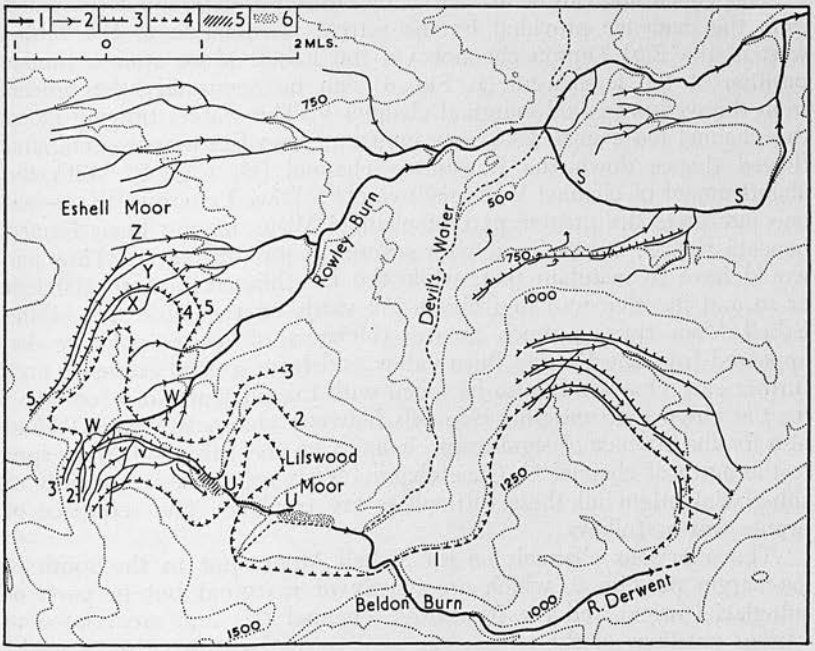


Fig. 4 The Beldon and East Dipton channels in relation to adjacent glacial drainage channels. 1. routes followed by large volumes of meltwater for considerable periods; 2. other glacial drainage channels; 3. positions of the ice-margin accurately mapped from marginal features; 4. other positions of the ice-margin; 5. site of former lake; 6. area of sand and gravel situated between channel U and the Beldon channel; S. Strothers Dale channel; U. Rowantree channel; V-Z other channels referred to in the text.

Ice-margin position 1 is for the most part deduced from the relationship between channel U and the Beldon channel and is to be regarded as provisional. It is possible that the Beldon Burn valley was filled with ice at this stage, in which case the deep trench that the burn occupies would have been cut sub-glacially.

Burn valley is strongly suggested by the precise alignment of the lower part of this valley with the upper part of the channel. It might appear from Figure 2 that channel NOP was formed along the edge of an ice-tongue projecting southwards into the Devil's Water valley. This explanation would be inadequate, however, for it would mean that on entering the East Dipton channel the meltwaters from NOP flowed uphill for several hundred yards (from P to Q). Furthermore it would

fail to explain how the East Dipton channel was initiated, for the altitude of the floor of the col into which this channel was cut was approximately 500 feet (unless one assumes, without evidence, a pre-existing channel in the floor of the col). Therefore, in order to develop the necessary hydrostatic head to permit the initiation by a sub-glacial stream flowing uphill of the western part of the East Dipton channel, the ice-margin at that time must have been situated a considerable distance to the south and south-west, in which case the whole of channel NOP and at least the eastern part of the Rowley Burn trench are of sub-glacial origin.

This conclusion can be arrived at by entirely independent reasoning from the evidence provided by the retreat phenomena to the south-west of the East Dipton channel. On the Eshell Moor spur a former position of the ice-margin (2, Fig. 4) can be accurately determined from the well-preserved marginal channel V. The waters flowing along this channel fed a small lake, now indicated by a flat marshy area, and flowed thence down the Rowantree channel (U, Fig. 4). With the abandonment of channel V the meltwaters followed channel W. Unless one interprets the greater part of channel W as having been formed beneath the ice one is faced with several major difficulties. Thus one would have to maintain that, while the ice thinned a mere 100 feet or so and its edge retreated about 250 yards on the west side of the Eshell Moor spur, a much greater thickness of ice completely disappeared from the Rowley Burn valley and from a very extensive area further east. One would also be faced with the difficulty of accounting for the absence of marginal channels between channels V and W and also for the absence of significant channels on the Lilswood Moor spur to the north of channel U. Once channel W is recognised as mainly of sub-glacial origin all these difficulties are resolved. The sequence of events was as follows.

The numerous channels on the Eshell Moor spur to the south of ice-margin position 2, which are mostly of marginal but in parts of sub-glacial origin, fed the Rowantree channel (U) and are related to various positions of the ice-margin as it retreated down the slopes of the spur. The last channel (V) to feed channel U was formed along ice-margin 2. While V was being formed increasing quantities of meltwater had found their way beneath the ice-margin, particularly along the three channels that join the western part of W. Eventually channel V was completely abandoned (and consequently channel U also) and all the meltwaters coming from the west followed channel W, disappearing beneath the ice occupying the Rowley Burn valley. Meanwhile the ice on the west side of the spur had retreated from position 2 to position 3. The absence of marginal channels between V and W is thus accounted for.

With continued retreat of the ice-margin channel W was eventually abandoned and the meltwaters crossing the Eshell Moor spur followed channels X, Y and Z, at successively lower levels to the north. As in the case of channel W, the waters flowing through these channels flowed beneath the ice occupying the Rowley Burn valley, for all four channels are deep trenches continuous with the Rowley Burn trench, which is in turn continuous with the Devil's Water trench. This implies

that a major sub-glacial stream followed the Rowley Burn to its junction with the Devil's Water, a conclusion supported by the absence of significant channels on the Lilswood Moor spur to the north of channel U. This also accords with the suggestion made above that the alignment of the upper part of channel NOP (Fig. 2) with the lower end of the Rowley Burn trench implies that most of the waters that cut channel NOP had come under the ice down this trench.

The sub-glacial Rowley Burn did not at first continue into channel NOP (Fig. 2), however. Instead it followed the Strothers Dale channel (S, Fig. 4). This channel, mapped in part by Dwerryhouse [14], resembles the East Dipton and Beldon channels in possessing an up-and-down profile (the "downstream" rise of its present floor is about 100 feet), but differs in that it is broader and less deeply cut (80 feet at most). Significantly the south-eastward-directed part of the first few hundred yards of its course as it ascends the east side of the Devil's Water valley is exactly aligned with the point at which the Rowley Burn enters the Devil's Water valley from the west. It thus appears that the sub-glacial waters descending the Rowley Burn valley first cut channel S, secondly channel NOP and its continuation as section PQR of the East Dipton channel, and finally the upper 50 or 60 feet of the Devil's Water trench (NT) and section TP of the East Dipton channel.

Thus the simple sequence of events described by Dwerryhouse [15] whereby the channels on the spur to the east of the Devil's Water were cut successively from south to north as the overflows of large ice-dammed lakes is incorrect. The three up-and-down channels were cut in order from south to north by sub-glacial waters, but independently of the other (smaller) channels. The large size of the three up-and-down channels clearly indicates that large volumes of water flowed through them for a considerable period of time. This is consistent with their mode of evolution as outlined above. The other channels on the spur, however, are much smaller: most of them start as faint benches or very shallow channels and increase in depth gradually towards the east. These channels are mostly marginal (and, in places, sub-glacial) channels. Their relatively small size is due to their waters being derived locally and to the short time they were in use, while their eastward increase in depth is largely due to the increased quantities of water available for their excavation in this direction (principally from melting of the adjacent ice). This is in direct contrast with the three sub-glacial channels which are of large dimensions throughout, having been cut by large volumes of water not derived from their own immediate vicinity.

Probable criticisms of the interpretation suggested above have been indicated by Peel [16], who states: "Many problems are of course involved in such a concept including the ability of sub-glacial streams to erode large channels in solid rock, the possibility of their eroding uphill under hydrostatic head, and perhaps even the possibility of major concentrated meltwater streams flowing at all under thick land ice..."

Although little attention has been given in Britain to erosion by sub-glacial streams its importance has long been recognised on the Continent. Mannerfelt [17] and others have described and mapped numerous valleys cut by waters that flowed from ice-free ground

beneath the ice where the latter terminated against steeply-rising ground. Examples abound in Britain but appear not to have been described as yet. The writer has mapped many channels of this type in East Lothian, where they range up to 100 feet in depth and are usually cut in solid rock [18].

The tunnel valleys of eastern Denmark (tunneldale) and north Germany (Rinnentäler) are regarded by most authorities as having been formed by sub-glacial streams. Thus Schou [19] states that the larger tunnel valleys in Denmark are up to 75 km. long and over a kilometre wide, while the surrounding glacial landscape often lies 100 metres above the valley bottom. Regarding their origin Schou [20] says: "These valleys were formed by the erosion of the meltwater under the ice-cap. In tunnels which they formed under the ice the enormous masses of water moved with such great erosive force that they were able to cut the tremendous sub-glacial gutters which now appear in the form of large valley systems or transgressed by the sea as in the case of the East Jutland fiords. The great thickness of the ice-cap and the resulting high watertable in the ice subjected the meltwater in the tunnels to high pressure, so that locally within these enclosed ducts it ran uphill. This phenomenon is also observable in present-day glaciers, where in some cases the meltwater has been seen to emerge from the 'gates' like a fountain, having run obliquely upwards along the final stretch." These large sub-glacial valleys greatly exceed in size the channels in Northumberland that the author suggests are of sub-glacial origin. In particular it may be noted that the 'downstream' (westward) rise in the floors of the sub-glacial valleys in Denmark sometimes approaches 100 metres, a figure greatly exceeding that in the Northumbrian examples.

Evidence of sub-glacial stream flow and its relation to ice-dammed lakes in areas at present undergoing glaciation is in accord with the deductions made concerning the Northumbrian channels. Proof that sub-glacial streams are eroding their beds, particularly that they are in some instances eroding when flowing uphill under hydrostatic pressure, is obviously difficult to obtain. However, Okko has stated [21] that in Iceland, at least in the case of Hoffelsjökull, Heinabergsjökull, Breidamerskurjökull and Skeidararjökull, sub-glacial waters "must flow uphill in order to emerge from the ice" and therefore "must.... flow under hydrostatic pressure". Ahlmann [22], Arnborg [23], Thorarinsson [24] and Jónsson [25] have made similar observations regarding streams issuing from the base of the ice in this area. Arnborg states that in these conditions "the sub-glacial rivers may have a great capacity and eroding force" [26], while Jónsson says: "There is no doubt that the ice-river does a tremendous erosive work beneath the ice where there is no rock to stop it. When the river has eroded off the loose material which is the bed, the ice must, of course, sink down and on account of the hydrostatic pressure, the ice-river may continue its erosion deeper down according as the thickness of the ice above" [27].

It should also be mentioned that the draining off of ice-dammed lakes by streams flowing beneath the ice is a common phenomenon in Iceland. The opening of a tunnel beneath the ice during the summer may rapidly empty an ice-dammed lake, the waters re-appearing at the snout of the glacier several miles away producing sudden floods.

Arnborg [28], using his own evidence and Thorarinsson's researches [29], suggests that, as the ice thins over the course of years, these floods probably become less intense (owing to the sub-glacial outlet being opened earlier in the summer), eventually ceasing completely when the sub-glacial tunnel exists as a permanent feature.

The explanation advanced above for the origin of the Beldon and East Dipton channels may appear unsatisfactory in that it requires sub-glacial streams to have followed valley floors and then suddenly to have abandoned them in order to climb valley sides and flow through pre-existing cols. It must therefore be pointed out that such sub-glacial stream courses are quite common, for in some areas the evidence of eskers clearly shows that sub-glacial streams often followed such routes. For example, Mikkola [30] mentions five esker systems that cross the mountain belt of northern Lapland "availing themselves of the gaps and valleys which interrupt it". He continues: "Thus they follow the general rule of adjusting themselves to the larger topographic features." A similar generalisation is made by Flint [31], who states: "..... an esker may climb up a valley for several miles, pass over a low divide, and descend the far slope. Crossings of divides are invariably at low saddles or gaps, and at such a crossing the esker is likely to be notably discontinuous". One esker in northern Lapland discussed in detail by Mikkola [32] follows the valley of the Sotajoki river for a considerable distance before leaving it at an abrupt bend (cf. Beldon and East Dipton). The esker then climbs 100 metres up the valley side to near the crest of the interfluve, where it is continued by a "zone of conspicuous erosion phenomena", which extends down to the floor of the next major river valley. Examples in Britain have been mapped by F. Synge, who kindly allows the author to quote from his unpublished work. One esker on the Aberdeenshire/Kincardineshire border, mapped by Synge, is several miles long and crosses two spurs on either side of the Water of Aven, a south-bank tributary of the eastward-flowing Water of Feugh. Commencing near the Feugh at about 400 feet the esker rises eastwards to 450 feet before descending below 400 feet where it has been breached by the Aven. It then climbs again and ends slightly above 450 feet in a meltwater channel cut into the floor of a pre-existing col. The esker commences again at the lower end of the channel and descends to near the floor of the Burn of Strathy, which it follows to the confluence of the Strathy with the Feugh. More striking examples have been mapped by Synge in Leix County in central Ireland. Here the Stradbally and Maryborough eskers climb steadily towards hilly country from altitudes of between 250 and 300 feet to altitudes of between 400 and 500 feet. Each esker makes for a pre-existing col where it is replaced by signs of meltwater erosion that merge into a well-developed meltwater channel on the further side of each col.

It is suggested that these notes on sub-glacial drainage in Iceland and in areas formerly ice-covered, especially as evidenced by eskers and related erosional features, coupled with the detailed evidence from southern Northumberland, leave little doubt as to the former importance of sub-glacial drainage in the latter area.

It is not proposed to consider here the application of the conclusions reached above to other parts of Britain. The author ventures to

suggest, however, that much that has been written in Britain on glacial retreat stages as deduced from glacial drainage channels is in need of considerable revision. It may be added that the author is not alone in making this suggestion for orthodox views have long been challenged by Carruthers [33] for very different reasons, while Peel [34] has rightly pointed out that in Britain "thought has been perhaps unduly influenced by Kendall's concept of freely-draining lakes despite the common and disturbing lack of any independent direct evidence for such lakes".

The author wishes to thank Professor R. F. Peel for discussing this paper in detail with him and for making several valuable suggestions. He would also like to thank Mr F. M. Synge for permitting him to quote from unpublished work.

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SUPPOSED ICE-DAMMED LAKES IN BRITAIN WITH PARTICULAR REFERENCE TO THE EDDLESTON VALLEY, SOUTHERN SCOTLAND

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Since the publication of Kendall's well-known paper in 1902 it has been widely accepted that the withdrawal of glacier ice from higher to lower ground caused small ice-dammed lakes to be formed in valleys in many parts of upland Britain, these lakes overflowing from one to another across spur crests, resulting in the formation of channels that are now streamless or contain misfit streams. In the last half century or so numerous papers have been published in which such former lakes are described and their former extent often shown on maps, and it might appear that the validity of Kendall's original hypothesis has been demonstrated many times. Nevertheless, it is the purpose of this paper to suggest that his hypothesis requires considerable modification.

It will be suggested that the sites of many postulated ice-dammed lakes were probably largely occupied by glacier ice at the times the lakes are supposed to have existed, and that the channels on the spurs between are usually not "overflow" channels from these lakes, but were cut by streams supplied mainly by water held up within, or flowing through, beneath, and/or over the ice that occupied the valleys between the spurs. In arriving at this interpretation the author has been strongly influenced by the publications of Scandinavian workers and in particular wishes to acknowledge his indebtedness to papers by Mannerfelt (1945, 1949).

The first part of this paper is concerned with demonstrating how the suggested interpretation applies to a selected area, namely the Eddleston valley and its immediate vicinity, situated some 12 to 19 miles south of Edinburgh. The relief of much of this area was ideal for the development of ice-dammed lakes draining from one to another across spur crests, but the numerous meltwater channels and the abundant deposits of sand and gravel show that no such simple pattern developed. Instead this evidence shows that large volumes of meltwater flowed through and beneath the ice, as well as along its margin. In the second part of the paper certain major difficulties encountered by the orthodox ice-dammed lake hypothesis will be considered and it will be suggested that the author's interpretation of the Eddleston valley evidence (and of that of other areas) may be applicable to considerable areas of Britain.

The Deglaciation of the Eddleston Valley Area

Principal physical features. The southern part of this area comprises a portion of the Southern Uplands, which here rise to maximum altitudes of 1,500 to 2,000 feet. The

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northern part is lower, reaching about 1,250 feet in the west and just under 1,000 feet in the east; here the relief is dominated by several broad ridges and hollows aligned approximately N. 30° E.—S. 30° W. The amplitude of these undulations is several hundred feet in the west but diminishes towards the east to 50 feet or less.

The greater part of the Eddleston valley runs almost due north and south and through it the Eddleston Water flows southwards to join the Tweed. North of the Southern Upland

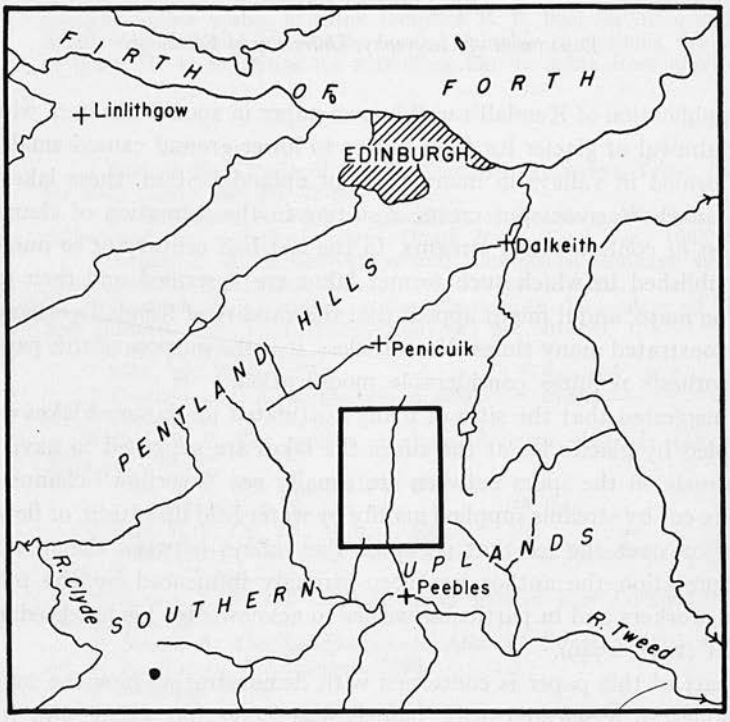


Fig. 1. Location of the Eddleston valley area.

edge the valley is fairly shallow, its floor lying only 100 to 150 feet below the general level of the surrounding country. Within the Southern Uplands it becomes a major feature, the hills on either side rising 900 or more feet above the level of the valley floor. This rise is not abrupt, however, for the Eddleston Water occupies a marked "valley within a valley". The inner valley (which term will be used hereafter) is generally steep-sided and some 200 to 300 feet deep. Between its upper edge and the final steep rise to the hill summits a number of benches occur, by far the most conspicuous one being situated on the western side of the valley between altitudes of about 900 and 1,100 feet, producing a marked contrast between the two sides of the valley. This contrast is related to the contrast in tributary valley development, for the western side of the valley lacks major tributary valleys whereas the eastern side possesses several deep, steep-sided valleys (Fig. 2).

For convenience of description that part of the Eddleston valley lying north of the Southern Upland edge will be subsequently referred to as the "upper Eddleston valley". The section south of the southern limit of the main mass of sand and gravel (Fig. 2) will be termed the "lower Eddleston valley". The part between will be called the "middle Eddleston valley".

Bedrock geology. The bedrock of almost the whole of the region consists of intensely-folded Ordovician rocks, mainly shales, mudstones and greywackes, striking between NE.—SW. and NNE.—SSW. Many of the valleys and ridges are closely related to this trend. In the extreme north-east and north-west gently-dipping Carboniferous rocks are present, mainly limestones and shales.

Previous literature. No detailed account of the glaciation of the Eddleston valley has been published. The drift boundaries in the area are shown of Geological Survey maps, but no memoir is available. Charlesworth (1926; 1957, p. 1204) considered the sand and gravel accumulations in the northern part of the Eddleston valley to be part of his Lammermuir-Stranraer kame-moraine, deposited along the margin of the Highland ice. Eckford (1952), however, concluded that similar deposits in the West Linton-Dolphinton area, a few miles to the west of the Eddleston valley, were laid down in relation to the Southern Upland ice.

Fieldwork. Prior to fieldwork the region was studied stereoscopically on vertical aerial photographs on a scale of approximately 1 : 10,000, details of retreat forms and superficial deposits, wherever identifiable, being transferred to sheets 36/25 and 36/24 of the Ordnance Survey 1 : 25,000 map. Drift boundaries, copied from Geological Survey maps, were also marked on these sheets. The map sheets were then taken into the field and corrections and numerous additions were made on a transparent overlay, the region being covered systematically one kilometre square at a time. In addition a map of the upper Eddleston valley on a scale of 1 : 4,000 was made.

Marginal and near-marginal channels. Many marginal channels exist in the region and they often occur in sequences one below the other on hillsides. They are especially numerous to the west of the Eddleston Water, owing partly to the favourable topography. The great majority of these channels are small features, generally less than 15 feet deep and no more than 20 to 30 yards wide. The smallest ones mapped are narrow benches, often only about 10 yards wide, cut by water flowing in channels one side of which was ice and the other bedrock or drift. On some slopes (e.g. 42[1]) very faint benches a few yards wide occur, but it was not possible to map these on the scale adopted.

The distribution of the channels shows that, as the ice wasted down, small ridges and hills gradually appeared above its surface. This is well illustrated, for example, by the channels on both sides of ridge 13. The downwasting of ice in valley 15 is demonstrated by the channels on both its north-western and south-eastern slopes. The emergence of the small hill 18 is clearly shown by the marginal channels that occur on all sides of it. In the

[1] All numbers referring to specific features are marked on Fig. 2, where they are arranged in order from west to east along one-kilometre strips starting in the north-west corner of the map.



Fig. 2. Glacial geomorphology of the Eddleston valley area. 1, Major meltwater channel. 2, Channel cut by meltwaters produced by melting of detached mass of ice. 3, Other meltwater channel. 4, Highest point on up-and-down channel. 5, Esker. 6, Kettle. 7. Major ice-contact slope. 8, Kame ridge. 9, Conical kame. 10, Contour. 11, Flat-topped kame or kame terrace. 12, Sand and gravel; symbol omitted in areas with symbol 11. 13, Peat. 14, Alluvium. 15, Water body. Till ommitted to simplify map. Drift geology modified after Geological Survey. Contours omitted where superficial deposits marked. National Grid lines at one kilometre intervals shown around borders of map.

extreme south the small hill 54, which is only about 100 feet high and has an area of only one-tenth of a square mile, is grooved by some 15 channels, which are mainly marginal but partly subglacial in origin. These channels permit the emergence of the hill above the ice surface to be traced in detail (Figs. 2 and 3).

It can be frequently shown that wastage of the ice occurred contemporaneously on both sides of a ridge. For example, channel 19 begins on the south-eastern side of ridge 18 and could have been formed only when ice was present on this side of the ridge. Having crossed the crest of the ridge, the channel does not run straight down the slope towards the valley floor but turns sharply to descend obliquely for some 200 yards. It could have been cut in this position only if ice was present on the north-western side of the ridge as well as on the south-eastern side when it was being formed.

Almost all the 150 or so marginal channels mapped in the area fall in height towards the NE. or NNE., indicating that the margin of the ice, and hence the free ice surface also, sloped down in this general direction. This in turn strongly suggests that the last glacier ice to occupy this region came from the Southern Uplands. Movement of ice in a north-easterly direction is indicated by the only two striae known in the region (Geological Survey One-Inch Map). It is also suggested by eskers 6 and 7, for it is generally agreed that the longest eskers in a region parallel the latest direction of ice-movement. This accords with the conclusions of Pringle (1948), Eckford (1952) and others, but not with those of Charlesworth (1926; 1957, p. 1204).

The marginal channels enable one to map accurately former positions of the ice-margin for short distances at numerous places to the west of the Eddleston Water. However, the ice-margin positions mapped from different channels cannot be correlated by *direct* field evidence one with another so as to delimit the extent of the ice over considerable areas at specific times. Correlation was therefore made by the indirect method of calculating the approximate gradient of the ice-margin from the gradients of the marginal channels, a method used by Mannerfelt (1945).

These gradients vary between about 1 in 15 and about 1 in 50. This considerable range is the result of most of the marginal channels having been cut by meltwaters that flowed along the ice-margin for short distances before flowing back on to, in to or beneath the ice. Thus many channels after running a few tens or hundreds of yards along a hillside terminate suddenly, their lower ends being left "hanging"; in such channels the lower end is usually the point at which the meltwaters flowed back in to or on to the ice. In other instances channels turn sharply through about 90 degrees to run straight down a hillside, indicating that the marginally-flowing meltwaters suddenly turned to flow beneath the ice and continued their erosive action there. Yet other channels, after running along a hillside at only a very slight angle with the contours, turn through a small angle to run more steeply, but still obliquely, down the slope, making a larger angle with the contours of the hillside; many of these channels were formed by waters that flowed first along the ice-margin and then obliquely down the slope beneath the ice. Channels of this last type are frequently joined by lower marginal channels formed later. Most of the marginal channels shown in Fig. 2 are related to subglacial channels or to former en-

glacial or supraglacial streams in one of these three ways. Consequently, the lower (downstream) part of a marginal channel does not necessarily reflect the gradient of the ice-margin, for a gradient steeper than that of the ice-margin was frequently developed here by headward erosion resulting from the increased velocity of the marginally-flowing meltwaters towards the point where they began to descend rapidly into or beneath the ice. This steepening of the long profile towards the lower end is seen in many marginal channels, and, as would be expected, is especially conspicuous in marginal channels that are continued by subglacial channels that plunge straight down hillsides. It follows that the gentlest channel profiles give the best approximation to the gradient of the former ice-margin. Such channels are generally the longer ones, for steepening towards the lower end has usually affected only a small part of the long profiles of these channels. Therefore, since the least-inclined marginal channels have gradients of about 1 in 50 and the great majority of the channels fall in height towards the NE. and NNE., this was assumed to be the approximate gradient of the ice-margin in this direction and the ice-margin positions mapped directly from different marginal channels were correlated one with another using this gradient. The results are shown in Fig. 3, along with ice-margin positions deduced from other evidence discussed later.

Many of the marginal channels are spaced at regular vertical intervals. Extensive sequences of such regularly-spaced channels in northern Sweden have been studied by Mannerfelt (1945, 1949) and Hoppe (1957). Mannerfelt's conclusions as to their origin, which are accepted by Hoppe (1950), and which the present author believes satisfactorily explain the similar channels of the Eddleston valley, are as follows (1949, p. 196). "The lateral drainage channels were formed mainly in the spring and early summer by the intensified melting of the snow. Sub-glacial drainage tunnels opened later in the summer. Through these the water was then drained off as the summer ablation lowered the surface of the ice, gradually leaving the lateral gullies of the spring flood dry. During next year's melting period another gully was eroded at a lower level. Wherever these gullies occur in large numbers, they are invariably so regularly spaced that they must be assumed to reflect the annual ablation, seeing that hardly any other climatological periodicity is conceivable. If that assumption is correct, the annual thinning by ablation must have averaged 3.5—5 m, which seem likely values if compared with the recent observations in glaciated districts." In the Eddleston valley area the best sequences of sub-parallel marginal channels are numbered 14, 25, 43 and 45 in Fig. 2. In order from north to south these sequences comprise 10 channels in a vertical interval of 110 feet, 10 channels in 90 feet, 14 channels in 150 feet and 7 channels in 60 feet, giving average vertical spacings between successive channels in the four sequences of approximately 12, 10, 11 $^1/_2$ and 10 feet respectively. It is suggested that this regular spacing indicates that the ice in the Eddleston valley area was reduced by thinning at an average rate of 10 to 12 feet per year (except where buried beneath sand and gravel, as discussed below). If, as suggested above, the ice surface sloped towards the NE. and NNE. at about 1 in 50, marginal retreat towards the SW. and SSW. occurred at an average rate of about 500 to 600 feet per year on level ground.

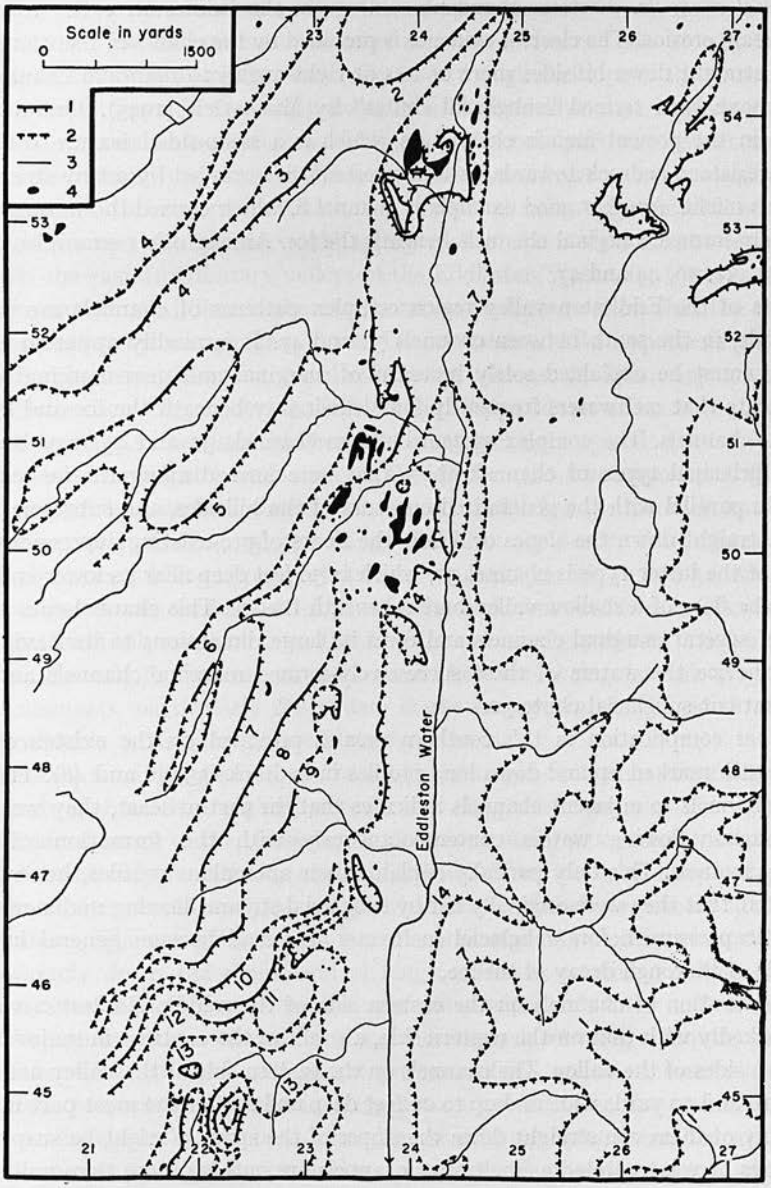


Fig. 3. Ice-margin positions in the Eddleston valley. 1, Ice-margin position accurately fixed by marginal feature. 2, Other ice-margin position. Owing to the scarcity of marginal features on the eastern side of the Eddleston valley ice-margin positions 1, 7 and 14 on that side of the valley have been deduced from the slope of the ice-margin as determined from the evidence on the western side of the valley. 3, Present stream. 4, Position of small detached mass of ice indicated by kettle.

Subglacial channels. There is abundant evidence in the Eddleston valley area of subglacial stream erosion. The clearest evidence is provided by the channels, mentioned above, that run straight down hillsides more or less at right-angles to marginal channels. Such channels have been termed "subglacial chutes" by Mannerfelt (1945). One of the best examples in the present area is channel 12, which is a steep-sided feature cut some 40 feet into resistant bedrock towards its lower end and is occupied by a tiny stream that is an obvious misfit. Another good example is channel 8, which carried the waters of several successively-formed marginal channels beneath the ice. Among other examples are channels 11, 31, 32, 39, 44 and 47.

In parts of the Eddleston valley region complex patterns of channels are found, as, for example, in the south between channels 32 and 47. It is readily apparent that such patterns cannot be explained solely in terms of marginal and near-marginal drainage: they indicate that meltwaters frequently found their way beneath the ice and excavated subglacial channels. The complex system between channels 32 and 47 may be resolved into two principal types of channel: those that were formed along the ice-margin and run nearly parallel with the generalized contours of the hillsides, and subglacial channels that run straight down the slopes or follow the floors of pre-existing depressions. A good example of the latter type is channel 31, which is 50 feet deep near its lower end and was cut into the floor of a shallow valley buried beneath the ice. This channel cuts across the courses of several marginal channels and owes its large dimensions to its having carried beneath the ice the waters of these successively-formed marginal channels and also the waters that cut subglacial chute 32.

A further complication in this southern area is provided by the existence of three channels with marked up-and-down long profiles in bedrock (33, 40 and 48). The relation of these channels to adjacent channels indicates that, in part at least, they were utilized by subglacially-flowing waters contemporaneously with the formation of adjacent marginal channels. This only partially explains their anomalous profiles, however, and it is suspected that they were originally cut by subglacial streams flowing under considerable hydrostatic pressure before subglacial meltwater flow had become general in the area as a result of thorough decay of the ice.

The distribution of channels on the eastern side of the middle Eddleston valley contrasts markedly with that on the western side, a result of the contrast in major landforms on the two sides of the valley. The channels on the eastern side of the valley are generally between 10 and 30 yards wide and up to 20 feet deep and are for the most part in bedrock. Since many of them run straight down the slopes of the spurs it might be suspected that the features here interpreted as meltwater channels are nothing more than gullies formed since the deglaciation, for such gullies are abundant in the Southern Uplands (Tivy, 1957). Attention is therefore drawn to the following facts. (i) Almost all the channels are streamless and their floors and the lower parts of their sides are rounded owing to weathering and mass-movement. (ii) The channels are restricted to the northern and north-western slopes of the spurs and are especially common immediately to the north and north-west of cols in these spurs. (iii) Some of the channels commence on the crest-lines of spurs and

most of them, unlike normal stream valleys, are of approximately-constant size throughout their lengths. (iv) In a considerable number of cases a channel falls steeply down the slope of a spur but then terminates suddenly, even though the slope of the spur continues to descend steeply towards the valley floor. It is inconceivable that such channels were cut by normal streams whose waters suddenly ceased to erode part way down steep slopes.

This last point might appear to indicate that the channels were cut as overflows from ice-dammed lakes, the abrupt termination of a channel indicating the level of the lake into which it drained. Conditions would appear to have been ideal for the formation of such lakes in the eastern tributary valleys of the Eddleston Water, for these valleys form deep embayments in the hills and there is abundant evidence to show that the Eddleston valley was occupied by a large mass of ice. The ice-dammed lake hypothesis provides an inadequate explanation, however, for the following reasons. (i) There appears to be no evidence of lake shorelines or bottom deposits. (ii) Deltas do not occur at the lower ends of the channels. (iii) The heads of the channels on any particular spur do not fall in altitude from east to west as might be expected on the ice-dammed lake hypothesis. (iv) If the channels were cut by overflow waters from ice-dammed lakes the channels cut into the floors of pre-existing cols (24, 35, 37 and 53) should be far deeper than other channels. This is not the case. (v) The upper parts of four channels (22, 23, 30 and 52) are cut into the *western* slopes of cols, that is, into slopes that descend from west to east. These channels cannot be satisfactorily explained on the ice-dammed lake hypothesis. They show that ice occupied the cols to the east of them when they were being formed. (vi) Complex systems of channels, particularly the system in the embayment to the north of col 53, appear inexplicable on the ice-dammed lake hypothesis.

It is suggested that the channels are of subglacial origin except for short segments at the upper ends of some channels that were formed along the ice-margin. This is indicated by the position of many of the channels, for they are usually so situated that they could have been formed only in intimate contact with ice. That ice must have occupied the cols to the east of channels 22, 23, 30 and 52 when these channels were being cut has been mentioned above. The complex channel system north of col 53 consists of short elements running obliquely down the slopes and of longer elements running straight down the slopes. These two types of channel are continuous with each other and were clearly formed contemporaneously. Therefore, since the channels running obliquely down the slopes could have been formed only in the presence of ice, one must conclude that the whole channel complex was formed in intimate contact with ice, the channels running straight down the slopes having been formed by streams plunging down beneath the ice. Another example is provided by channel 51, which is cut into the summit of a gentle rise on the crest of the southernmost spur. The ground slopes down from the edge of the channel both to the east and to the west, so that when the channel was formed ice must have been present on both sides of it. Channel 36 on the central spur is a more striking example since it runs for about 300 yards along the crest of a minor side-spur (shown by the contours in Fig. 2). Again, ice must have been present on both sides of the channel when it was being formed: the ice, in fact, formed the walls of a subglacial tunnel. That

many other channels on the eastern side of the middle Eddleston valley could have been formed only in intimate contact with ice can be deduced from their relation to the contour as shown in Fig. 2.

Since all the channels are found on the northern and north-western slopes of spurs it is clear that the flow of meltwaters in this region was consistently from south to north. Furthermore, since many of the channels commence exactly on the crest lines of the spurs it is also clear that these east-west spurs functioned as barriers to the free flow of water from south to north. It has already been shown that the channels cannot be interpreted as former overflows from ice-dammed lakes, and it is suggested that the following explanation is the logical consequence of the preceding facts and deductions.

Ice covered the region to a minimum altitude of 1,500 feet above present sea-level (the altitude of the highest channel). As the ice decayed it became extensively broken by numerous openings, through which water flowed freely. This free flow of water was hampered, however, by the three east-west spurs buried or partly buried beneath the ice, so that considerable volumes of water were held up within the ice on the southern side of each spur. This water escaped northwards down the northern slope of each spur cutting the numerous channels. In some instances the headward segments of the channels were cut by waters flowing along or just beneath the ice-margin, but these waters soon plunged beneath the ice to cut subglacial channels. Other channels were entirely formed subglacially. Several of the channels descended some 250 to 300 feet beneath the ice, strongly suggesting that at least the upper 300 feet of the ice was in a much decayed condition when they were cut. Theoretically the altitude of the lower end of a subglacial channel might mark the point below which subglacially-flowing waters were unable to penetrate owing to lack of openings in the ice. Alternatively it might mark the upper limit of saturation within the ice controlled by a spur to the north buried or partly buried beneath the ice. The latter explanation is considered the more likely and evidence in favour of it will be considered later.

It will be observed that the hypothesis outlined above is in some respects similar to the conventional ice-dammed lake hypothesis: the channels are, in a sense, the overflows from englacial lakes impounded by spurs buried or partly buried beneath the ice. The use of the word "lake" is rather misleading, however, and it is suggested that the upper limit of saturation within the ice should be referred to as the "englacial water-table". (Mannerfelt [1949, p. 194] used the term "ground water level".) Above the englacial water-table water descended more or less freely through the ice and streams flowing down the slopes of the buried spurs were able to cut subglacial channels. Below the level of the water-table erosion was impossible: instead, where the topography of the land buried beneath the ice was favourable and abundant material was being transported by the meltwater streams, kames or eskers were formed, as on the western side of the Eddleston valley (see below).

The altitude of the lower limit of meltwater erosion. The lower ends of many channels and channel sequences lie between altitudes of 850 and 950 feet. Along the western side of the upper Eddleston valley channels 1, 2, 4, 9 and 17 all terminate at or slightly below

the 900-foot contour. Channel 16 is utilized by the small stream draining valley 15, but its fashioning by meltwaters is revealed by its steep sides, flat floor and slightly-winding course (composed of large-radius curves that could have been cut only by large volumes of water), all of which features suddenly terminate at an altitude of about 880 feet. Downstream from this point the valley is much narrower, has no flat floor, and the stream tumbles down a series of rapids. These features indicate that the section above about 880 feet is a meltwater channel whose excavation was controlled by the same general base-level as controlled the cutting of other channels on the western side of upper Eddleston valley.

Further south on the western side of the Eddleston valley many channel sequences terminate between altitudes of about 850 and 925 feet. Only two large channels, whose significance is discussed later, continue below 850 feet. On the eastern side of the valley no channel descends below 850 feet and most of the channels on the southernmost spur and on the western part of the central spur terminate between altitudes of 850 and 950 feet.

This widespread evidence of a lower limit to meltwater erosion might appear to suggest that an ice-dammed lake with a surface level between about 850 and 950 feet existed in the Eddleston valley. It would appear reasonable to expect that such a lake would come into being and increase in area as the ice that occupied the valley retreated towards the south. The sequence of events was by no means so simple, however. It has been concluded above, for example, that the channels on the eastern side of the middle Eddleston valley did not drain into an ice-dammed lake but were for the most part cut subglacially. Much additional evidence of the sequence of drainage changes is provided by depositional features, to the consideration of which we may now turn.

Depositional features. Deposits of sand and gravel are very abundant along the western side of the inner valley of the Eddleston Water, but are of very limited extent on its eastern side except in the extreme north. This distribution implies that the sand and gravel were carried into the area by the meltwater streams that cut the channels to the west of the Eddleston Water. This is confirmed by the frequent abrupt termination of channels exactly at the upper edge of the depositional area and by several sections in gravel pits where the bedding consistently indicates flow of water towards the north-east or north.

The abundant sand and gravel deposits do not continue into the lower Eddleston valley, however, for they stop abruptly at 50, Fig. 2. Instead they extend continuously (apart from two very short breaks due to subsequent channel-cutting) into the Meldon valley, where they are concentrated on the western side. This distribution shows, as do the abundant channels at higher levels, that the ice occupying the Meldon and Eddleston valleys prevented meltwaters escaping to the Tweed valley and deflected them northwards.

The sand and gravel deposits have an extremely irregular surface form. There are numerous kames and kettles as well as several eskers and kame terraces. The abundance of ice-contact forms shows that the sand and gravel were deposited, not in an open lake, but in intimate association with dead ice.

In the northern part of the Eddleston valley nearly all the kames are flat-topped or possess very gently-undulating tops. When viewed from certain positions the tops of the

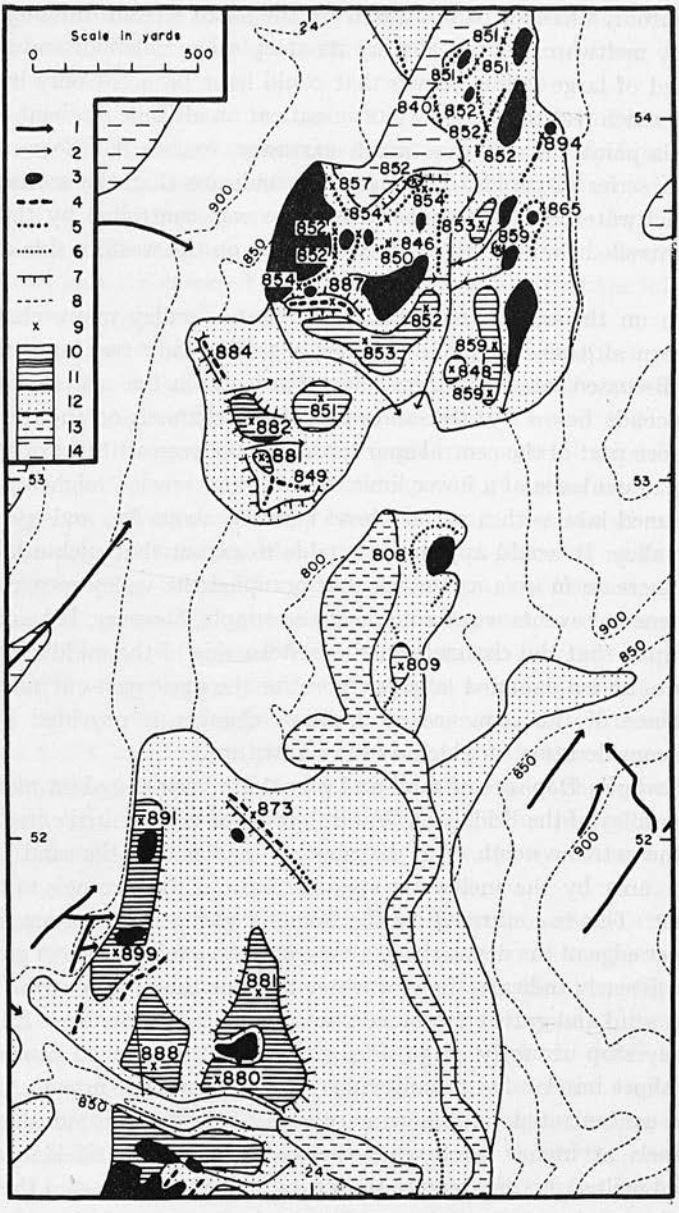


Fig. 4. Glacial geomorphology of the upper Eddleston valley. 2, Channel cut by meltwaters produced by melting of detached mass of ice. 1, Other meltwater channel. 3, Kettle. 4, Kame ridge related to the higher level of aggradation. 5, Kame ridge related to the lower level of aggradation. 6, Other kame ridge. 7, Major ice-contact slope. 8, Contour. 9, Aneroid station with altitude in feet. 10, Kame terrace or flat-topped kame related to higher level of aggradation. 11, Kame terrace or flat-topped kame related to lower level of aggradation. 12, Sand and gravel; symbol omitted in areas with symbols 10 and 11. 13, Alluvium. 14, Peat. Drift geology modified after Geological Survey.

kames appear to fall into two distinct altitude groupings. Since such impressions are liable to be misleading the altitudes of the kames were determined with an aneroid, readings being taken on the highest point of each kame, on each of the approximately-flat areas of sand and gravel that occur on the flanks of a few kames and, in the case of kames with gently-undulating crest lines, on the top of each slight rise. The altitude data obtained are shown in Figs. 4 and 5. The existence of two main levels of deposition—between altitudes of about 880 and 900 feet and between about 850 and 860 feet—is clearly revealed by Fig. 5.

The upper level of deposition is best represented in the south-western part of the area shown in Fig. 4. Here the surface of a considerable area of sand and gravel lies between 880 and 900 feet except where pitted by kettles and partly-closed depressions. This feature is bounded along much of its eastern edge by a steep ice-contact slope that accurately delimits the western edge of the main mass of ice occupying the Eddleston valley at the end of the period during which the feature accumulated. That this material was not derived (except possibly to a very minor degree) from the ice that occupied the valley, but was instead washed in by the streams that cut the meltwater channels to the west and south-west, is suggested by its large volume, and is proved by the bedding exposed in gravel pits in two of the kames, which demonstrates flow of waters towards the north-east.

The remaining kames built up to the 880—900-foot level of deposition are of limited extent (Fig. 4). They are all largely surrounded by steep ice-contact slopes, which indicate that they accumulated in crevasses in the ice that occupied the valley. The close correspondence in altitude of the kame crests implies that the openings in which they were formed penetrated down to the base of the ice: had they accumulated on top of ice they would have been lowered to different levels when the ice melted. The fact that these kames, despite the intervening ice, were built up to the same level as the spread of sand and gravel a half mile or more to the south indicates that the ice was extensively decayed, with water flowing freely through and over it.

Numerous kames related to the lower level of aggradation are concentrated in the extreme northern part of the Eddleston valley (Fig. 4). The southern members of this kame concentration are up to 30 feet high, but the northernmost ones are merely low mounds that rise a few feet above the surface of an extensive peat flat. The tops of the kames, however, all lie between altitudes of 840 and 865 feet and the great majority are between 850 and 860 feet. Their ice-contact slopes indicate that they were mostly deposited in a large irregularly-shaped opening that had formed in the ice as a result of the enlargement and merging together of the small openings in which the kames built up to the 880—900-foot level had been deposited. One may therefore visualize the upper Eddleston valley at the end of the period of aggradation represented by the lower group of kames as containing irregularly-shaped masses of ice surrounding and surrounded by, and probably often buried by, a sheet of sand and gravel built up to an altitude of between 850 and 860 feet. The melting of these masses of ice at a later date gave rise to the numerous kettles and partly-closed depressions that characterize the northern part of the Eddleston valley.

The water produced by the melting of this ice often cut small channels, thus further modifying the original smooth-surfaced spread of sand and gravel.

The large volumes of meltwater that entered the upper Eddleston valley from the west and south-west during the period that the two levels of deposition operated escaped north-eastwards via cols 3 and 5 (Fig. 2), which are the lowest points in the watershed surrounding the Eddleston valley. The present floor of the more northerly col is at an altitude of about

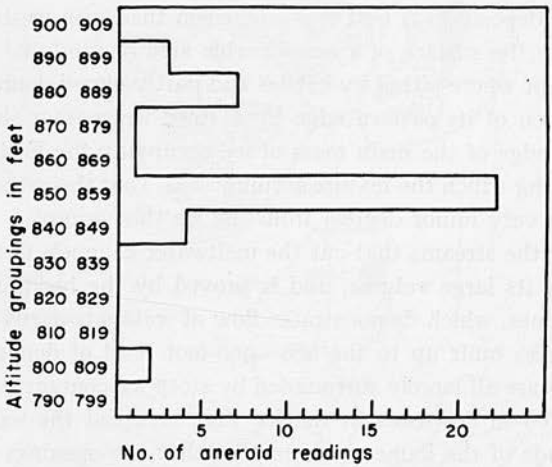


Fig. 5. Kame and kame terrace altitudes in the northern part of the Eddleston valley.

890 feet and that of the more southerly at about 865 feet, the bedrock floor in each case being a few feet lower. The more northerly col provided the escape route for meltwaters from the upper Eddleston valley during the period of aggradation to the 880—900-foot level and the more southerly one during the 850—860-foot stage.

In the *middle* Eddleston valley the most conspicuous depositional feature is the extensive mass of sand and gravel that occupies its northern part. Sections exposed in a pit show over 100 feet of these deposits, the bedding indicating deposition by northward-flowing waters. Although the surface of the deposits is extremely irregular, forming a complex of kames and kettles, most of the higher points lie between 850 and 870 feet, indicating that the upper limit of deposition was controlled by col 5. The southern edge of this mass of sand and gravel is a steep ice-contact slope up to 50 feet high (21). This ice-contact slope probably indicates the position of the northern edge of the main mass of ice in the Eddleston valley at the end of the 850—860-foot period of aggradation. At this time many detached masses of ice still existed further north in the valley (Fig. 6) and their subsequent melting helped to produce the present irregular topography.

To the south of ice-contact slope 21 the sand and gravel deposits, although much thinner, mantle almost the whole of the steep western slope of the inner valley and form a large number of kames and other ice-contact forms. That these deposits accumulated mainly as a result of material transported by the meltwater streams that cut the channels at higher levels being washed onto, into and beneath the ice that was stagnating in the deep, steep-sided inner valley is strongly suggested by the fact that many of the kames run up and down the steep slope of the valley side, and is clearly demonstrated by the eskers. Esker 41 is a steep-sided, gently-winding ridge, sufficiently conspicuous to be named Kaim Rig on the Ordnance Survey 1 : 25,000 map. Since it begins exactly at the lower end of a meltwater channel there can be little doubt that the stream that cut the channel also deposited the esker. The change-over from erosion to deposition occurs at an altitude of between 860 and 870 feet. A similar relationship exists about a mile and a half to the north, where the stream that cut channel 26 changed over to deposition at an altitude of between 890 and 900 feet, forming eskers 27 and 28. A fourth esker (34), for which no feeding channel was found, starts at an altitude of about 870 feet. These four eskers demonstrate that waters flowed into the ice stagnating in the deep inner valley of the Eddleston Water. The fact that their upper ends lie within the altitude range 860—870 to 890—900 feet indicates that, at the times of their formation, the upper limit of deposition (alternatively, the base-level for erosion) was controlled by one or other of the two outlets to the east of the upper Eddleston valley. Since this control could have operated only through the ice itself, the ice must have been extensively decayed with water flowing freely through it. The englacial water-table was unable to fall below the altitude of the floor of col 5, and the level at which it intersected the steep valley side at any particular time was the lowest level at which erosion was possible and the upper limit of abundant deposition.

It is not proposed to discuss in detail the complex relations of deposition and erosion on the western side of the middle Eddleston valley. Briefly, the upper limit of the extensive sand and gravel area, which generally lies between 850 and 925 feet, was controlled by cols 3 and 5. Correlation of particular features or groups of features with one or other of the two levels of aggradation in the upper Eddleston valley, is generally uncertain, however. In general, the upper limit of abundant deposition was lowered owing to col 5 replacing col 3 as the outlet, and probably owing to the meltwaters being able to flow with increasing freedom through the ice as decay progressed.

Reference has already been made to the fact that on the eastern side of the middle Eddleston valley none of the channels descends below 850 feet, while most of the channels on the southernmost spur and on the western part of the middle spur end between 850 and 950 feet. Reasons have been given for believing that these channels were, for the most part, formed subglacially. It is a logical step from the discussion in the last paragraph to conclude that the lower limit of channel cutting on the eastern side of the Eddleston valley, as on the western side, was determined by the level of the englacial water-table controlled by cols 3 and 5. This conclusion also follows from the earlier discussion of the channels on the eastern side of the valley, as a result of which it was suggested that the

spurs buried or partly buried beneath the ice controlled the level of the water-table to the south of them. The water-table between about 850 and 950 feet controlled by cols 3 and 5 was thus the final one of a series that had existed at lower and lower levels as the ice decayed.

Several of the channels on the eastern side of the middle Eddleston valley that end between 850 and 950 feet begin on the spur crests at altitudes of around 1,150 to 1,200 feet, showing that when they were cut the surface of the ice in their immediate vicinity was not lower than 1,150 to 1,200 feet. This indicates that the ice was so deeply decayed that streams were able to plunge down the slopes of the buried spurs as much as 300 feet below its surface to the englacial water-table controlled by cols 3 and 5 situated some 4 to 6 miles to the north.

The establishment of drainage to the south in the Eddleston valley. In the preceding paragraphs the extensive decay of the ice has been demonstrated and it may therefore appear somewhat contradictory that it has been implied that at the same time the ice in the Eddleston valley prevented meltwaters escaping southwards to the Tweed valley. This apparent contradiction may be resolved by considering probable ice thicknesses at a particular time. For example, as shown above, cols 3 and 5 began to influence the lower limit of subglacial channel cutting on the central spur on the eastern side of the middle Eddleston valley when the ice surface there was situated at an altitude of at least 1,200 feet. If the surface of the ice rose towards the SW. and SSW. at a gradient of about 1 in 50, as suggested on p. 7, the altitude of the ice surface over the present site of Peebles at the junction of the Tweed and Eddleston valleys would have approached 1,500 feet at this time, and its thickness over the valley floors hereabouts would have been nearly 1,000 feet. One would expect such a mass of ice to have formed an effective barrier to the passage of meltwaters down the Eddleston valley into the Tweed valley, especially as the eastern edge of the ice occupying the latter valley must have been situated a considerable distance to the east of the junction of the two valleys. However, as time progressed, thinning would lower the surface of the ice and at the same time the zone of decay would penetrate deeper and deeper into the ice, until eventually there would come a time when the ice could no longer prevent the meltwaters in the Eddleston valley escaping down the valley to the Tweed valley. Col 5 would then cease to function as the escape route for the meltwaters in the Eddleston valley.

The evidence is entirely consistent with this deduction and indicates that the establishment of a subglacial escape route resulted in the rapid draining of the middle and upper Eddleston valley. The abundance of flat-topped kames and kames with accordant summit levels in the northern part of the valley has been referred to above. South of the major ice-contact slope 21 and below an altitude of about 850 feet, however, kames of this type are absent and there are no terrace-like spreads of sand and gravel on the valley sides. Many of the kames have gently-rounded outlines, while steep ice-contact slopes, although present in places, are far less common than in areas where the kame crests rise to altitudes of 850 feet or more. Furthermore, below an altitude of 800 feet kettles are completely absent except for one very shallow example. These features indicate that after the aban-

donment of the 850—860-foot level of aggradation no lower levels of aggradation operated in the Eddleston valley: the water held up in the ice was drained away and the englacial water-table ceased to exist. The kames below about 850 feet owe their more subdued and generally-rounded forms to their constituent materials having been let down onto the ground during the melting of a considerable thickness of ice after deposition in their vicinity had ceased.

Proof that large volumes of water flowed beneath the ice in the Eddleston valley is provided by channel 49. This large channel, which is cut about 80 feet into bedrock near its lower end, terminates at an altitude of about 660 feet, only 50 feet above the present valley floor. The lower end of the channel merges into a sand and gravel spread that slopes continuously down to present valley floor level. This channel was able to function later than adjacent channels because it carried water from the ice decaying in the Meldon valley (Fig. 6). That ice at least 150 feet thick was still present in the middle Eddleston valley when the channel was initiated, is shown by the turn towards the north that the channel makes near its lower end. Here only a narrow ridge of bedrock veneered with drift separates the channel from the steep 150-foot descent to the floor of the Eddleston valley. The meltwaters could have been prevented from flowing straight down this steep slope only by a mass of ice in the Eddleston valley.

One other channel continued to function after drainage to the south had been established. This channel (20) descends to the present floor level of the Eddleston valley, which here lies at about 730 feet. It continued to operate after the surrounding channels had been abandoned because it carried meltwaters from the ice in valley 38, outside the area considered in this paper.

In Fig. 6 an attempt has been made to portray conditions in the Eddleston valley shortly after col 5 had been abandoned and southward drainage had been established. Most of the evidence on which this map is based has already been discussed, but two points require explanation. (i) That detached masses of ice still existed to the north of the margin of the main ice-mass at this time is shown by the small channels that breach the kettle rims. Most of these channels are cut below 850 feet and show that the melting out of most of the detached masses to form kettles occurred after the 850—860-foot water level in the valley had ceased to operate. Ice was able to survive in the inner valley of the Eddleston Water for a considerable time after surrounding higher ground had been deglaciated owing to its being buried by the abundant sand and gravel. Where the ice in the upper Eddleston valley was not extensively covered with these deposits it melted out before drainage to the south was established causing areas of open water to develop as it did so. The largest such area, now floored with a thin layer of silty material, is marked 10 in Fig. 2. (ii) The precise course of the stream that drained southwards beneath the ice in the Eddleston valley could not be ascertained owing partly to the later deposition of alluvium. It is reasonable to assume that it followed the eastern side of the present valley floor since the ice would have been thinner there owing to differential ablation caused partly by aspect but mainly by the abundant drift on the western side of the valley. This appears to be confirmed at one point on the eastern side of the valley (29) where

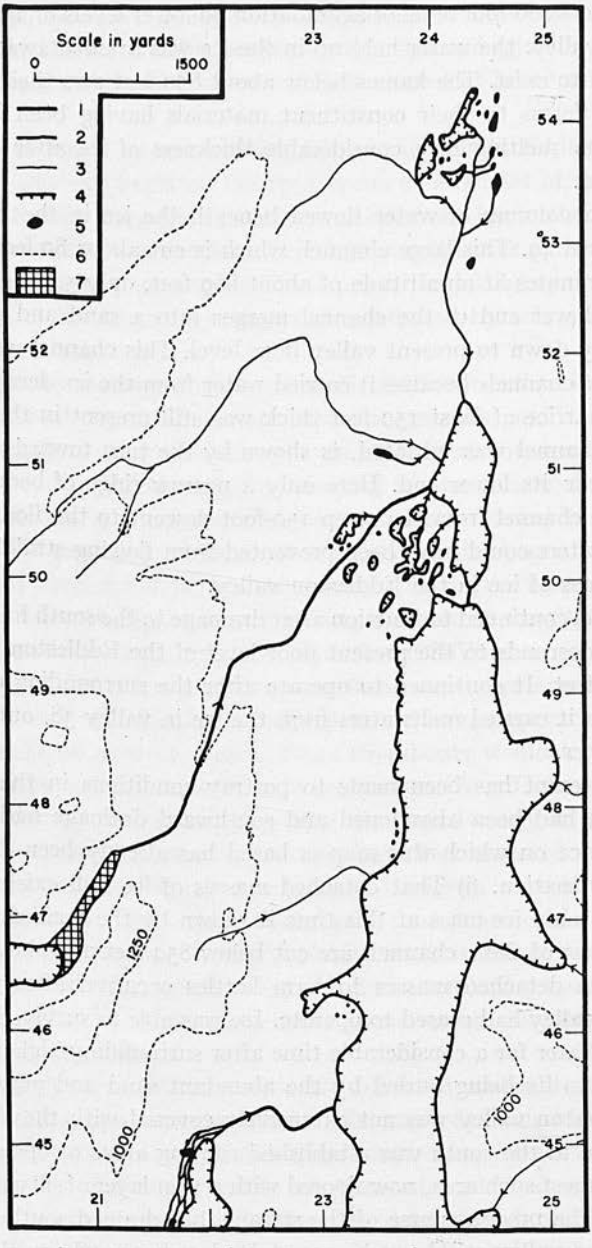


Fig. 6. Drainage and ice limits in the Eddleston valley shortly after the establishment of drainage to the south. 1, Major meltwater stream. 2, Minor meltwater stream. 3, Other stream. 4, Approximate position of ice-margin. 5, Very small detached mass of ice. 6, Contour. 7, Lake. Note: most of the ice shown on this map was probably buried beneath sand and gravel except in the extreme south and east.

a steep rock slope more than 100 feet high, concave towards the west, suggests undercutting by much larger volumes of water than are ever carried by the present stream.

Conclusion. The final deglaciation of the Eddleston valley has now been traced in some detail and it will have been apparent that in order to explain both the erosional and depositional features it is necessary to postulate extensive drainage of meltwaters through and beneath the ice. This interpretation is consistent with the evidence from present-day temperate glaciers, where, as numerous descriptions have made clear, subglacial flow of meltwaters in particular is widespread. Such drainage has been described in the Alps, Norway, Iceland, parts of Greenland, Alaska, the Himalayas and the Caucasus (e.g. Rabot, 1905, with numerous references; Blackwelder, 1907; Tarr, 1908—09, 1909; Tarr and Martin, 1914; Wentworth and Ray, 1936; Thorarinsson, 1939; Liestøl, 1955; Arnborg, 1955; Charlesworth, 1957). As Flint (1957) has pointed out, however, the retreat of modern glaciers is often not comparable in certain respects with that of many former glaciers. This generalization applies to the glaciers that occupied the Eddleston valley. Almost all retreating glaciers of the present day are withdrawing towards higher ground, often along valleys that have considerable gradients, so that the glaciers often remain active. In the Eddleston valley, however, the ice had come from a part of the Southern Uplands lying to the south-west and it retreated down the valley. Consequently, as thinning of the ice progressed during the deglaciation, the ice in the valley became increasingly cut off from its source of supply and probably became more or less inactive at an early stage. It was certainly completely stagnant in the final stages as it wasted away in the inner valley of the Eddleston Water. These conditions favoured the extensive development of subglacial and englacial drainage as described above. The deglaciation of the Eddleston valley area was thus closely comparable with that of parts of New England and northern Sweden described respectively by Flint (1929, 1930) and Mannerfelt (1945).

Supposed ice-dammed lakes in Britain

It was suggested in the introduction to this paper that the view that ice-dammed lakes were frequently formed in upland Britain as glacier ice withdrew from higher to lower ground may require considerable modification. In the preceding pages an attempt has been made to show that the ice-dammed lake hypothesis, despite favourable conditions, is inapplicable to a selected area, and the author has at the same time attempted to demonstrate the importance in this area of subglacial and englacial flow of meltwaters, ice thinning and stagnation. It is now proposed to indicate why there is good reason to suspect that such conditions were not peculiar to the Eddleston valley (or to other areas in Britain that the author has studied in detail).

Ice-dammed lakes formed as ice retreated from higher to lower ground have been described in many parts of upland Britain, for example around the Welsh uplands (Charlesworth, 1929; Jones and Pugh, 1935; Embleton, 1956), the Pennines (Dwerryhouse, 1902; Carter, 1905; Jowett, 1914; Jowett and Charlesworth, 1929; Raistrick, 1931, 1933; Anderson, 1940; Wray, 1948), the Lake District hills (Smith, 1912, 1932; Smith et al., 1931; Dixon

et al., 1926; Eastwood, 1930), the Lincolnshire Wolds (Twidale, 1956b), the Yorkshire Wolds (de Boer, 1944, 1958), and the Cheviot Hills (Kendall and Muff, 1901, 1903) and in the Midland Valley of Scotland (Kendall and Bailey, 1907—08). The evidence from which these former lakes have been deduced is best considered under the four headings used by Kendall in 1902 in his paper on the Cleveland Hills (on which the studies referred to above, as well as many others, were modelled), namely beaches, deltas, floor deposits and "overflow" channels.

(i) *Beaches.* Almost all the supposed ice-dammed lake sites in upland Britain lack definite beaches. Some papers on ice-dammed lakes contain no reference to shore features and in others the author states that only doubtful features exist. For example, Kendall, in his study of the Cleveland Hills stated, regarding beaches (p. 474): "I am unable to cite a single well-marked example in the Cleveland area . . .". Common (1954, p. 133), discussing the eastern part of the Cheviot Hills, where former ice-dammed lakes were described by Kendall and Muff (1901, 1903), stated: "Evidence of pro-glacial lakes is variable and generally poor". Twidale (1956b, p. 48) postulated ice-dammed lakes in North Lincolnshire but said: "Few good lacustrine phenomena have been found".

The absence of such evidence is normally imputed to one or more of several causes, which are conveniently summarized by Flint in his recent text-book as follows (1957, p. 145): "Around the shores of many former small or narrow glacial lakes shore features are absent as a result of: (1) feeble waves and currents, (2) short life of lake, (3) fluctuation of water level by changes in the outlet or by crustal warping, (4) susceptibility of faint shore features to destruction by mass-wasting, or (5) presence of firm bedrock or glacier ice along parts of shores." It is possible, however, that many of these supposed former lake sites lack shore features because lakes never existed there.

In this context the famous shore-lines of Glen Roy and adjacent glens (Jamieson, 1863, 1892) are significant. They are unique in Britain as regards their clarity and extent: Carruthers (1939, p. 323) referred to them as "the brilliant exception". Yet most or all of the adverse factors listed by Flint are present: in particular the valleys are narrow and deep so that waves and currents were probably feeble, and the shore-lines occur on steep valley sides in an area of very high precipitation, thus favouring rapid mass-wasting. Although these shore-lines were produced late in the glacial history of Britain, their formation and preservation despite adverse factors are nevertheless remarkable when one considers the complete absence of definite shore-line features around the margins of most supposed ice-dammed lake sites elsewhere in Britain, often in locations where conditions both for the formation and preservation of such features would have been much more favourable than in the Glen Roy area. It is not unreasonable to suggest that the explanation of this anomaly may be that most of the ice-dammed lakes supposed to have existed elsewhere in upland Britain did not in fact exist.

(ii) *Deltas.* Many of the meltwater channels in Britain that have been interpreted as overflows from lakes are of considerable dimensions and were clearly cut by large volumes of water flowing with considerable velocity. The sudden checking of these waters where

they entered an ice-dammed lake should have resulted in the deposition of large amounts of coarse material (unless, as rarely, the lithology was such that little or no coarse material was produced). Yet many of the channels interpreted as having been cut by meltwaters flowing into ice-dammed lakes do not have deltas, or features that have been interpreted as deltas, at their lower ends. For example, Kendall stated (1902, p. 475): "In the Cleveland area such deltas are rarely seen, but some examples exist, especially in the valley of the Esk." Where deposits do occur at the lower end of a channel their volume is frequently surprisingly small in relation to the size of the channel, even when due allowance is made for non-deposition of fine material.

A further aspect of the features that have been interpreted as deltas formed in ice-dammed lakes is that they frequently do not possess the gently-inclined top surface and steep frontal slope normally associated with deltas. For example, Smith (1932, p. 67), describing a deposit in Cumberland, said that it was "strictly deltaic, but does not exhibit typical deltaic forms". Many of the features interpreted as deltas are undulating areas or ill-defined spreads of sand and gravel. For example, Kendall (1902, p. 475), discussing the Cleveland area, said they are usually "simply patches of current-bedded sand and gravel". It seems probable that many of these features would not have been interpreted as deltas had not deltas been expected to occur on the ice-dammed lake hypothesis. A further anomaly was recorded by Smith (1932, p. 67), for he observed that gravels had been deposited below the level of his Lake Eskdale in the south-western part of the Lake-District. Embleton (1956) encountered the same difficulty in north-east Wales, for he said (p. 403): "the outer edges of these deltas usually extended as much as twenty or thirty feet below lake-level as determined from the overflow channels". He suggested "suspension currents" as the cause of the anomaly but admitted that "this remains an hypothesis at present".

(iii) *Floor deposits.* The deposits of supposed ice-dammed lakes, in particular the laminated clays generally interpreted as floor deposits, have been discussed in detail by Carruthers (1939) in his treatment of the glacial drifts of northern England. He stated (p. 326): "The laminated clays are not in any sense water-laid. They are the 'banded dirts' of the englacial detritus, released by the rising bottom melt . . .". Whether or not one accepts Carruthers's interpretations—and they do not appear to have received the consideration they deserve—one must agree that the detailed evidence he has presented renders highly suspect many former interpretations of supposed lake-floor deposits.

(iv) *"Overflow" channels.* The principal evidence on which the great majority of the numerous ice-dammed lakes postulated in upland Britain are based is the "overflow" channels. Almost always the evidence of supposed beaches, deltas and floor deposits, where present, is used only as supporting evidence. This was made clear by Kendall (1902, p. 480) in discussing channels, for he stated: "I have placed my chief reliance upon them in the identification of the series of lakes which form the subject of this paper." As another example, one may cite Jowett and Charlesworth (1929, pp. 319—320): "Large extra-glacial lakes were held up by the ice-barrier in the valleys and against the rising

ground of the South-Western Pennines, as is demonstrated by the frequent trenches along the hill-flanks, parallel with the contours, the numerous gashes across the hill-spurs, and the much rarer fossil deltas and lake-floor deposits."

In discussing channels in the introduction to his study of the Cleveland Hills Kendall stated (1902, p. 480): "When an ice-sheet or glacier obstructs the drainage of a country, the water is impounded so as to form a lake, which will find an escape through, or over, the ice; or, by overflowing, by some col or spur in the surrounding watershed." He thus recognised the theoretical possibility of flow through or over the ice, but he did not refer to a single example in his hundred-page paper on the Cleveland Hills. He made no reference whatever to subglacial flow of meltwaters. The absence of such references is entirely consistent with his hypothesis that the ice formed a "dam" or "barrier" by which the lakes were "impounded", terms frequently used by later authors. Yet a number of authors who have postulated ice-dammed lakes have at the same time found the orthodox ice-dammed lake hypothesis inadequate to explain certain channels and have suggested supraglacial, englacial or subglacial flow. For example, Dwerryhouse as early as 1902 (p. 585) suggested that in one instance "water probably flowed on to the ice, or may have found its way under or through the glacier, as do some of the streams on the Malaspina Glacier in Alaska". Similar brief references are to be found in studies by Smith (1912, pp. 425, 443 and 447), Eastwood (1930, pp. 109 and 111), Hollingworth (1931, pp. 309, 315 and 319), Edwards (1938) and de Boer (1958, p. 193). Considerable significance was attached to subglacial drainage by Dixon (Dixon et al., 1926, pp. 58 and 60) in his discussion of the deglaciation of the Carlisle area. Carruthers, discussing the Cheviot Hills, went a stage further and abandoned the ice-dammed lake hypothesis, stating that "remnants of terracing at the mouths of the cuts are so rare as to suggest drainage from a shallow 'bergschrund', or even directly across the ice rather than from impounded lakelets" (Carruthers, 1932, p. 134). Common (1957), in his study of some of the Cheviot channels, agreed with Carruthers and, in addition, discussed examples of subglacially-formed channels. With the probable exception of the last three authors cited, however, drainage beneath, through or over the ice was apparently suggested only as a last resort when other explanations had failed. The present author considers that englacial and, in particular, subglacial flow of meltwaters was far more important during the deglaciations of Britain than has yet been recognised. In the preceding account of the deglaciation of the Eddleston valley an attempt has been made to demonstrate the importance of such drainage in a specific area. The author has also examined channel systems in many parts of Britain where ice-dammed lakes have been postulated and has found little or no evidence of lakes of the dimensions described but has found abundant evidence of subglacial drainage. He has attempted to demonstrate the importance of such drainage in studies of East Lothian (1958a) and southern Northumberland (1958b). The former area was studied by Kendall and Bailey (1907—08) who claimed that an ice-dammed lake 14 miles long had existed. The author found no evidence of such an extensive lake but, on the other hand, found numerous subglacially-formed channels. In the latter area Dwerryhouse (1902) described series of ice-dammed lakes, yet the present author found instead

abundant evidence of an extensive subglacial drainage system which includes the deep rock gorges occupied by some of the present streams.

The inadequacy of the ice-dammed lake hypothesis as put forward by Dwerryhouse for southern Northumberland was pointed out by Trotter (1929, p. 583) and again by Peel (1951), the latter drawing attention to the remarkable up-and-down long profiles of two large channels that had been interpreted by Dwerryhouse as lake overflows. Peel considers that "quite a number of channels, in northern England at least" have up-and-down profiles (1956, p. 483), a generalization that is probably true of a much larger part of glaciated Britain. Channels of this type have been recorded as occurring in the North Yorkshire Moors (Kendall, 1902, pp. 525 and 555; Peel, 1956, p. 483), Lincolnshire (Twidale, 1956a), Weardale (Peel, 1956, p. 483), the Cheviots (Common, 1957, pp. 96—97) and East Lothian (Kendall and Bailey, 1907—08), and around the margins of the Lake District (Dixon, 1922, pp. 125—6; Dixon et al., 1926, p. 57; Smith, 1932, p. 68). Several explanations of the origin of these channels have been suggested by these authors, the most popular one being reversal of drainage between ice-dammed lakes, favoured by Kendall, Smith, Peel and Twidale. Peel, however, has subsequently informed the author that he no longer favours this explanation for the two channels in southern Northumberland. The author has suggested elsewhere (1958b) that the two Northumberland channels were formed by subglacial streams flowing under considerable hydrostatic pressure, an explanation that, unknown to the author, Mannerfelt had earlier suggested to Peel. This explanation was adopted by Dixon as long ago as 1922 (p. 126) to explain two up-and-down channels in Ennerdale, Cumberland, following de Geer's suggestion that a Swedish channel of this type was formed in this way (de Geer, 1909, p. 534). The author suspects that many of the British up-and-down channels in bedrock were produced in this manner. Whether or not this suspicion proves correct, however, it will be apparent that the number of such channels already recorded, coupled with the probability that many more exist, renders the orthodox ice-dammed lake hypothesis highly suspect.

The significance of glacier thinning. Current explanations of ice retreat from high to low ground in upland Britain frequently encounter a major difficulty. A typical example is shown in Fig. 7. In order to maintain the approximately-parallel retreat of the ice-margin from position ABCDE to position FG it is necessary to assume that the thick ice in the valley melted at a far more rapid rate than the much thinner ice on the spurs on either side. This appears highly improbable, even when allowance is made for some melting of the ice in the valley by transfer of heat from lake waters (assuming that a lake existed). It is difficult to see how such retreat could have taken place even if one thinks in terms of backwasting alone. When one takes downwasting into account, however, such retreat appears impossible: in terms of downwasting the ice-margin would take on positions such as AHCKE, LCME and NO, thus becoming adapted to the relief. Such relationships have long been recognised as applying in low-lying areas of Britain where ice-contact features abound and have been illustrated in detail especially by Dixon (1922), Trotter (1929) and Hollingworth (1931).

In this context it is important to recognise that at the time Kendall put forward his

ice-dammed lake interpretation for north-eastern Yorkshire the importance of glacier thinning was not appreciated. Abundant evidence has now been accumulated, however, both on present-day retreating glaciers (Charlesworth, 1957) and in formerly-glaciated areas, showing that, except for calving glaciers, the wastage of glaciers is essentially by thinning and only to an insignificant extent by backwasting at the terminus. As Hollingworth (1931), Flint (1929, 1930), Mannerfelt (1945, 1949), Hoppe (1950), and many others have shown in formerly-glaciated areas, downwasting resulted in a very close relationship

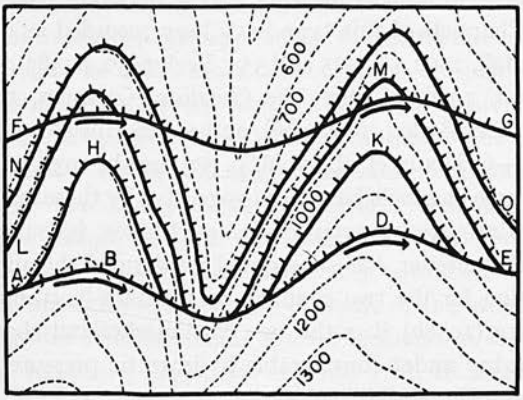


Fig. 7. Map of a typical area in upland Britain to illustrate two types of relationship between ice-margin positions and relief.

between the position of the ice-margin and the form of the ground, even small ridges and hills gradually emerging above the surface of the ice, the ice remaining longest in valleys and lowlands. The significance of these relationships in terms of the ice-dammed lake hypothesis as applied to upland Britain has not been generally appreciated, however: glacier thinning would have caused the sites of the supposed lakes to have been occupied by ice at the times the lakes should have existed. The ice in the valleys in many instances would have become increasingly cut off from its source of supply as its surface was lowered by thinning, as was recognised by Hollingworth (1931) in his study of western Edenside. Consequently the ice would have tended to become stagnant, thus facilitating the extensive development of subglacial, and englacial drainage, as discussed in detail by Mannerfelt (1945).

The ice-margin positions deduced in those parts of upland Britain where glacial drainage channels exist on spurs are normally carried across the intervening valleys either as nearly straight lines or, more often, with a slight convex curvature up-valley (as shown in Fig. 7). Such ice-margin positions have almost always been deduced *only from the channels*. Yet the channels indicate positions of the ice-margin *only on the spurs* (and here only strictly marginal channels fix the position of the ice edge accurately). As Trotter (1929, p. 580) has pointed out, "it is rarely possible to trace the position of the ice-front in open country between the spurs". There appears to be no reason why, in most cases, the ice-margin

positions should not be drawn to delimit tongues of ice extending up the valleys, the edge of the ice at any particular time corresponding closely with the contours of the valley sides. As has been shown above, simple theoretical considerations force one to conclude that this is how they should be drawn.

Re-interpretation. Once this is recognised the relations of many channels to the ice-margin are seen to be quite different from those normally assumed in Britain. Channels on spurs are no longer mostly "overflow" channels from ice-dammed lakes: they are mostly the channels cut by streams supplied mainly by water held up within or flowing through, beneath and/or over the ice that occupied the intervening valleys. Some channels were cut subaerially, some subglacially, and others were formed in both ways. Some of these relationships have been considered above in the discussion of the channels of the Eddleston valley, and others have been discussed elsewhere (Sissons, 1958a, 1958b).

Depositional features also take on a different significance. Deposits interpreted as deltas or shore-lines formed in ice-dammed lakes were laid down between the ice-margin and a hillside or deposited beneath, within or on top of the marginal zone of the ice. These deposits were laid down in very narrow marginal lakes or in stream beds. Where the deposits accumulated on top of, within or beneath the ice their form was modified by later melting of the ice, thus accounting for the frequent absence of true delta forms. Where the materials accumulated between the ice edge and a hillside a kame terrace remained after the ice had melted, if the materials were sufficiently abundant and the slope of the hillside not too steep. The terraces described by Embleton (1956) in north-east Wales are kame terraces, not shore features, an origin indicated by "the frequency of kettle, holes and other irregular depressions" in them. The horizontality of these terraces, which appears to favour their interpretation as shore features, is due to their accumulation having been controlled by the level of the water-table within the ice. In East Lothian the author has mapped an almost horizontal terrace of similar origin some three miles in length; in this case the marginal origin is proved by the relation of the terrace to marginal channels. The interpretation suggested here also allows the extension of deltas below supposed lake levels remarked on by Smith (1932, p. 67) and Embleton (1956, p. 403) to be explained as due to parts of the deltas having been deposited on top of or within the ice and subsequently lowered when the ice melted.

The frequent absence of deposits at the lower ends of meltwater channels, or the small volume of such deposits compared with the size of the channel from which they were apparently derived, is in some instances due to the stream that cut a channel having flowed into the ice with a velocity sufficient to enable it to carry away the material produced by channel cutting, as in parts of southern Northumberland. In other instances the material may have been deposited over a considerable area beneath, within or on top of the ice, in which case it may have been redistributed during the melting of the ice and possibly covered by or mixed with other englacial and supraglacial material so as to be no longer readily recognizable. The interpretation suggested here also enables the large accumulations of sand and gravel found in some localities to be explained. For example, the abundance of such materials in the middle Tyne valley remarked on by Trotter (1929)

may be due to their having been largely supplied by the extensive subglacial drainage system that developed on the hills to the south during the last deglaciation. The fact that these deposits form an extensive "ose-train" (Trotter, 1929, p. 578) is in accord with this interpretation.

Three consequences of the hypothesis outlined above may be briefly noted. (i) Postulated oscillations of an ice-margin based on the relative altitudes of channels on the same spur (for example, Kendall, 1902, pp. 509, 523, 538, 542, 543, 546, 548, 555—6; Jowett and Muff, 1904, p. 232; Kendall and Bailey, 1907—08; Dixon et al., 1926, p. 59; Stephens, 1953, pp. 134—5) are probable invalid, for channels were not necessarily formed in the order in which they now occur on a spur. (ii) The conclusions of some authors that certain meltwater channels and related deposits indicate the presence of ice from two distinct sources results from an insufficient appreciation of the significance of thinning and may well be incorrect. For example, Trotter (1929) suggested that, as the Lake District-Galloway ice retreated down East Allendale in the northern Pennines, a local glacier advanced at the same rate, maintaining contact with the former ice-mass. Trotter suggested this rather unlikely explanation to account for certain marginal channels that are simply explained in terms of thinning as having been formed solely along the margin of the Lake District-Galloway ice. It has been suggested elsewhere (Sissons, 1958a) that Kendall and Bailey's (1907—08) conclusion that local glaciers existed on the Lammermuirs is based on a misinterpretation of the meltwater channels. Smith's (1932) conclusion that the Lake District valley glaciers and the Irish Sea ice separated from each other near the lower ends of Eskdale and other valleys certainly requires re-examination. Embleton's (1956) conclusion that the Welsh ice and Irish Sea ice parted company in the area he studied in north-east Wales, retreating in opposite directions down two valleys heading in the same col, is more probably the result of thinning of a single mass of ice resulting in the emergence of the col while ice still occupied the valleys on either side. (iii) If the ice-dammed lake hypothesis is untenable in certain areas it follows that reconstructions of ice-margin positions in these areas from the altitudes of supposed lakes and "overflow" channels may be incorrect also. It would appear preferable to reconstruct ice-margin positions primarily by attempting to determine the gradient of the ice-margin from strictly-marginal features (as discussed on pp. 6—8 above), supplementing this with other evidence wherever possible, although this method might present considerable difficulties owing to inadequate evidence in the less recently glaciated parts of Britain.

Conclusion. It may appear improbable to the reader that numerous workers in Britain over the course of more than half a century should have consistently misinterpreted the evidence that has led them to postulate the former existence of ice-dammed lakes. The author would therefore strongly emphasize that the basic reason for this is that, ever since Kendall's classic paper on the Cleveland Hills was published, meltwater channels and related depositional features in upland Britain have almost always been interpreted *on the assumptions that ice-dammed lakes existed and that most of the channels cut across spurs are "overflow channels" or "spillways".* Alternative interpretations have been seriously considered only occasionally. This is illustrated by the introductory paragraphs

of numerous papers on this subject, of which a single example will suffice as an illustration. The first sentence of Smith's 1932 paper reads: "The occurrence in former times of glacier-lakes in this country, due to the impounding of drainage from a locally-glaciated mountain group by an impinging ice-sheet, is nowadays accepted as an undeniable fact." One consequence of this widespread acceptance of the ice-dammed lake hypothesis is that the abundant evidence in Britain of englacial and, in particular, subglacial flow of meltwaters has remained almost unrecognised. Once it is realised that Kendall's hypothesis is not the only hypothesis to be taken into account, a quite different interpretation of the complex patterns of meltwater flow in association with wasting ice is very frequently revealed.

It may be noted that two authors, at least, have expressed doubts as to the validity of Kendall's hypothesis. Regarding northern England Carruthers concluded that, "all things considered, it seems probable that ... the sites of these supposed late-Glacial lakes were mostly occupied by stagnant ice" (1939, p. 325). Peel (1956, p. 486), referring to subglacial drainage, stated: "This idea does not seem to have as yet received much serious consideration in Britain, where thought has been perhaps unduly influenced by Kendall's concept of freely draining lakes despite the common and disturbing lack of any independent direct evidence for such lakes." It is also relevant to note that similar suggestions have been made regarding areas outside Britain. For example, Flint (1929, 1930), in his studies of Connecticut, emphasized the importance of thinning and suggested that many supposed ice-dammed lake sites were largely occupied by ice, kame terraces being formed in narrow marginal lakes. Hoppe (1957, pp. 14—15), discussing Halden's "almost classical" map showing a large number of elongated ice-dammed lakes in the Swedish mountain chain stated: "Glacial lakes several tens of kilometres long were in fact much rarer than has been thought; one is tempted to add, 'if they ever occurred at all' ".

Finally, it must be made clear that it is not suggested that no ice-dammed lakes ever existed in upland Britain. The author has not been concerned here with areas where active valley glaciers were retreating up their valleys towards their sources in high ground. Many ice-dammed lakes may have developed in such locations, as described, for example, by Charlesworth (1956), although their drainage was almost certainly far more complex than has been generally assumed. The author has been concerned here with those considerable areas of Britain where ice from a more or less distant source was receding from high ground that it had previously partly or completely covered. Even in these areas it would be rash to suggest that no ice-dammed lakes ever existed (for example, if undisturbed lake floor deposits occur this can hardly be disputed). However, the author does suggest that in these areas there are very good reasons for suspecting that most ice-dammed lakes of the dimensions postulated by numerous authors since the beginning of this century did not exist.

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GLACIAL LAND FORMS OF THE SYRACUSE-ONEIDA AREA, NEW YORK

a, large steep-sided glacial-drainage channel; b, large glacial-drainage channel bounded by gentle to moderate slopes (broken lines indicate limits of flattish floor); c, other glacial-drainage channel; d, waterfall site; e, plunge basin; f, steep slope of side of large glacial-drainage channel, of limestone scarp, and bounding terrace; g, end moraine; h, drumlin; i, kettle; j, kame ridge; k, esker; l, kame terrace, river terrace, outwash terrace, or delta surface; m, fluvioglacial sand and gravel (symbol omitted on terrace and delta surfaces); n, Lake Iroquois cliff; o, bar of Lake Iroquois; p, cliff line above Iroquois level; q, small shore features, mostly above Iroquois level; r contour, interval 50 feet (omitted where other symbols shown). Contours from U. S. Geological Survey maps.

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SUBGLACIAL, MARGINAL, AND OTHER GLACIAL DRAINAGE IN THE SYRACUSE-ONEIDA AREA, NEW YORK

BY J. B. SISSONS

ABSTRACT

In his numerous papers on central New York, H. L. Fairchild interpreted almost all the glacial-drainage features as of marginal origin. In the Syracuse-Oneida area, however, there is morphological evidence of subglacial stream erosion by both local and Great Lakes waters, of subglacial deposition by local waters, and of englacial flow in places. Evidence also indicates that the Great Lakes waters in some places flowed in channels walled by ice. Only limited evidence of ice-dammed lakes exists in the Syracuse-Oneida area itself, except at the lowest altitudes, where numerous small shore features indicate a former local lake above the highest Iroquois shore line. There is evidence of two periods of Great Lakes drainage across the hillslopes of the Syracuse-Oneida area during the oscillatory retreat from the Wisconsin maximum. The writer suggests, however, that these two periods of Great Lakes drainage alone are insufficient to account for the vast, complex channel system.

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INTRODUCTION

"The glacier acting as a barrier to northward drainage is the fundamental fact to be apprehended by the reader. The ice sheet was a melting dam during both its advance and its retreat, and waters were flowing copiously from it, not into it. Valleys or land depressions sloping toward the ice front were by the ice barrier made into lake basins."

With these words Fairchild (1909) opened a paper on glacial drainage in central New York in which he summarized his conclusions given in many of his numerous earlier publications. The quotation shows that Fairchild, like many other writers in North America and Britain, erected his interpretation of glacial drainage on the assumption that glacier ice is an impenetrable barrier to waters outside it. This paper's main purpose is to show that during the last retreat of the ice from the Syracuse-Oneida area, meltwaters frequently flowed into, beneath, and across the ice, giving rise to a variety of erosional and depositional features. This recognition of the complexities of the glacial drainage necessitates certain modifications of Fairchild's interpretation of the glacial retreat, an interpretation that appears to have remained essentially unchallenged.

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METHODS

The geomorphic features of the Syracuse-Oneida area were initially mapped from air

photographs, and this mapping was subsequently checked, and modified where necessary, by systematic traversing on foot. The generalized results of the mapping are shown on Plate 1, numerous details and minor features having been omitted in order to produce a map of reasonable dimensions.

Major Relief

The area lies on the northern margin of the Alleghany Plateau province. In the south the plateau is 1200 to 1700 feet in altitude, but the ground descends abruptly northward to the Iroquois Lake plain at about 400 feet above sea level. The upland is broken into five major sections by the deep, steep-sided valleys of the northward-flowing Onondaga, Butternut, Limestone, Chittenango, Cowaselon, and Oneida creeks. These five sections are lettered A to E in Plate 1, and points referred to in the text are numbered, *e.g.*, A3, E5. Since constant reference to this plate is required throughout the text, specific mention of it is generally omitted.

The bedrocks of the area are mainly limestones and shales dipping gently southward and southwestward. Shales crop out on the higher parts of the spurs, and in the southwestern part of the area their outcrop is roughly co-extensive with the drumlin region. Massive limestones beneath the shales correspond with a well-marked bench that rises eastward from an altitude of about 700–800 feet near Syracuse to about 1200–1300 feet near Oneida. Northward the limestone bench in many places terminates in a very steep escarpment, or in several escarpments, overlooking the lower ground to the north developed on the underlying shales.

Previous Work

The only comprehensive treatment of the glacial drainage of the Syracuse-Oneida area (and adjacent areas) is by Fairchild, who published numerous papers during a period of 35 years on aspects of the glaciation of central New York. His conclusions regarding the glacial history changed slightly as he accumulated evidence, but his final conclusions (1909; modified in 1932a; 1932b), which Flint (1953; 1956; 1957, p. 343–348) apparently accepted, were, briefly, as follows.

(1) As the ice sheet withdrew from the northern slopes of the uplands of central New York, lakes were impounded in the valleys of the Finger Lakes region and in the valleys to the east. The earliest drainage of these lakes was southward through the cols at the valley heads.

(2) This southward drainage was gradually replaced by westward drainage into Lake Warren, the outlet of which was the Grand River valley in western Michigan. The westward drainage certainly commenced as far east as the Onondaga valley and probably as far east as the Limestone valley.

(3) The westward drainage was replaced by eastward drainage, the latter developing first in the eastern part of central New York and slowly extending farther west. This eastward drainage did not involve the Warren waters for a considerable time, however, since the ice margin in the Batavia area (100 miles west of Syracuse) remained stationary and continued to hold back the Warren waters while it was retreating farther east. Consequently during this retreat Great Lakes waters did not escape across the hillslopes of the Syracuse-Oneida area, which were affected only by local waters.

(4) Eventually the ice margin in the Batavia area began to retreat, and the Warren waters were thus able to escape eastward, a series of channels being cut in the Batavia-Rochester area from 870 feet down to about 480 feet. These waters flowed only through the low ground in the Rochester-Syracuse-Oneida-Rome district, escaping through the Mohawk Valley.

(5) After retreating an unknown distance, the ice readvanced and Lake Warren was re-formed with its outlet again through the Grand River valley. This time, however, the ice margin was highest in the Syracuse area, not the Batavia area. Consequently, during this readvance and the final retreat of the ice, huge channels were cut near Syracuse by the large volumes of water escaping eastward to the Mohawk Valley.

(6) With continued retreat of the ice, Lake Iroquois was established at various altitudes up to 460 feet at Rome and 435 feet at Rochester (this difference in altitude being partly due to isostatic recovery of the area).

Local Glacial Drainage

Except in the extreme eastern part of the area, the features produced by locally derived waters occur mainly on the spurs and are found above a zone descending from about 900–950 feet in the west to about 650 feet in the east. They comprise numerous meltwater

channels and kames, a few small deltas, and two small esker systems.

Most of the channels are small, ranging from 2 or 3 feet deep to 40 or 50 feet deep. A few channels attain depths of 100 feet for short distances, but they are much narrower than the channels formed by Great Lakes waters. The channels were excavated by waters that ultimately found their way eastward to the Mohawk Valley, excepting a few in the extreme southwest that carried meltwaters westward. These channels (A1, A2, B1) are consistent with Fairchild's conclusion that drainage was at first westward on spurs A and B.

Although numerous marginal channels occur above the levels affected by Great Lakes waters it has not been possible to correlate them throughout the area studied and thus map successive positions of the ice margin. In fact, it is rarely possible to deduce the position of the ice margin from marginal channels for more than a mile in any one locality. The principal exception occurs on spur D and even there the ice-margin position can be traced only 3 miles. Channel D1 starts at about 1230 feet and ends in a delta whose surface lies at about 1220 feet. The shallow channel D2 begins at the edge of the delta and descends to about 1210 feet. A short distance to the east channel D3 begins at 1210 feet and ends in a delta whose surface lies between 1200 and 1170 feet. A mile farther northeast channel D7 commences at 1170 feet and terminates at about 1140 feet. This sequence thus provides an example of very small ice-dammed lakes draining from one to another by channels cut across spurs. Fairchild (1909, p. 39) described part of this sequence. Significantly, it appears to be unique above the levels of the region affected by Great Lakes waters in the area studied. According to the marginal-drainage hypothesis, one would expect innumerable examples, for at every position of the ice margin during its retreat, there should have been along or near the ice edge either meltwater erosion, producing channels, or deposition, forming kames, kame terraces, deltas, or lake-floor deposits. Fairchild recognized this, and his maps of the Syracuse-Oneida area (and of adjacent areas) show such features in abundance (Fairchild, 1909): in particular, numerous channels clearly convey the impression of meltwaters sweeping along the ice margin where it lay against the slopes of the successive spurs. Fairchild's maps are misleading in this respect, however, since many of the smaller channels marked on them do not appear to exist.

Plate 1 shows that there is no evidence of marginal drainage in numerous places where one would expect to find it on the basis of Fairchild's hypothesis. The absence of such evidence is by no means uncommon in similar localities elsewhere in North America and in Europe and has been frequently explained as due to (1) subsequent destruction of the evidence or (2) adjacent ice-dammed lakes having become confluent across intervening spurs. Regarding the first point, it is probable that a few small channels have been destroyed or obscured, but this explanation is clearly inadequate in many instances. For example, on the basis of the marginal-drainage hypothesis, one would expect series of marginal channels and related deposits to occur on spur D immediately below the level of the well-defined series D1, D2, D3, and D7 described heretofore. Yet only two shallow marginal channels and one small deltaic area occur here, the rest of this area apparently being devoid of marginal features. The second explanation might account for the absence of marginal features at a few points but can hardly be invoked repeatedly. Furthermore, it presupposes the existence of marginal lakes of considerable depth. There is very little evidence of even small marginal lakes having existed on the spurs, however, and no evidence at all of large ones: kames, deltas, and lacustrine deposits are of limited extent on the spurs.

If then, the absence of marginal features from many parts of the spurs cannot be explained in either of these ways, it would seem that one must conclude that the meltwaters frequently flowed beneath, through, and/or over the ice itself. Concerning the two last, evidence is obviously difficult to obtain. There is abundant evidence of the first, however. A considerable number of channels run along the contours of hillsides for distances ranging between a few hundred feet and about a mile before turning through a right angle, in places abruptly, to run straight down the slopes. Examples include C2, C4, D10, D11, D12, D16, and several channels in area 3 on spur E. The parts of such channels running along the slopes were cut by meltwaters flowing along the ice margin, and the down-slope segments were produced by these same waters flowing beneath the ice. The sharp bend in a channel of this type is the point at which the meltwaters turned to flow beneath the ice. Many such channels have been

described in northern Sweden by Mannerfelt (1945), who called them "subglacial chutes".

Some subglacial chutes carried beneath the ice the meltwaters from a number of marginal channels at different stages of the retreat. Channel system D14 is an example. The upper part of this system, aligned approximately west-east and cut about 10–20 feet into limestone, was formed along the ice margin. Traced eastward the main channel is found to increase abruptly in depth to about 80 feet as it swings around through 90° to run steeply down-slope. This down-slope segment, in part of subglacial origin, runs at right angles to several marginal channels, some of which are clearly developed on its western side but do not exist on its eastern side. Thus the down-slope channel must have carried the waters of these marginal channels beneath the ice. Other examples of channel systems of this type are shown in plate 1, especially on spur D.

This interpretation of certain valleys running down hillsides, including some valleys containing streams, as wholly or partly of subglacial origin is supported by other evidence. In the case of marginal channels that turn through a right angle to run down slopes, any other origin than a subglacial one for the down-slope section is difficult to visualize, for both parts of such channels are continuous with each other and form a single feature that can be explained only in terms of erosion by waters flowing from one end to the other (*i.e.*, marginally to the ice, then subglacially). Furthermore, the down-slope section of some of these channels ends abruptly part way down a steep slope (*e.g.*, around E3). It is difficult to see how a normal stream could suddenly cease to erode part way down a steep slope. On the other hand, the sudden cessation of bedrock erosion by a subglacial stream could be due either to its having become englacial or to its having descended to the level of the water table in the ice.

The presence of streams in some subglacial channels is to be expected, for such channels, after their abandonment by glacial waters, provided ready-made routes for later drainage. In many instances the present streams are very small in relation to the size of the valleys they occupy. The existence of misfit streams in many marginal channels has long been recognized and is to be expected in many subglacial ones also.

Some subglacial channels, however, are occupied by streams of considerable size as, for example, the valley of Limestone Creek between D11 and D26 (to be distinguished from the much larger Limestone Creek farther west). The creek rises at D6 and enters the channel at D11. A tributary rises at D8 and enters the channel at the head of the latter. The west-east part of the channel, running along the hillslope, was formed along the ice margin. Where the feature swings around to a northward direction, a steep bluff up to 20 feet high extends for a quarter of a mile along its eastern side and continues on the southern side of the channel. Presence of this bluff excludes the possibility that the eastward-flowing meltwaters that cut the marginal west-east part of the channel continued eastward beyond it and shows that the meltwaters turned to flow beneath the ice. A mile to the north two more marginal channels (D23) join the Limestone valley, indicating that their waters also followed the valley beneath the ice. A short distance farther north another marginal channel (D24) functioned in the same manner. Finally, the Limestone valley is joined by channel D25, which is an example of a marginal channel that turns abruptly through a right angle to become a subglacial chute. It thus appears that Limestone Creek between points D11 and D26 utilizes routes followed by glacial waters, and its course is therefore marked as a channel in Plate 1.

In view of the evidence of meltwater flow into the ice one must inquire at this point as to its subsequent history. In some instances meltwaters appear to have emerged from the ice at a point east of their point of entry. This is suggested by the fact that many marginal channels are of considerable size at their heads (*i.e.*, at their western ends). The ice-dammed lake hypothesis would explain this as a result of these channels having been lake overflows; this explanation is unsatisfactory since lacustrine deposits are generally absent. More probably these marginal channels have been formed by streams that flowed partly along the ice margin and partly within and beneath (and perhaps on) the ice.

In other instances subglacial chutes show that meltwaters plunged down the slopes of the land surface buried beneath the ice, in places descending as much as 200 feet. Deposits are absent at the lower ends of most of the chutes. This suggests that the streams that cut them continued to flow with undiminished velocity after becoming englacial. On the other hand, in a few areas, particularly along the middle part of spur D below the main limestone scarp, areas of sand and gravel, the latter predominating, appear to have been deposited by streams flowing beneath and into the ice. The ice-dammed lake hypothesis would explain these

deposits as having been laid down by streams flowing along the ice margin. The following evidence is against this view and favors the interpretation put forward here.

(1) The distribution of the deposits appears to be more closely related to the subglacial than to the marginal channels. For example, a small patch of sand and gravel occurs near the northern end of channel system D14, and a larger patch (D17) suggests deposition by the meltwaters that plunged down the limestone scarp beneath the ice margin in excavating channels D15 and D16.

(2) Where not subsequently dissected, the deposits commonly form low, irregular mounds that may be termed kames. These kames, however, do not have the steep ice-contact slopes and flattish surfaces that one might expect to find if they had been formed between the ice margin and the hillslopes.

(3) All the deposits are in places cut into by both marginal and subglacial channels. They must have been laid down therefore before either type of channel was cut at the points where dissection occurs. The deposits could not have been laid down between the ice edge and the hillslopes since subglacial dissection of them could not then have taken place. This suggests that deposition occurred, in part at least, at a time when the ice edge lay against or a little south of the steep, north-facing limestone scarp. Meltwaters flowed for short distances along the ice margin but then plunged down the scarp to flow beneath the ice, cutting some of the numerous gashes that now occur in the scarp. The materials produced by this erosion were partly redeposited farther down the slope beneath, and probably also partly within, the ice. As the ice continued to thin, and its edge continued to retreat down the slope, the waters flowing down the slope beneath the ice eroded to lower and lower levels, so that the deposits previously laid down by subglacial and englacial streams were partly dissected by the former. As the ice margin itself retreated across the deposits, further dissection occurred where waters flowed along it before themselves plunging beneath the ice.

Additional evidence of drainage beneath and through the ice is provided by the southern part of the valley of Cowaselon Creek. The valley is here about a mile wide and 400 feet deep. Its upper edge is generally a sharp convex break of slope, below which its sides descend steeply for some 200 to 250 feet. The central part of the valley, except for a strip along the course of the creek, is filled with deposits of sand and gravel, which in places extend up the steep valley sides. These deposits form numerous kames, ranging in height from a few feet to about 50 feet. The great majority of the kames are long and narrow and are aligned either straight up and down the valley sides or at a small angle with the line of steepest slope pointing slightly down-valley. Such kames extend to the head of the Cowaselon valley, about 2½ miles south of the southern limit of the area mapped in Plate 1.

Since the kames lie perpendicular to the direction of ice movement, they show that at the time they were formed a mass of dead ice occupied the Cowaselon valley. The alignment of the kames also indicates that they were deposited by supraglacial, englacial, and/or subglacial streams flowing toward the axis of the valley. The steep sides of many of the kames indicate that many of these streams were, in fact, subglacial. The abundance of kames on the western side of the valley north of D9 and their almost complete absence from the corresponding portion of the eastern side of the valley is largely the result of meltwaters from the west flowing into the valley.

At point D19 a deep subglacial chute, which continues a marginal channel of similar dimensions, runs straight down the valley side through a vertical distance of 300 feet. Along one side of the chute is a straight, steep-sided kame. This indicates that the waters flowing down the steep western side of the Cowaselon valley were at first engaged in deposition and built the kame in a steeply inclined subglacial tunnel. The meltwaters flowing through the tunnel later became erosive, probably owing to lowering of the englacial water table, and cut the subglacial chute, destroying part of the kame in the process.

A short distance to the north another subglacial chute (D20), which continues a short but well-defined marginal channel, descends part way down the valley side and is then continued by a kame. It appears that the subglacial waters changed from an agent of erosion to an agent of deposition at this point, probably because this was the level of the englacial water table at the time these features were formed.

Immediately to the north another mass of sand and gravel (D21) extends down the valley side, tapering from about 1000 feet wide at the top to about half this at the bottom. It was deposited by the waters that cut channel system D22. The main channel of this system runs straight down the side of the Cowaselon valley along the middle of the deposit, dividing it into

two steep-sided ridges. This indicates that, as in the case of the kame and channel at D19 described heretofore, deposition occurred in a steeply inclined subglacial tunnel, and this was followed by erosion by a subglacial stream flowing through the tunnel. The valley down the center of the depositional area (here interpreted as a subglacial chute) could hardly be of postglacial origin, since this would require a postglacial stream to have developed along the crest of a sand and gravel ridge running down the valley side.

Similar, but smaller, features to those at D21 occur at D18. Farther south, at D13, a kame runs straight down the valley side through a vertical distance of about 300 feet, and a subglacial chute follows one side of it. Still farther south, at D4, another kame extends through a similar vertical distance. This last kame does not appear to be related to a meltwater channel. Instead, it represents the deposits of a land stream that flowed into the ice occupying the Cowaselon valley, for it continues exactly the direction of the upper course of a small stream that enters the valley from the southwest, which stream changes direction at the beginning of the kame to wind its way down the valley side.

The kames of the Cowaselon valley are composed for the most part of coarse gravel with irregular thin beds of sand, except in the extreme southern part of the area studied, where the kames in the center of the valley are formed of silts and sands. The upper limit of deposition in this southern area is commonly very well defined at an altitude of approximately 900 feet. At this level most of the kames disappear, especially on the eastern side of the valley, where their crests show a marked flattening between 850 and 900 feet. Were these kames a little wider, their upper parts would form a kame terrace. It thus appears that during part of the time the dead ice in the Cowaselon valley was wasting away, a lake existed in the southern part of the valley at a level of about 900 feet. One might therefore anticipate the existence at the southern end of the valley of a channel at about 900 feet through which the lake waters escaped. A well-marked channel does exist there, but its floor level is slightly below 1100 feet. Since this channel floor is the lowest point in the drainage divide, the lake waters must have escaped northward, either along the ice margin or through the ice itself. A short narrow northward extension of the lake along the eastern margin of the ice occupying the valley is indicated by the flat-topped kames mentioned, but north of these kames for about 2 miles there is neither erosional nor depositional evidence at the appropriate altitude of marginally flowing meltwaters on the eastern side of the valley. It appears, therefore, that the meltwaters here flowed through the ice. Some of these waters then followed channels E1 and E2.

The phase of deposition up to the 900-foot level is the most clearly represented in the Cowaselon valley. Less well-defined phases occurred at slightly below 1100 feet, controlled by the channel at the southern end of the valley, and at about 750 feet, controlled by the channels at E4.

Thus, the Cowaselon valley evidence presents the following picture. As the ice margin retreated down the slopes of the spurs, a long tongue of ice remained in the valley. This tongue became thoroughly decayed, and water was able to flow freely through it. The level of the water in the ice was controlled by rock outlets, first at the southern end of the valley, and then on the eastern side of the valley toward its northern end. An englacial water table was thus developed. Where the surface of the ice lay below the level of this water table, a lake existed on top of the ice, as is demonstrated by the lacustrine deposits forming kames at the southern limit of the mapped area. The englacial water table also controlled the level of deposition within and beneath the ice: its level thus determined the upper limit of subglacial kame formation at any time and the lower limit of subglacial stream erosion. The lowering of the water table permitted the subglacial streams to dissect the long kames they had previously deposited on the western side of the valley.

It should perhaps be added that the retreat features of the Cowaselon valley are by no means uncommon; similar features have been described in Sweden (Mannerfelt, 1945) and Britain (Sissons, 1958). It may be noted also that Holmes (1952, p. 1008) concluded from studying the lithologic composition of a kame terrace in the Keeney valley, about 20 miles southwest of the Cowaselon valley, that upland drainage had carried material down against stagnant valley ice during deglaciation.

Although no definite stages in the retreat of the ice from the Syracuse-Oneida area before Great Lakes waters flowed through it can be recognized throughout the area, evidence indicates the general form of the ice margin. The evidence of the Cowaselon valley clearly shows that a narrow tongue of ice remained in that

valley after the higher ground on either side had become ice free. Kames and kame terraces in the Oneida valley show that a tongue of dead ice lay in that valley also. In the main valleys to the west such depositional evidence appears to be absent within the area studied but it is a reasonable presumption that they too were occupied by tongues of ice. This is implied by the marginal channels, which show that the ice margin closely followed the contours of the spur slopes and hence projected into the valleys. The same conclusion is indicated by the trend of the small end moraine on spur C.

This moraine appears to provide the only clear evidence that the ice was active at any time during its final retreat from the area. It consists of scattered mounds up to 30 feet high that appear to be composed mainly of till with beds of poorly stratified sand and gravel in places. In its eastern part (C2), angular limestone blocks up to 5 feet across lie on the morainic mounds or project from their grass-covered slopes. The apparent restriction of the moraine to spur C may be due to topographic factors, since it is possible that here an advance of the ice was retarded by the sharply rising ground immediately south of the moraine. Apart from this moraine the evidence suggests that the marginal zone of the ice was essentially stagnant during its final decay. Stagnation of the ice tongues in the deep valleys is expectable and in the Cowaselon and Oneida valleys is demonstrated by the ice-contact forms. Eskers near D5 and C1 indicate stagnation of the ice in the areas where they occur since they are not parallel with the direction of ice movement. In general, the preservation of the abundant evidence of subglacial drainage and the apparent absence of till moraines (apart from the exception noted) seem to suggest an essentially stagnant rather than an active ice-marginal zone.

Great Lakes Drainage

Most of the erosional and depositional features near Syracuse have been described by Fairchild, so only a brief description is necessary here. The principal channels, such as A3, A5, B6, C11, and C12, are 2–3 miles long, 100–150 feet deep, commonly have flattish floors 600–1200 feet across, and have steep sides which are commonly vertical where cut in limestone. They begin well above the levels of the valley bottoms to the west of them, and for the most part their floors slope down gently eastward. Toward the eastern end of channels A3, A5, and C12, abandoned waterfall sites occur, associated with plunge basins in channel C12. Numerous other waterfall sites, some of them associated with plunge basins, occur in a belt about 10 miles long and up to 1½ miles wide in the limestone region immediately south of channels A5, B6, and C11. Some of these are quite small, but waterfall sites at A4, B4, B5, and C10 are up to 150 feet high and occur at the heads of vertical-sided flat-bottomed gorges. Deposition by Great Lakes waters took place mainly in and near the south-north valleys (Onondaga, Butternut, and Limestone). The deposits, which are mainly coarse gravel (in some places composed largely of limestone boulders 1–2 feet across, with occasional larger blocks), have been subsequently dissected by slightly later meltwater erosion and by postglacial-stream erosion to form terraces bordering the south-north valleys and the channels.

According to Fairchild these features were formed by Great Lakes waters flowing along a retreating ice margin aligned southwest-northeast. This conclusion was unavoidable in view of his marginal-drainage interpretation. In his own words (1909, p. 36),

> "The production of cataracts on the west side of each valley . . .proves that the waters stood much lower in each valley, successively, than in the next valley on the west. Such discordance of the water-level could not have existed if the ice front had extended in an east and west direction across the ridges and valleys."

This view is difficult to reconcile, however, with the evidence already discussed indicating that prior to the establishment of Great Lakes drainage the ice margin was closely related to the contours of the hill slopes, tongues of ice extending up the major valleys. Fairchild overcame this difficulty in his final papers (1932a; 1932b) by suggesting that the features in the Syracuse area were formed by Great Lakes waters during a re-advance of the ice and the subsequent retreat. He thus regarded the features on the hillslopes south of channel C11 and east of the Limestone valley to the eastern limit of the area considered here as having been formed entirely by local waters during the penultimate retreat of the ice.

The problem of the orientation of the ice margin during the Great Lakes drainage through the Syracuse-Oneida area appears to be best approached by first ascertaining the upper limit of the region affected by this drainage. On spur A this limit is fixed by the edges of

channel A3 and by the presence of uneroded and partly eroded drumlins as lying between altitudes of 920 and 940 feet. On spur B channel B3 indicates that Great Lakes waters affected this area to just above 900 feet. On spur C the lip of the former waterfall at C9 is at 920 feet. Farther east on this spur large channels which are cut into the limestone scarp and the slope beneath it extend up to altitudes of between 900 and 950 feet. Outside the area considered here, on the spur to the west of the Onondaga valley, Fairchild (1909, p. 24) reported that the highest features cut by Great Lakes waters appeared to be between 920 and 940 feet. Thus over a west-east distance of about 18 miles there is evidence that the maximum altitude reached by Great Lakes waters when flowing through the area was between 900 and 950 feet. If, as seems probable, this evidence indicates the approximate position of the ice margin at the time Great Lakes waters last began to drain eastward, the eastward slope of the ice margin must have been very gentle (although not quite as gentle as these figures suggest, since allowance must be made for subsequent isostatic recovery). Such a gentle gradient seems more in accord with that to be expected along the margin of a large ice sheet than does the steep gradient required by Fairchild's hypothesis. It is also expectable because immediately before Great Lakes waters began to flow through the Syracuse-Oneida area the meltwater drainage was westward in the extreme west (A1 and A2), implying a westward slope of the ice margin there at that time. As the ice margin retreated prior to the Great Lakes drainage phase, the location of its highest point apparently shifted westward. When channels B1 and C3 were cut, the highest point of the ice margin lay between them, for these channels start at approximately the same altitude (1100 feet) but descend in opposite directions. By the time channels A2 and B2 were cut by waters flowing in opposite directions, the highest point of the ice margin lay farther west (between spurs A and B) and was about 100 feet lower, for these channels start at around 1000 feet. Further retreat, possibly accompanied by a further westward shifting of the highest point of the ice margin, allowed eastward drainage of Great Lakes waters to be established.

The highest feature produced by Great Lakes drainage on spur D is channel D27. The floor of this channel at its head is at an altitude of about 810 feet, and its downcutting probably commenced between 880 and 900 feet. This channel is far larger than any of the channels on the slopes above it and cannot be ascribed to local meltwater drainage. It also differs from these smaller channels in its length, for it extends for 6 miles to the Cowaselon valley. For the last mile and a half of its course it is a steep-sided trench about 100 feet deep, with a flat floor 300 to 500 feet across cut in shale.

On spur E the highest feature related to Great Lakes waters is channel E5. This channel is much larger than the channels at higher levels cut by local waters, being 100 or more feet deep for most of its course, and it also differs from them in that it forms a continuous feature from one side of the spur to the other. It is continued across the Oneida valley by an almost flat-topped mass of gravel (E6) whose surface lies between 630 and 650 feet. Thus it appears that at the time Great Lakes waters began to escape through the Syracuse-Oneida area the altitude of the ice margin declined gently eastward in the western half of the area, lying, in terms of present altitudes, between 950 and 900 feet over a distance of 18 miles, and declined somewhat more steeply in the eastern half of the area from around 950–900 feet to about 650 feet in a distance of 12 miles.

The failure of Fairchild and others to recognize that Great Lakes waters flowed through channels D27 and E5 appears to be due to two related groups of facts. First, these two channels, although far larger than the channels at higher levels on the same spurs, are, nevertheless, smaller than the vast channels cut by Great Lakes waters near Syracuse. This is obviously difficult to explain by the marginal-drainage hypothesis. Secondly, successive sequences of large channels and related features do not occur on spurs D and E below the level of channels D27 and E5. Both these groups of facts are readily explained on the interpretation put forward here, for they resulted from increasing quantities of Great Lakes waters having found their way through and beneath the ice.

Clear evidence of this is provided by a group of channels on spur C. Channel C7 is a great trench up to 150 feet deep and 700 feet across running straight down the steep western side of the Chittenango valley through a vertical distance of about 400 feet. The present tiny stream could not have formed this feature. It is a great subglacial chute cut by some of the Great Lakes waters plunging beneath the ice. The various routes followed by these waters along the ice margin before they turned to flow

beneath the ice are indicated by the channels that join channel C7 from the northwest. A short distance to the southeast are two smaller subglacial chutes (C5 and C6) whose origin is shown by their relationship to marginal channels. A short distance northwest of channel C7 a large marginal channel (C8) 600 feet broad and 50 to 100 feet deep follows the scarp face for a mile. The channel then suddenly turns through a right angle and runs straight down the hillside as a subglacial chute, descending 200 feet in three-quarters of a mile. Other large channels that carried Great Lakes waters first along the ice margin and then beneath the ice occur to the west-northwest. The westernmost feature of this group is channel C10, which, for the first half mile of its course, is 100 feet deep and 600 feet broad with vertical sides. The orientation of this part of the channel suggests it was cut along the ice margin, but it appears that the waters then flowed beneath the ice, for the channel changes direction and descends abruptly toward the floor of channel C11. Evidence of flow beneath the ice is also provided by waterfall site C9. This waterfall site is a vertical-walled semicircular recess about 600 feet across and 130 feet high cut into the limestone scarp, at the top of which is a water-worn limestone platform. For these features to have been formed, the ice must have lain against the limestone scarp, and its edge must have been at least level with the top of the scarp. Since marginal channels on spur C consistently indicate that the ice margin followed the scarp edge, the waters that cut the recess must have done so as they plunged down between the scarp face and a wall of ice, subsequently flowing beneath the ice.

The features on spur C described in the preceding paragraph resemble those produced by local waters on spur D except that they are on a much larger scale. They are repeated on an even larger scale on spurs A and B, south and southeast of Syracuse. The channels here fall into two distinct groups. One group, represented by channels A3, A4, and A5, B4, and B6, as well as by a large number of smaller channels, trends approximately west-east. The other group, represented by channel B5, the Butternut valley north of its junction with channel A3, the Onondaga valley, part of the Limestone valley, and a number of smaller features, trends more nearly south-north. Fairchild's southwest-northeast ice margins appear to represent a compromise between the directions of these two groups of features and do not permit either group to be explained satisfactorily.

Among the difficulties encountered by Fairchild's interpretation are the following.

(1) He concluded that channels A5, B6, and C12 formed a continuous route for the glacial waters at one stage in the retreat, and that at a later stage these waters followed channel A5 and the Butternut valley. That the glacial waters followed these routes is demonstrated by both channels and terraces. But that they did so along the ice margin in both cases is difficult to believe. According to the marginal-drainage view, the edge of the ice could not have been more than a mile north of channels A5, B6, and C12 at the time they functioned together, since otherwise the waters would have followed a route farther north. Therefore the ice edge must have trended west-east across spurs A and B. But by the time channel A5 and the Butternut valley carried the glacial waters, the ice margin must have changed direction to trend southwest-northeast. The ice margin must have therefore remained more or less constant in position at the head of channel A5 to allow that channel to continue functioning while only a short distance away it retreated 4 or 5 miles from near the lower end of channel C12 to some point west of the Butternut valley. Such an occurrence seems highly improbable.

(2) As already noted, the highest altitude affected by Great Lakes waters on spur A is between 920 and 940 feet, this being the maximum altitude of the edges of channel A3. The floor of this channel at its head is just below 800 feet. The maximum altitude of the water-swept limestone region around channel A4 is also just below 800 feet. Abundant evidence of meltwater erosion occurs below this altitude down to 550 feet, which is the altitude of the floors of channels A5 and A10 at their heads. This evidence is consistent with these features having been formed along an ice margin as Fairchild maintained. But it indicates that the ice margin trended, not southwest-northeast, as he concluded, but approximately west-east.

(3) On spur B approximately west–east-trending ice-margin positions are demonstrated by channels B4 and B6 and by the numerous smaller features between them. But if the ice margin trended west and east across spurs A and B, channel B5 and the part of the Butternut valley between the western ends of channels A3 and A5 could not also have carried waters along the ice margin, for they lie almost per-

pendicular to and cross the marginal features on spurs A and B.

(4) If, as Fairchild maintained, the glacial waters flowed only along the ice margin there should be continuous sequences of erosional and/or depositional features down the slopes of the spurs crossed by Great Lakes waters. On spurs B and C, however, gaps occur in the sequences. The floor of channel B6 at its head is at 530 feet. A mile to the north is a col (B7) at 510 feet. According to Fairchild's hypothesis, this col should have carried Great Lakes waters after they ceased to flow through channel B6. That it did not do so, however, is indicated by the uneroded drumlins on its southern slope, by the absence of sand and gravel deposits from its immediate vicinity, and by its somewhat irregular floor and rather subdued character compared with other channels on the same spur. It seems probable that the col is a meltwater channel, but that it was formed prior to the last retreat of the ice from the area. On spur C there is a gap in the sequence between channel C10, commencing at 840 feet, and channels C11 and C12, the upper limits of whose sides are below 700 feet.

There is thus abundant evidence of marginal drainage on parts of spurs A and B, as recognized by Fairchild, although the ice margin trended approximately west-east, not southwest-northeast as he concluded. On other parts of these spurs the marginal hypothesis is inadequate, however. Here the approximately south-north drainage evidenced by certain channels and deposits points to drainage beneath and through the ice. It is only by postulating such drainage that the apparently contradictory evidence of gentle ice-margin gradients and great waterfall sites in the same area can be reconciled. The great embayment about half way along the south side of channel A5 is walled by a cliff 150 feet high at the top of which is an area of water-worn limestone, with deep gullies and underground passages along joint and bedding planes that carried waters to the cliff face where they plunged down in great falls. These falls could have existed only if the ice margin was situated immediately to the north, and they must have developed between the limestone cliff and a wall of ice. Similar evidence points to a similar origin for channel B5, and smaller waterfall sites, some of them with plunge basins at their bases, occur along the scarp face forming the southern wall of channel B6 between channel B5 and the Butternut valley.

The waters that followed these and other routes probably flowed beneath the ice initially. In view of the great volumes of water involved and the probably limited thickness of the ice, however, the subglacial interpretation appears unsatisfactory as a complete explanation. Furthermore, it is clearly invalid in the case of the Butternut valley north of its junction with channel A5, for here extensive outwash terraces occur. Since, as pointed out, it is highly improbable that this valley carried meltwaters along the ice margin, it appears that the only satisfactory alternative is that they followed this valley in a channel walled by ice. It seems likely that flow along the valley was initiated subglacially, the roof of the tunnel subsequently collapsing to produce an open trench through the ice. The abandonment of channel A5 for the route northward down the Onondaga valley to Syracuse probably took place in a similar manner. On a smaller scale, the terraces in the Butternut valley near the eastern end of channel A3 and in the Cowaselon valley near the eastern end of the long channel D27 indicate that the flow of Great Lakes waters into these valleys caused the tongues of dead ice extending up them to be severed. The best example occurs in the extreme eastern part of the area, in the Oneida valley. Here a large mass of sand and gravel (E6) forms a feature that resembles a great dam built part way across the valley. The mass extends west-east for a mile, has an almost flat top a quarter of a mile broad that inclines very gently eastward, and is bounded by steep slopes up to 100 feet high. The feature continues exactly along the direction of channel E5, and it clearly represents the infilling of a deep trench through the ice, its steep slopes to north and south being ice-contact slopes: thus Great Lakes waters cut straight across the dead ice tongue occupying the Oneida valley.

In summary, it appears that during the final retreat of the ice from the Syracuse-Oneida area, eastward drainage from the Great Lakes was initiated at a time when the ice edge declined in altitude from between 950 and 900 feet in the western half of the area to about 650 feet in the extreme east (in terms of present altitudes). This ice mass was already much decayed in its marginal zone, as indicated by the abundant evidence of subglacial and englacial drainage developed before the establishment of Great Lakes drainage. The earliest Great Lakes flow, however, is represented by long marginal channels, such as D27 and E5, the lengths of which contrast with the shortness of adjacent,

slightly higher channels, produced by locally derived waters. The probable explanation of this is that the openings in the ice developed by local waters prior to the Great Lakes flow were incapable of absorbing this flow immediately, so that an important marginal drainage was established. The subsequent history of the Great Lakes drainage is largely one of the successive abandonment of marginal routes for others leading more directly to the low ground in the north. A short cut across the ice tongue in the Oneida valley was rapidly established by the waters that had passed through channel E5. Subsequently the waters followed a route slightly farther north, laying down the kame terraces and associated deposits around E8. The waters then abandoned spur E and flowed from channel D27 northward along the northern part of the Cowaselon valley, cutting off the remnant of the dead ice tongue in that valley (as indicated by flat-topped kames according in level with the terrace around the valley mouth) and also separating from the main ice mass an area of dead ice along the northern foot of spur E (part of which on its subsequent decay produced the kame and kettle topography south of Oneida). The volume of water flowing through the long channel D27 decreased as the series of large subglacial channels on spur C developed until, eventually, flow across spur D ceased, the last phase of flow across this spur being recorded by the kames with steep, north-facing ice-contact slopes extending from D30 to D28 and by channel D29, whose floor is at approximately the same level as the nearly accordant summits of the kames. Similar developments took place farther west, the major routes by which Great Lakes waters took short cuts to the lower ground, detaching portions of the marginal zone from the main ice mass as they did so, being represented by channel C12 and the northern sections of the Butternut and Onondaga valleys.

Former Lakes

Only the highest shore features of Lake Iroquois are shown in Plate 1. They increase in altitude eastward, as described by Fairchild (1909; 1916), from about 430–440 feet at the northern end of the Butternut valley to about 460 feet around Oneida. Above the Iroquois shore features, numerous very small, but nevertheless distinct, shore features occur east of the Chittenango valley, extending up to a maximum altitude of about 515 feet. As many as nine faint steps can be distinguished in places in a vertical interval of 40 feet. Bars are small and uncommon and occur only where abundant unconsolidated material was available from kame areas as at D31 and 32. These kame areas are cliffed in places, the cliff base being at about 500 feet. Subdued mounds of sand and gravel that occur below this altitude near the kame areas (above the level affected by Lake Iroquois) appear to be wave-modified kames.

The outlet (or outlets) of the lake in which these features formed was presumably east of Oneida. The absence of evidence of meltwater erosion related to the last retreat of the ice below an altitude of about 500 feet east of the Chittenango valley indicates that the lake existed there while the ice withdrew. To the south and east of C13 a large area of sand and gravel, delta-bedded in places, has a flat top at about 500 feet. The position and bedding of this deposit show that it was laid down by meltwaters that had come through channel C12. Therefore this channel originally discharged into the lake. These sand and gravel deposits are, however, dissected by large channels. Furthermore, the floor of channel C12 northeast of the abandoned waterfall site is below 420 feet in altitude. The lake must have been drained, therefore, while channel C12 was being used by Great Lakes waters. Considerably later, submergence to altitudes up to 460 feet occurred owing to encroachment of Lake Iroquois as a result of isostatic tilting.

Fairchild's Syracuse-Oneida Moraine

Fairchild (1932b) described an end moraine extending from Syracuse to Oneida along the hill-foot zone and continuing both west and east beyond the limits of the area considered here. The moraine as marked on Fairchild's map (1932b, opp. p. 627) includes the kames and kame terraces south of Oneida, the kames along the northern end of spur D, the outwash terraces at the northern ends of the Cowaselon, Limestone and Butternut valleys, and the dissected outwash south and east of C13. Most of these deposits cannot legitimately be described as "moraine", and their relationships to each other and to the meltwater channels show that they were not formed contemporaneously. Since there appears to be no other evidence of end moraines along this belt it must be concluded that the Syracuse-Oneida moraine does not exist.

Evidence of More Than One Period of Great Lakes Drainage

Evidence indicates that Great Lakes waters flowed through the Syracuse-Oneida area on at least two occasions. Channel E7 appears to have been cut by these waters, for it is a quarter of a mile broad, its northern side is up to 70 feet high, and its southern side is up to 160 feet high. Its eastern end, however, is almost blocked by a mass of sand and gravel forming kames and kame terraces that extend up to 60 feet above the channel floor. These deposits show that channel E7 was not formed during the last retreat of the ice from the area. The floor altitude of channel D34 suggests that it was formed contemporaneously with channel E7. That channel D34 was not formed during the last retreat of the ice is also suggested by the absence of related outwash deposits from its eastern end.

Channel D33 is 4½ miles long and ranges in width between a quarter of a mile and half a mile. The eastern end of this channel is blocked by a mass of sand whose nearly flat top, about 50 feet above the channel floor, is bounded by well-defined, sinuous ice-contact slopes, indicating that dead ice occupied the channel at some time after its formation by Great Lakes waters. The same conclusion is indicated for channel C11 by the low kames and the sand and gravel terrace that obscure the first half mile of its course.

Reasons have been given for concluding that channel B7 was not formed during the last retreat of the ice. Its position and altitude suggest that it carried Great Lakes waters that had passed through channel A10, which, along with channel A7, also appears to have been formed prior to the last retreat. Farther south on spur A the great loop channel on the south side of channel A3 was apparently partly occupied by ice when the west-east part of channel A3 was carrying Great Lakes waters during the last retreat. The entrance to the loop is occupied by a mass of sand and gravel up to 50 feet thick, which has an irregular boundary on its southern side suggesting an ice-contact slope. A detached mass of dead ice in the loop would explain the sudden termination of the sand and gravel deposit on its southern side and would also account for the marshy area, suggestive of a kettle, immediately south of the deposit. In the eastern part of the loop other sand and gravel deposits form mounds and a small terrace. These mounds appear to be kames formed in association with dead ice, and not the result of dissection of an outwash mass by the present stream, for a section in one of them shows that the dip of the beds accords approximately with the slopes of the feature.

There is thus adequate evidence that some of the great channels in the Syracuse-Oneida area were occupied by dead ice during the final deglaciation. It is possible to argue that this resulted from frequent minor oscillations of the ice margin during the final retreat, so that after a certain channel was formed the ice advanced again and occupied it, subsequently becoming stagnant therein. This explanation seems inadequate, however, since, apart from the complex ice-margin oscillations it would necessitate, it does not accord with the pattern of retreat established from the marginal, subglacial, and englacial drainage evidence discussed heretofore, and the numerous moraines that one might expect to have been formed at the limits of at least some of the readvances do not exist. It appears, then, that the evidence indicates at least two periods of Great Lakes drainage through the area.

In addition to the channels just discussed, certain other major channels, which did carry Great Lakes waters during the last deglaciation of the area, were formed for the most part at an earlier date. The waterfall sites along the southern walls of channels B6 and A5 indicate that these channels were already in existence before the last retreat and were occupied by ice. At the bottom of the waterfall site on the southern side of channel A5 is a deposit of sand and gravel containing limestone blocks up to 10 feet across. It was laid down at the time the waterfalls functioned and was not swept away subsequently by the waters flowing through A5 since it lies in the embayment on the southern side of the channel. The level of the bedrock surface beneath the deposit is approximately the same as the level of the bedrock floor of channel A5. This implies that this channel had been excavated to its present depth before the last retreat of the ice. That channel B6 was excavated to its present depth prior to the last retreat is suggested by the accordant junction with it of channel B5. Since the relative altitudes of channels A5, B6, and C12 show that they functioned at the same time, it follows that C12 (at least above the waterfall site) was also excavated to its present depth prior to the last retreat. The same conclusion applies to channel A3, since the level of its bedrock floor is approximately the same as that of the bed-

rock surface beneath the sands and gravels in the big loop channel on its southern side.

Thus, four great channels in the Syracuse area, which were utilized by Great Lakes waters during the final deglaciation, as terraced outwash masses prove, appear to have been excavated to their present depths at some time prior to the last retreat. The main erosional effects of the Great Lakes waters in these channels during the last retreat were the addition or modification of the embayments due to waterfalls on the southern sides of A5 and B6, and the modification of the waterfall sites at the eastern ends of A3, A5, and C12.

The drumlins near Syracuse suggest that two periods of Great Lakes drainage occurred during the oscillatory retreat from the Wisconsin maximum. In the small area north of channel A5 and west of the Butternut valley are about 50 drumlins. Many of these drumlins are 100–150 feet high, 2000–3000 feet long, and 700–1000 feet broad. Within the same area, and commonly lying between the big drumlins, are smaller drumlins, which are generally between 10 and 40 feet high, 800 to 2500 feet long, and 200 to 300 feet broad. Thus the smaller drumlins are long, low, and narrow compared with the others. If, as appears to be generally accepted, the form and size of drumlins is related, in part, to the intensity of former ice movement, it would seem that two distinct periods of drumlin formation occurred in this area, the larger drumlins having been formed during the earlier period. This is also suggested by the existence on several of the bigger drumlins of long, low, narrow tails of dimensions similar to those of the smaller drumlins.

The final utilization of the large channel A7 by Great Lakes waters postdates the period of formation of the large drumlins, since the channel cuts cleanly across two of them. It predates the formation of the small drumlins since two of them (A8) extend a short distance onto the channel floor, and their rounded forms are unmodified, whereas a third unmodified small drumlin (A6) extends into the area that Great Lakes waters probably flowed through before they plunged over the limestone cliff at the head of channel A7. That the channel was not utilized by glacial waters during the final deglaciation is also indicated by its slightly irregular, partly till-covered floor, the small, partly obscured, plunge basins at its head, and the reverse slope at its eastern end due to outwash deposited by waters flowing through channel A5.

Since it seems highly improbable that the large drumlins have survived from a pre-Wisconsin glaciation, it appears that the evidence associated with channel A7 indicates two phases of Great Lakes drainage through the Syracuse area during Wisconsin times. This accords with Fairchild's conclusion (1932a; 1932b) which is accepted by Flint (1953; 1956; 1957). In terms of Flint's suggested correlations the small drumlins may have been produced by ice advancing to the Valders maximum, one period of Great Lakes drainage through the area having taken place during the advance to that maximum, the other during the retreat from it. If only these two periods of Great Lakes drainage through the Syracuse-Oneida area occurred, then the preceding retreat (culminating in the Ontario basin equivalent of the Bowmanville low-water phase) did not result in Great Lakes drainage through the area, being associated only with local drainage as Fairchild concluded (1932a; 1932b). The evidence for this conclusion of Fairchild lies outside the area considered here and cannot be discussed therefore. It may be remarked, however, that his conclusion may require modification in the light of modern ideas on glacial drainage. For example, his interpretation of the low-level channels (*e.g.*, Fairport-Lyons and Victor-Phelps) as marginal features encounters many difficulties, and there are good reasons for believing that these channels were formed by Great Lakes waters flowing across the dead marginal zone of the ice itself between walls of ice.

Whether or not only two periods of Great Lakes drainage through the Syracuse-Oneida area occurred in Wisconsin times, this writer suggests that the vast channel system has a much longer and more complex history than the preceding discussion might appear to indicate. Firstly, it seems reasonable to presume that the Great Lakes drained through the area in pre-Wisconsin times, for the ice borders appear to have been broadly similar during the different glaciations. Secondly, there is evidence of Great Lakes drainage before the formation of the larger drumlins. Thus the drumlin that forms part of the southern side of channel B7 belongs to this group, indicating that this channel was formed before these drumlins. Channel A10 was formed before the smaller group of drumlins, since two of these ascend its southern slope from the level of its floor (A9). This channel also appears to have been formed before the larger drumlins were produced (although it may have been utilized later), for the distribu-

tion of these drumlins appears to be partly controlled by it: to the south of the channel these drumlins commence exactly at its lip, although none of them have been eroded by the waters that cut the channel. Thus channels A10 and B7 both suggest Great Lakes drainage before the larger drumlins were formed, and these indications are mutually consistent, for the relative position and altitude of the two channels strongly suggest that they functioned together. Finally, the author finds it difficult to believe that the tens of miles of large channels could have been produced by only two periods of Great Lakes flow through the area, bearing in mind the rate of retreat of Niagara Falls, which is of dimensions comparable with many of the falls that formerly existed in the Syracuse area, and the limited time available, as indicated by carbon-14 dates. In the Syracuse-Oneida area alone there are nearly 40 miles of channels about 100 or more feet deep and averaging about a quarter of a mile wide, produced by Great Lakes waters, besides many smaller channels cut by these waters. The total length of channels of these dimensions cut by eastward-draining Great Lakes waters in the whole of central New York is greatly in excess of this figure. It has been shown heretofore that the last period of Great Lakes drainage did not utilize many of the large channels, and it produced only limited modifications of several other great channels. This means that, if only two periods of Great Lakes drainage through the area are postulated, almost all the vast channel system was produced during the earlier period. It seems much more likely that Great Lakes waters have flowed through this area on a number of occasions, each occasion after the first resulting in limited additions and modifications, until eventually the present vast channel system was produced.

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EROSION SURFACES, CYCLIC SLOPES AND DRAINAGE SYSTEMS IN SOUTHERN SCOTLAND AND NORTHERN ENGLAND

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THE origins of the erosion surfaces and drainage systems of southern Scotland and northern England, as of other parts of Britain, have been explained in various ways and there is still no agreement concerning most of the major issues. This paper discusses some of the interpretations that have been published and puts forward a hypothesis that may help to reconcile some of the divergent views prevailing at present. The area considered extends from the Southern Upland Fault southward to the Stainmore depression and includes the Lake District.

Since the sedimentary sequence in this area terminates at the beginning of the Jurassic and is not resumed till the Pleistocene, this attempt to explain the origin of the major relief features is based mainly on a study of the three aspects of the drainage and landforms that are especially important. These are: (i) groups of streams that, owing to their apparently discordant relationships to the present geological structures, appear to be satisfactorily explicable only by postulating their superimposition from an initial slope or slopes coincident with, or passing above, the adjacent hill-tops (i.e. consequent streams); (ii) areas of subdued relief, at any altitude, developed across rocks of variable resistance to erosion, that appear to indicate that at the time they were initiated the levels down to which erosional processes were able to operate were higher than at present (i.e. erosion surfaces); and (iii) the slopes separating the erosion surfaces, which may be termed cyclic slopes if, as is generally accepted, the erosion surfaces are related to past higher base-levels.

Most of the published views on the evolution of the major landforms of southern Scotland and northern England are concerned largely or entirely with discordant streams. They fall into three main types: (i) hypotheses based on the assumption that the original streams flowed mainly in a single direction over large areas; (ii) a hypothesis that recognizes groups of discordant streams flowing in several directions, but combines them with trunk streams flowing in a single direction; and (iii) hypotheses in which unidirectional discordant streams are not recognized.

The first type of hypothesis has been advocated by A. Geikie, H. J. Mackinder, B. N. Peach and J. Horne, F. Mort, J. W. Gregory, F. M. Trotter, M. and A. G. Macgregor, J. Pringle, and others.[1] These authors (excepting Trotter, who did not consider Scotland) favoured an original south-eastward-flowing drainage over a large part of Scotland, Mackinder, Peach and Horne, Gregory, Mort, the Macgregors and Pringle suggesting that the original south-eastward-flowing streams of the Southern Uplands had their sources in the Highlands.

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The second type of hypothesis has been put forward by D. L. Linton.[2] He has concluded that the discordant drainage of most of Britain was initiated on a widespread cover of Chalk and implies that this Chalk surface, in addition to being tilted towards the east, was warped along west-east lines. On this interpretation the original Tweed flowed approximately eastward along a synclinal axis and ran parallel with west-east watersheds on either side corresponding with anticlinal axes. The present Tweed is thus essentially a trunk stream towards which flow consequent streams from the south and from the north-north-west.

The third type of hypothesis has been advocated by A. Stevens, S. E. Hollingworth, T. N. George and others. It has long been accepted that the initial drainage of the Lake District was radial. Stevens[3] stated that in his opinion 'the drainage of the Southern Uplands developed on a roll of the ground whose core was Silurian, and whose crest lay along the line of the present fault which runs inland from Dunbar'. George[4] has suggested that the drainage basins of the Clyde, Nith and Annan are related to 'Hollingworth's platform residuals as initial centres of stream deployment'. Hollingworth[5] concluded from a statistical study of one-inch Ordnance Survey sheets that a series of essentially unwarped erosion surfaces exists in south-west Scotland, the Cheviot Hills and the Lake District (and in other areas of western Britain to the south), and considered these surfaces to be most satisfactorily accounted for as having been formed by marine erosion. This led him to state[6] that 'considerable revision of current theories of the evolution of British rivers will ultimately be required'. He suggested[7] that the earliest river system may have been 'superimposed on the older rocks beneath the Cretaceous cover during a great continental period which was followed by a big submergence in Miocene-Pliocene times', the rivers formed during the subsequent intermittent emergence in some cases following the earlier river courses and in others developing new courses.

The first type of hypothesis need not be discussed here, since Linton[8] has given adequate reasons for considering it unsatisfactory, pointing out, for example, that the discordant streams that flow northward are not explained by it. The second and third types of hypothesis are best considered in relation to a map of discordant streams (Fig. 1). This map was constructed by comparing the course of every stream in southern Scotland and northern England with the geology as shown on the Geological Survey one-inch sheets, this map evidence being supplemented by reference to Geological Survey memoirs and other geological publications. All streams exceeding 2 miles in length whose courses do not appear to be controlled by differential rock resistance, faults or shatter-belts are shown. In the case of streams whose courses appear to be partly related to the structures, and partly unrelated, only the unrelated parts are mapped. Blank areas on the map occur where there is insufficient information for the relation of streams to structures to be determined or where there has been considerable glacial modification of the drainage.

It is probable that some of the streams shown on the map are subsequent streams that have developed, for example, along unrecognized faults, while a few others probably owe their present courses to the effects of glaciation. The only hypothesis that appears to account satisfactorily for the origin of the great majority of the discordant streams, however, is that of superimposition. In any particular part of the area almost all the discordant streams flow in a similar direction and only a small minority have the opposite direction of flow. The former, which often flow in the same direction as the regional dip of the post-Silurian rocks, are presumed to be consequent streams and the latter obsequent streams.

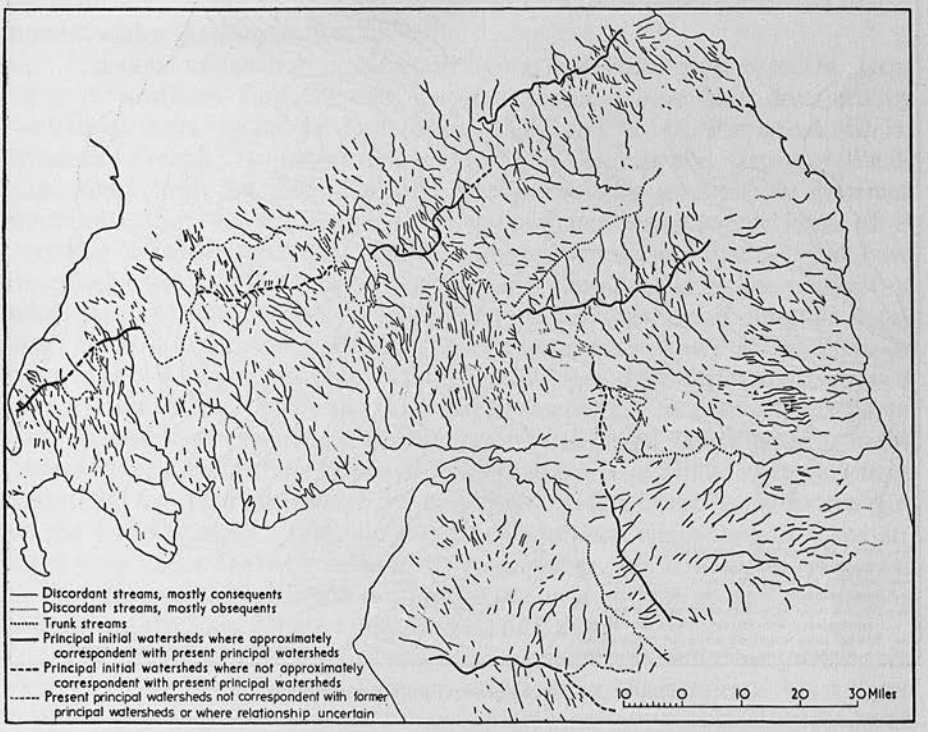


FIGURE 1—Discordant streams and principal watersheds.

Figure 1 shows that these presumed consequent streams of southern Scotland and northern England tend to radiate from the major upland areas or to flow in opposite directions away from the line of highest ground, and to converge on the major lowland areas in which trunk streams flow or on the Solway Firth. It will be noted that in the Tweed drainage area the consequent streams converge on the Teviot, whose line is continued by the lower Tweed. This is not in accord with the views of Mackinder, Mort, Gregory and Linton, who considered the Tweed to be the principal initial stream of this area. It

seems that in reaching this conclusion these authors were influenced by the Biggar gap, which leads from the Clyde valley to the Tweed valley, and which they interpreted as having been cut by an early Tweed. It is hoped to discuss the Biggar gap in detail elsewhere. For the present it may be remarked that this gap is not the only one in the upper Tweed region and that it appears to be unrelated to the present discordant stream pattern (as may be seen in Fig. 1 where it is marked — aligned east, then east-south-east, then east again — since it contains a very small stream). It seems likely that the Biggar gap owes

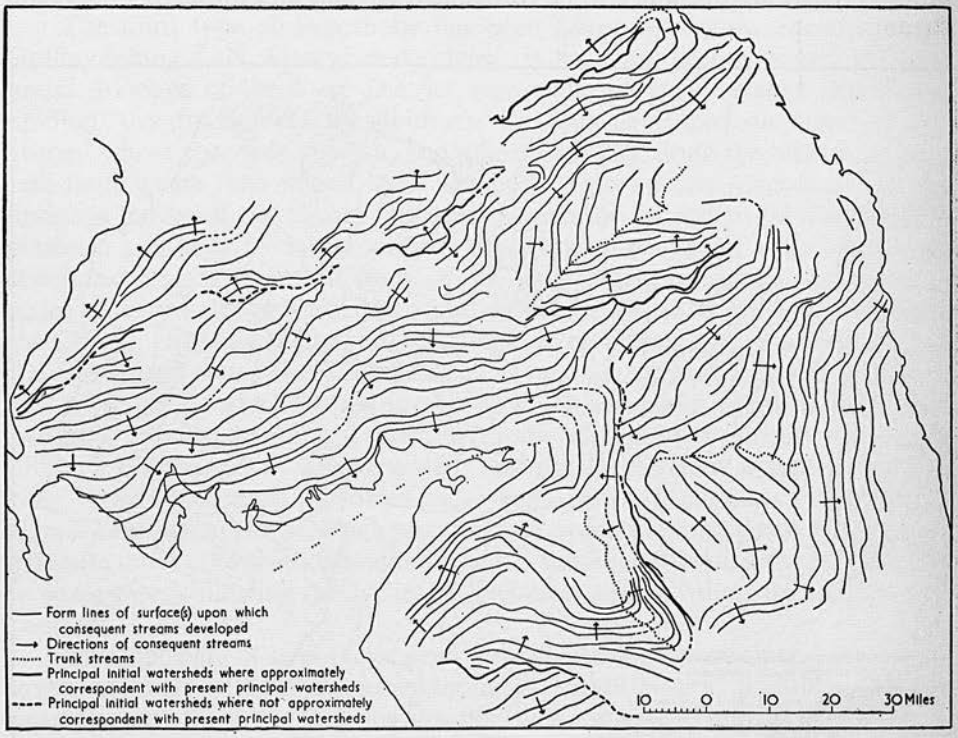


FIGURE 2—Form lines and initial watersheds.

its present dimensions to glacial erosion and that, like many other similar gaps in Scotland, it is of little or no significance in the reconstruction of former drainage systems.

The discordant drainage of southern Scotland and northern England as shown in Figure 1 appears to be more satisfactorily explicable in terms of the views of Hollingworth and George than in terms of the views of those who consider the initial drainage to have developed partly or wholly in relation to a unidirectionally inclined surface. Figure 2 attempts to emphasize this point and to depict the general form of the surface or surfaces on which the initial drainage commenced to flow. If, as appears to be generally accepted, conse-

quent streams generally flowed down the maximum slope of the surface(s) upon which they developed, lines drawn on a map at right angles to consequent streams will approximate to form-lines of the initial surface(s). O. T. Jones[9] used this method in studying the drainage of Wales and it has been applied in much greater detail in Figure 2. The spacing of the form-lines is arbitrary and was determined by cartographic considerations. The fairly uniform spacing of the lines on a given part of the map does not therefore necessarily imply a uniform initial slope, but the *directions* of the initial slopes indicated by the form-lines are very similar to the slopes of the present major relief features. This correlation strongly suggests that the consequent streams were developed on a surface or surfaces whose general form was very similar to that of the present major landforms.

The same conclusion is suggested by the evidence of river captures. Captures in northern England and southern Scotland have been described by various authors, including J. E. Marr, Marr and W. G. Fearnsides, G. H. Mitchell, Trotter, Hollingworth, Linton, J. A. McIver and George.[10] Wind-gaps show that the great majority of the drainage adjustments described by these authors have taken place below the present local summit levels. It is arguable, however, that this is not significant since these authors may have described river captures only where captures are indicated by the presence of wind-gaps. One of Hollingworth's papers[11] is not open to this objection, however, since it comprises a detailed and comprehensive study of the drainage adjustments of a limited area. Hollingworth studied the left-bank tributaries of the river Eden. Analysis of his paper shows that he describes sixty-four instances of river capture, of which sixty are associated with wind-gaps in the appropriate divides. The floors of these gaps occur at altitudes varying from about 350 feet near the river Eden to about 2000 feet in the north-eastern part of the Lake District. If Hollingworth's deductions concerning captures are valid they imply that the present drainage system was initiated at, or not far above, the present local summit levels (i.e. on a surface or surfaces sloping down from the Lake District into the Vale of Eden).

The depth of the wind-gaps is also significant. Of the sixty wind-gaps recognized by Hollingworth, forty-six are less than 350 feet deep. Most of the fourteen gaps with depths greater than 350 feet have been considerably deepened by ice or meltwaters. If the gaps that have been definitely deepened in this manner are omitted, there remain forty-five gaps, of which forty-two are less than 350 feet deep. The average depth of the forty-five is about 180 feet.

Comparison of the altitude of the floor of each wind-gap with the altitude of the associated elbow of capture shows that there has usually been several hundred feet of stream incision since capture occurred. The average amount of incision since capture for the forty-five cases under consideration is approximately 390 feet.

If Hollingworth's deductions are valid the evidence presented above is very difficult to reconcile with the view that the present drainage of western

Edenside was initiated on a Chalk cover in the manner envisaged by Linton. According to Linton's map[12] the restored sub-Cenomanian surface would be at an altitude of about 3000 feet in this area. On his hypothesis, therefore, the consequent streams of western Edenside, having breached the Chalk, must have cut down through many hundreds of feet of varied strata over a long period of time during which only four captures occurred (i.e. the four captures not associated with wind-gaps); then, for no obvious reason, numerous captures occurred when the streams had cut down to just below the present local summit levels; captures then ceased or almost ceased until glacial times (Fig. 3). It is suggested that such a sequence of events is highly improbable. If, however, the drainage system was initiated at or not much above the present local summit levels on a surface or surfaces declining in altitude towards the Vale of Eden, the numerous relatively shallow wind-gaps can be explained as due to the rapid development of subsequent streams, resulting in captures, before the newly developed drainage system had become deeply incised; captures then ceasing or almost ceasing, once the drainage had become well established, until the advent of glaciation.

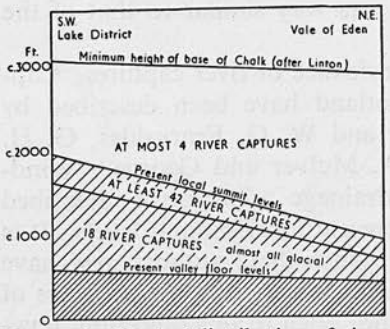


FIGURE 3—Vertical distribution of river captures recognized by S. E. Hollingworth west of the river Eden. For explanation, see text.

The slopes of the upland areas of southern Scotland and northern England have so far been referred to in general terms as if they are smooth slopes. Various geologists and geographers have shown, however, that these slopes consist of series of dissected erosion surfaces. The statistical work of Hollingworth[13] in the Lake District, Cheviot Hills and south-west Scotland has been referred to above. A. G. Ogilvie[14] published a map showing erosion surfaces in south-east Scotland and C. A. Halstead[15] has reported on a hypsographic analysis of the whole of the Southern Uplands. Erosion surfaces recognized in the field in the central Southern Uplands have been described by George,[16] in Galloway by W. G. Jardine,[17] in the Lake District by Hollingworth[18] and R. B. McConnell,[19] and in the Cheviot Hills by R. Common.[20] Opinions differ as to whether the surfaces recognized are of subaerial or marine origin. Neither is it established whether the surfaces (or some of them) are horizontal or not. Hollingworth concluded from his map analyses that the surfaces are essentially horizontal but Halstead, also from map analyses, concluded that the surfaces in southern Scotland are inclined.

Although there is no agreement as yet concerning certain important aspects of the erosion surfaces of southern Scotland and northern England, it has been adequately demonstrated that the interfluves of the major upland areas are characterized by flights of erosion surfaces and that these surfaces often extend along the sides of the major valleys. In the preceding pages it has been shown that the discordant streams radiate from the major upland areas.

It is often on the crests of the spurs between these streams that erosion surfaces have been identified. It therefore seems highly probable that a close relationship exists between these spur-crest erosion surfaces and the discordant streams. This relationship can be explained if these surfaces are considered to have been formed by marine erosion (or to have developed from surfaces of marine erosion), for the discordant streams can then be accounted for as having formed by extension across the erosion surfaces as these surfaces were successively formed and exposed during a period of intermittent emergence. Such an interpretation has been suggested by Hollingworth[21] for the Lake District, Cheviots and south-west Scotland (and other areas farther south) and is favoured by George[22] for part of the central Southern Uplands. This interpretation is also in accord with the conclusion reached above, from the evidence of river captures, that the present drainage was initiated at, or not very far above, the present local summit levels.

The view that the present consequent drainage of upland Britain or of parts of it developed on a series of marine erosion surfaces has received only limited support. Many geomorphologists appear unwilling to consider this hypothesis seriously because of the apparent absence of Tertiary marine deposits over most of Britain, and this point is repeatedly raised as an argument against the hypothesis. It is suggested, however, that this argument is far from conclusive on the following grounds:

(i) By analogy with the rock platforms forming parts of our present shores or with the recently uplifted rock platforms of western Scotland, where deposits are generally of very limited extent and thickness, it seems reasonable to suggest that any such deposits laid down on marine erosion surfaces developed on the slopes of the uplands of southern Scotland and northern England were initially of very limited extent and thickness also. It also appears probable, on the hypothesis advanced here, that marine deposits laid down offshore at any stage in the emergence of this area would be subsequently eroded and transported farther seawards with each fall of sea-level as they came within the zone of erosion, ultimately finding their present resting places on the existing sea floors.

(ii) If the removal of any remaining marine deposits appears to present difficulties it should be noted that these are less serious than the difficulties encountered in this respect by some other hypotheses. For example, the Chalk cover hypothesis as put forward by Linton requires the removal of some 3000 to 4000 feet of Jurassic rocks from the eastern part of the Irish Sea area 'by normal erosion in Tertiary time',[23] a volume of rock amounting to several thousand cubic miles.

(iii) In England Tertiary marine deposits have been located only in the unglaciated south or just within the glaciated area where the ice-sheets were depositional rather than erosional agents. This strongly suggests that the absence of marine deposits farther north in England, and in southern Scotland, is largely due to glacial action, probably assisted by periglacial processes. On the other hand, if this suggestion is not accepted and it is argued that erosion

surfaces can be of marine origin only where marine deposits lie on them, one must conclude that not only has the London Basin west of a line through about Hitchin recently been uplifted by several hundred feet relative to East Anglia,[24] but that this part of the London Basin has been recently uplifted by a similar amount relative to most of the rest of England.

(iv) The fact is, of course, that few, if any, British geomorphologists insist on such a rigid requirement in all cases, and many erosion surfaces at lower levels in southern England and the peripheral areas of Wales have been repeatedly interpreted as of marine origin although marine deposits are more often than not entirely lacking. Certain geographers, however, appear unwilling to accept erosion surfaces above about 700 feet as of marine origin, even where the general form and distribution of these surfaces are the same as those of surfaces below this level in the same area. This attitude appears to be largely due to the significance attached to Wooldridge and Linton's interpretation of south-east England, for these authors concluded that the 'Pliocene' submergence in this area did not exceed 700 feet and that those parts of south-east England situated above this altitude are in a second cycle of erosion following peneplanation.[25] This interpretation is not entirely irreconcilable with the view that marine erosion surfaces exist at altitudes exceeding 700 feet elsewhere in Britain as pointed out by W. G. V. Balchin and admitted by Wooldridge.[26] The Mio-Pliocene peneplain of south-east England may merely represent an exceptionally long still-stand in the general emergence of Britain. Alternatively, it is possible that some of Pinchemel's criticisms of Wooldridge and Linton's work are valid and that the drainage of the whole of that area is in a first cycle of erosion,[27] in which case events in south-east England can be more readily reconciled with the evidence of marine erosion surfaces at higher levels elsewhere in Britain.

In the preceding pages it has been implied that many of the erosion surfaces that form steps on the slopes of the upland areas of northern England and southern Scotland are of marine origin in the sense that they were actually cut by the waves of the sea. This interpretation represents the most extreme form of the marine erosion surface hypothesis and has been suggested, for various parts of upland Britain, by Hollingworth, E. H. Brown, Balchin, George and the present writer.[28] At the other extreme are those, such as Linton, K. M. Clayton, H. R. Wilkinson and S. Gregory and (more recently) Brown,[29] who favour the view that erosion surfaces in the higher parts of Britain are entirely of subaerial origin, although the way in which they were formed is seldom discussed in detail. These two modes of origin generally seem to be regarded as incompatible, but it seems to the author that to a considerable extent they can be reconciled.

Those who favour the wholly subaerial hypothesis often appear to fail to take into account the fact that, contemporaneously with the formation of each of the subaerial surfaces they have described, a marine abrasion platform must have been formed also. The early stages of the formation of such platforms are relatively rapid, as is witnessed by the rock platforms around parts of our

present shores and by platforms recently raised above sea-level in western Scotland or buried beneath drift in Holderness. These features, it will be noted, are not associated with subaerial erosion surfaces of comparable dimensions. Thus the subaerial planation hypothesis requires the formation and subsequent destruction of a series of marine platforms, although this is not pointed out by its advocates.

On the other hand, the marine hypothesis fails to take into account the probability of significant displacement of at least the older marine erosion surfaces by subaerial agencies. Dissection of the surfaces by rivers, differential weathering and erosion of the rocks across which the surfaces were cut, and a reduction in the slope of the original cliff have been described by various authors, but the possibility that the surfaces themselves have been displaced appears to have been largely neglected. That an uplifted marine surface can remain in its original location for a long period of time seems to the author highly improbable in the light of the evidence of slope retreat that has been rapidly increasing in recent years. It seems likely that the cliff-lines of at least the older marine surfaces have retreated under the influence of weathering and mass-movement, resulting in the backward extension of the surfaces at the bases of the cliffs, possibly in some instances causing the original marine surfaces to be replaced by surfaces of subaerial origin. It is suggested that the apparently opposed subaerial and marine hypotheses regarding the origin of erosion surfaces might well be largely reconciled on these lines.

Since, despite the work of W. Penck, L. C. King and others, there appears to be some reluctance on the part of many British geomorphologists to recognize slope retreat without active undercutting by the sea or rivers as a major process responsible for the formation or extension of erosion surfaces, some elaboration of the preceding paragraph is required. That such slope retreat is an active agent in erosion surface development is illustrated by valley-side benches. Where river deposits lie on such benches lateral river planation appears to account satisfactorily for the origin of the benches, but where, as is almost always the case at levels more than at most 200 or 300 feet above present valley floors, river deposits are absent, this explanation is not necessarily valid. In fact, as one ascends a sequence of valley-side benches an explanation purely in terms of river planation becomes increasingly difficult to apply, for it encounters the obvious difficulty that it requires a river to have become more and more constricted in its wanderings as incision progressed. This difficulty is greatest when the benches of adjacent valleys are traced downstream and found to merge together as an extensive bench extending across the crest of the intervening spur. It may appear that one way round this difficulty is to argue that the benches truncating the spur crests are of marine origin and that these benches grade into river-cut benches extending along the valley sides. This explanation is clearly unsatisfactory, however, when such features are developed in locations where surrounding higher ground renders former marine erosion highly improbable. Since, therefore, active undercutting by neither a stream nor

the sea can account for many such erosion surfaces, it appears that the only possible explanation is that they have been produced at the bases of slopes retreating owing to weathering, mass-movement, gullying, rill-wash and related processes. Since it has been shown by many authors, a few of whom have been cited above, that the erosional benches on many valley sides can be correlated on the basis of altitude and are related to former base-levels, it follows that these surfaces have developed from river-cut surfaces of limited width now destroyed by later stream incision. The surfaces on the crests of interfluves, into which the valley-side benches merge, may also be of subaerial origin, having been created by the intersection of slopes retreating from opposite directions. Alternatively they may be marine erosion surfaces, or may have developed from marine erosion surfaces, for such surfaces must have been extended by the same processes that have resulted in the extension of benches formed by river erosion.

Figure 4 attempts to illustrate these points diagrammatically. It is assumed that an area or homogeneous rock with equal and uniform slopes in opposite directions emerges intermittently from beneath the sea. It is also assumed that the periods of negative movement of base-level are of equal amount and duration, and that they occupy short periods of time compared with the periods of constant sea-level (all of equal duration) that separate them. During each period of constant sea-level the sea cuts into the land and produces a marine abrasion platform backed by a cliff. With each negative movement of base-level the streams are extended seaward across the newly exposed marine abrasion platform and associated marine deposits. During each period of constant sea-level a stream flowing into the sea attempts to adjust its course to the new sea-level and therefore becomes incised in its lower course, the incision extending upstream as the knick-point retreats. Thus each period of constant sea-level results in the production of cyclic slopes of marine and fluvial origin. When a negative movement of base-level causes the marine cliff to be abandoned by the sea the cliff continues to retreat under the influence of weathering, mass-movement and other processes. The valley-side slopes also continue to retreat under the influence of the same processes and are extended farther upstream as the knick-point works headwards.

In Figure 4, diagram A shows the imaginary area at the conclusion of the first period of constant sea-level. The cliffed coasts, incised streams and initial position of the shore-line are indicated. Diagram B shows the area at the end of a second period of constant sea-level preceded by a relatively rapid negative movement of base-level. The marine abrasion platforms related to sea-level 1 are beginning to be dissected as streams try to adjust themselves to sea-level 2. Meanwhile the slopes produced by stream incision and marine attack during sea-level 1 have retreated and the surfaces left at their bases are continuous with the marine platforms. In diagram C further slope retreat has produced a sinuous watershed between the stream sources in place of the initial straight one. By stage D the original surface has been completely destroyed in the summit

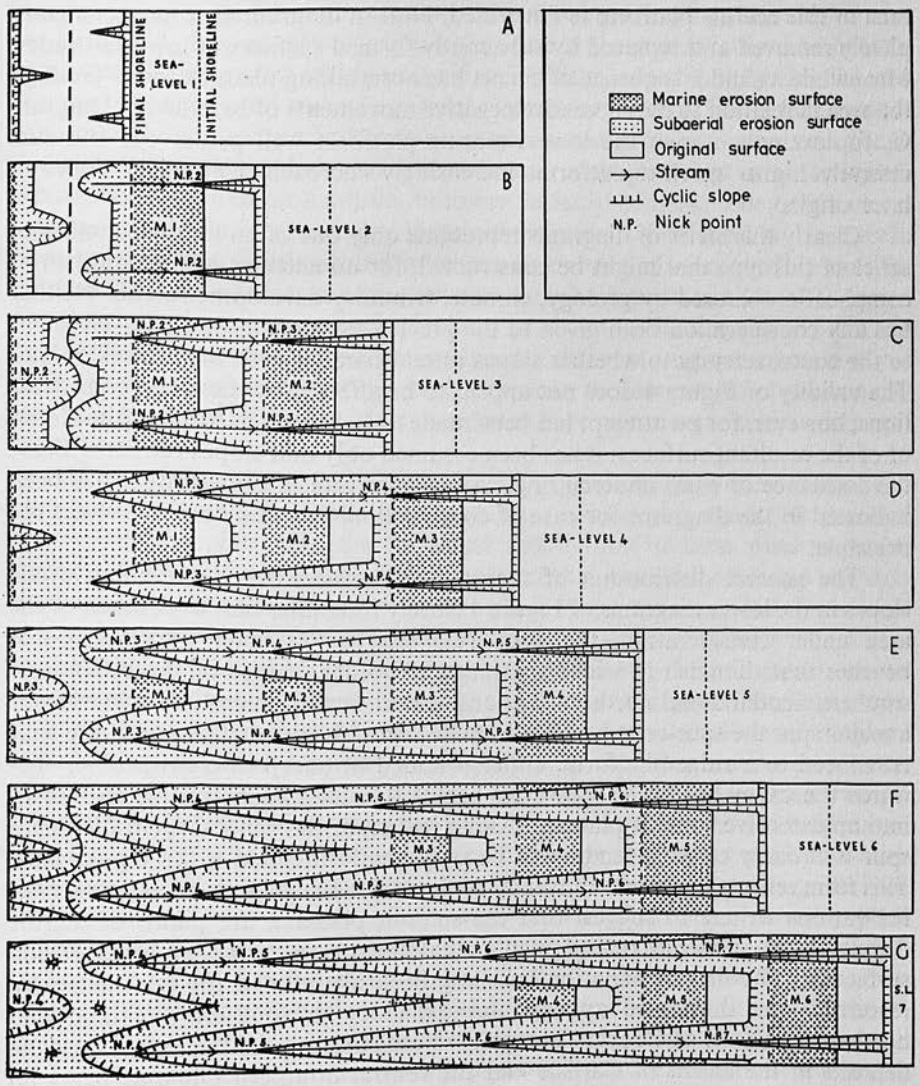


FIGURE 4—Diagrammatic representation of the evolution of polycyclic landforms in an area of uniform rock.

area, which is now an erosion surface formed by slope retreat under the influence of weathering, mass-movement and related processes, merging into the remains of marine platform 1. In diagrams E and F the further reduction in area of this marine platform is illustrated, until in diagram G it has been completely removed and replaced by subaerially-formed surfaces at lower altitudes. Meanwhile a similar sequence of events has been taking place at lower levels in the area in relation to the successive negative movements of base-level. Diagram G, for example, shows the lowest marine platform well preserved, and successively higher marine platforms increasingly encroached upon by slopes of later origin.

Clearly this series of diagrams represents only one of an infinite number of series of this type that might be constructed, for no account has been taken of complications caused by geology, climate, tributary development, etc. Neither has any consideration been given to the precise mechanism of slope retreat or to the controversy as to whether slopes retreat parallel with themselves or not. The validity of Figure 4 does not appear to be affected by these last considerations, however, for no attempt has been made to indicate the form of the slopes or of the resultant surfaces: it has been assumed only that slopes retreat without the assistance of basal undercutting. A constant rate of slope retreat has been indicated in the diagrams for ease of construction, but this does not affect the principle.

The general distribution of the erosion surfaces and intervening cyclic slopes in the lower diagrams of Figure 4 closely resembles that over much of the area under consideration, the spur-crest benches merging into valley-side benches that diminish in width upstream. Another feature of the uplands of southern Scotland and northern England is that very frequently as one ascends a major spur the spur-crest benches become successively narrower until the spur is reduced to a ridge-like form, either rounded or sharp in cross-profile. Towards the valley heads the spur crest begins to widen again and often merges into an extensive summit plateau. Along its length the general form of such a spur is broadly convex, tending to be only gently inclined in its higher part. This form reflects the broad convexity of the major British upland areas and has led various writers to suggest that the summit plateaux are plains of marine denudation,[30] subaerial peneplains[31] or exhumed sub-Permo-Triassic erosion surfaces.[32] The first explanation has long been abandoned for such extensive features. The third also appears unsatisfactory in southern Scotland and northern England: as George[33] has pointed out, the base of the Permo-Triassic deposits in the basins of Carlisle and the central Southern Uplands bears no relation to the present plateau surface of those areas, while the sub-Permo-Triassic surface in the Vale of Eden and eastern Durham is inclined much more steeply than the present slopes of the eastern Lake District and northern Pennines respectively.[34] All three interpretations require an ancient land surface to have survived in recognizable form over a very considerable period of time; and they are, therefore, open to the objection that they neglect the probability

of its very considerable modification owing to weathering, mass-movement and slope retreat under a variety of climatic conditions. It is possible that a fourth explanation, suggested by Figure 4, is of some significance.

In diagram G, for example, a well preserved marine bench is shown near the coastline, but at successively higher levels the marine benches become narrower, the spur crest taking on a ridge-like form. At still higher levels in the diagram the spur crest widens again into summit plateaux. The replacement of the original land surface shown in diagrams A and B by surfaces developed at the bases of retreating slopes implies a lowering of the higher parts of the region, which, when viewed as a whole, becomes broadly convex. The forms of the region depicted in Figure 4 are, therefore, very similar to those of the upland areas of southern Scotland and northern England, and it is possible that the summit plateaux of this area have developed in the manner shown in this figure. On this hypothesis the summit plateaux need not be distinguished from the erosional steps at lower levels and assigned a different origin as some authors have considered necessary: they simply represent the later stages of the destruction of a land surface by the same processes as are operating and have operated at lower levels.

If, as has been suggested above, the consequent drainage of southern Scotland and northern England developed on a series of successively exposed marine erosion platforms, the form and distribution of these platforms, and hence the directions of the consequent streams, would have been strongly influenced by the relief of the mass into which they were cut. Had the relief of this mass been considerable, the present consequent drainage pattern would have been complex, comprising both old and new elements. The simplicity of this pattern over most of southern Scotland and northern England suggests, therefore, that the relief of the land mass attacked by the sea was sufficiently subdued over most of the area to permit the formation, at each stage of the emergence, of marine platforms across which a new consequent drainage system of simple pattern could be extended when the level of the sea relative to the land subsequently fell. Unless, contrary to present theory, one assumes that marine planation can produce an extensive area of subdued relief, this surface must have been produced by subaerial agencies. It would therefore seem that a peneplain or pediplain, perhaps diversified by residuals developed on the most resistant rocks, such as the granites of south-west Scotland, had been developed across southern Scotland and northern England prior to the final submergence. In terms of the sequence recognized in south-east England such a surface would be the continuation of the Mio-Pliocene, sub-Eocene or sub-Cenomanian surfaces.

In order to produce the present consequent drainage pattern this peneplain or pediplain must have been warped and faulted before, and possibly during, the final emergence. The general form of this warped and faulted surface is shown in Figure 2. The fact that the initial watersheds are, as a whole, straight or gently curved over considerable distances is consistent with this hypothesis,

as also is their frequent correspondence (and that of the trunk streams) with old lines of warping and faulting established from geological evidence. The three major initial watersheds of southern Scotland and the Borders are therefore interpreted as anticlinal axes and the Teviot/lower Tweed line as a synclinal axis. Between south-west Scotland and the roughly east-west anticlinal axis of the Lake District, recognized by W. Hopkins[35] and Marr,[36] lies the Solway Firth syncline. Trotter's[37] conclusion that Tertiary uplift of the Alston Block occurred along the Pennine and Stublick faults, together with gentle warping along the Tees anticline is also in accord with the hypothesis put forward here (Fig. 5). In this way it is possible to account for the present distribution of the principal areas of high and low ground and for the fact that the latter areas correspond with the major outcrops of the less resistant rocks on which consequent streams converge. The peneplain or pediplain into which the marine erosion surfaces were cut is considered to have corresponded roughly with the generalized slope of the present local hill-tops at lower levels, but to have departed increasingly from it at higher levels owing to the increasing importance of slope retreat.

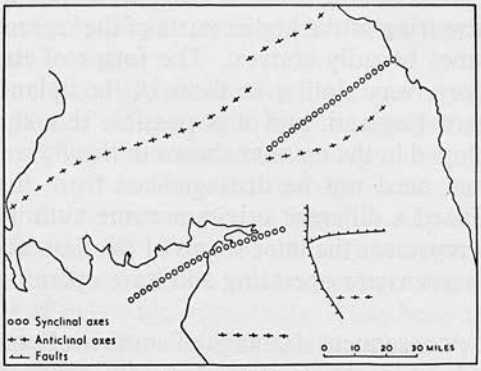


FIGURE 5—Suggested Tertiary structures.

In summary, the hypothesis put forward in the preceding pages is as follows:

(i) the reduction of southern Scotland and northern England principally by subaerial erosion to subdued relief, except perhaps for occasional residuals of the most resistant rocks;
(ii) submergence of the area accompanied by warping and some faulting;
(iii) intermittent emergence, possibly accompanied by further earth-movements, resulting in the development of marine platforms on which the present consequent drainage was initiated, these platforms and the related valley-floor surfaces having been subsequently extended by slope retreat.

ACKNOWLEDGMENT

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SOME ASPECTS OF GLACIAL DRAINAGE CHANNELS IN BRITAIN

PART I

J. B. SISSONS

THE interpretation of the glacial drainage channels of the Cleveland Hills put forward by Kendall in 1902 [1] has since that time been taken as the model for most other British studies of such channels. This interpretation assumed that many (perhaps most) channels were formed by the overflowing waters of ice-dammed lakes. It also assumed that meltwaters rarely, if ever, flowed under, over or through the ice. The ice was thus regarded as an almost impenetrable 'barrier' or 'dam' terms that have been subsequently used by many authors.

Kendall's interpretation is not in accord with the evidence from receding temperate glaciers of the present day, however. The classic studies of Russell, Tarr and von Engeln, as well as later studies, make it clear that almost always the land drainage and marginal drainage, as well as the supraglacial drainage, rapidly find their way into and beneath the ice itself. Where open lakes exist their waters may sometimes escape over cols, but far more often they are drained from time to time by englacial or subglacial tunnels. Furthermore, Kendall's hypothesis fails to explain many aspects of the form and distribution of glacial drainage channels.

In Part I of this paper these criticisms of Kendall's interpretation are amplified. This theme was introduced in an earlier paper [2]. In the present paper much additional material is presented. In Part I evidence in the British literature is discussed. It is shown that this evidence, rather than supporting Kendall's interpretation of the origins of meltwater channels, often points to conclusions differing considerably from those he reached. The difficulties encountered by many authors in applying Kendall's hypothesis to various parts of Britain, and considered by these authors to be exceptional, form the basis of a re-interpretation of the evidence in which interpretations of the type put forward by

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Kendall are themselves the exceptions. In Part II certain types and patterns of meltwater channels will be described and explanations of their modes of origin suggested, particular reference being made to the drainage of modern glaciers and to channel systems mapped by the writer in Britain.

Glacial drainage channels and the sites of former ice-dammed lakes had been described in various parts of Britain before Kendall published his well-known paper. For example, in 1869 A. Geikie provided a brief, clear description of channels in the upper Tweed valley, referring to the "remarkable hollows or trenches, like old water-courses, running along the slopes of some of the hills, but without any stream" and suggested: "Possibly they should be referred to a time when the drainage of the district was greatly modified by large accumulations of snow and ice [3]". The former existence of ice-dammed lakes had been deduced particularly by Carvill Lewis in the Pennines [4] and by various writers in the Glen Roy area [5].

The Glen Roy controversy is especially interesting since it throws light on the different views of glacier drainage that were held at the time. While Jamieson [6] argued for ice-dammed lakes overflowing solely through various cols, others attacked this view. For example, Prestwich pointed out that the Marjelen See (later to be mentioned by Kendall [7]) "has been adduced in illustration of the power of glacier barriers", but that "the water escapes almost every year, and the lake drains in the course of a few hours through crevasses in the great glacier [8]". He also mentioned the drainage of Himalayan ice-dammed lakes through crevasses in the ice. Prestwich also suggested that glacier ice could not withstand the hydrostatic pressure of such large lakes as Jamieson claimed for Glen Roy, a point that has been fairly recently discussed by Glen [9] with regard to ice-dammed lakes in general. Livingston [10], although he reached a different conclusion regarding the origin of the "roads", made similar criticisms to those of Prestwich concerning the lakes. Livingston mentioned the annual draining of Norwegian ice-dammed lakes. Concerning the Marjelen See he wrote: "If it proves anything at all, it proves that the supposed Lochaber lakes were an impossibility [11]". In a summary of his conclusions, he stated: "These barriers of decomposing and crevassed ice would be tunnelled underneath by a glacier river, from which, owing to the depth of the lakes behind the barriers, an enormous pressure would be exerted upon the over-arching ice, causing it to disrupt [12]".

These quotations, and others that might have been given, show that some British workers on problems of glacier drainage in the late nineteenth and early twentieth century were attempting to apply to Britain the evidence of modern glacier drainage. In the twenty years up to 1912 much further evidence became available from studies on Alaskan glaciers by Russell [13], Tarr [14], Von Engeln [15] and others, and from studies elsewhere by workers such as Tarr [16], Garwood [17] and Rabot [18]. These publications on Alaska still provide the best body of evidence on drainage conditions associated with temperate glaciers.

Kendall was familiar with this work, for he referred to it in some

of his publications. Yet he did not apply it. In his Cleveland Hills paper he nowhere referred to examples of drainage into, onto or beneath the ice, although he mentioned the theoretical possibility in his introduction [19]. Some twenty years later he still maintained the same hypothesis in the *Geology of Yorkshire.* This is clearly shown at one point in his discussion of channels when, having put forward what he obviously felt was a rather inadequate explanation for some channels in the Dallow Gill area of north Yorkshire, he went on to say: "The alternative is the even less plausible one of supposing a subglacial drainage system such as on occasion taps the waters of the chain of lakes margining the great Malaspina Glacier where it abuts against the Chaix Hills [20]".

Nevertheless Kendall was quite clear about the limitations of the evidence he presented for his ice-dammed-lake view. As noted later he was fully aware of the paucity of depositional evidence for the lakes and he built up his hypothesis on the assumption that most of the Cleveland channels are lake overflows. Discussing channels in his introduction he said: "I have placed my chief reliance upon them in the identification of the series of lakes which form the subject of this paper [21]". Yet he also recognised that many channels show features difficult to explain by his hypothesis. For example, the problem of up-and-down channels, which has recently provoked discussion [22], was early considered by him [23]. He noted too the fact that in some cols, where one would expect to find channels according to his hypothesis, there are no channels. He was also aware of the difficulty of applying his hypothesis to an upland where the channels show that, at the time many of them were cut, the upland was entirely surrounded by ice.

Despite Kendall's clear statements concerning the limitations of some of his own evidence and despite the quite different implications of drainage studies on existing glaciers, his views have received widespread acceptance. Rare criticisms, such as those of Bonney [24], were ineffective. One can only presume that the simplicity of Kendall's interpretation and the fact that, once put forward, it appeared so obvious, were major factors in this general acceptance. His views have been subsequently applied in numerous papers to many parts of Britain. After a time many writers did not even consider it necessary to present the detailed evidence for their conclusions. For example, Bremner, who carried out much valuable research on glacial problems in the Highlands, nevertheless wrote: "The cause, character and significance of overflow channels is now so well understood that there is no need for any detailed description of those in the area under discussion [25]".

Although papers in which Kendall's interpretation is accepted are still published quite frequently, doubts as to the validity of his views have been expressed by a few writers. Foremost among these is Carruthers [26], who believes that the lakes did not exist, their supposed sites being occupied by stagnant ice. Peel has pointed out that in Britain "thought has been perhaps unduly influenced by Kendall's concept of freely draining lakes despite the common and disturbing lack of any independent direct evidence for such lakes [27]". Common wrote: "The last word on the subject of meltwater channels has by

no means been written" and "there is still a great deal to be learnt about them and from them [28]", a statement that contrasts greatly with Bremner's opinion quoted above.

Despite the fact that Kendall's hypothesis has been widely accepted, there appears in almost every paper dealing with channels evidence that does not accord with his hypothesis. Most writers have considered such evidence in their respective areas as local minor anomalies and not as invalidating Kendall's interpretation. Analysis of the literature shows, however, that the same anomalies occur in many different areas and, taken together, they present a significant body of evidence that necessitates thorough reconsideration of the orthodox ice-dammed-lake hypothesis. Some of this evidence, as well as some other points revealed by the literature, is considered in the following pages.

The rarity of shoreline features. As pointed out elsewhere [29], almost all the supposed ice-dammed lake-sites in upland Britain lack definite shoreline features. For example, Kendall wrote in his classic paper: "I am unable to cite a single well-marked example in the Cleveland area [30]". Some twenty years later he extended this generalisation, saying: "Lake beaches are disappointingly infrequent in English areas [31]". It is arguable that this infrequency of shorelines in England may be due to subsequent weathering and mass-movement. This seems less likely, however, in Scotland, especially in the Highlands. Yet in the latter area Charlesworth [32] has recently catalogued and indicated on maps hundreds of former ice-dammed lakes without shorelines. Glen Roy, with adjacent glens, still remains the "brilliant exception [33]", the shorelines of Rannoch Moor being similar in form but much less extensively developed [34].

The rarity of definite deltas. Since many channels have clearly been cut by large volumes of water flowing with considerable velocity, there should be masses of sand and gravel at channel exits if these channels debouched into ice-dammed lakes, a result of the sudden checking of the velocity of the meltwaters. These deposits should be roughly proportional to the volume of rock removed in the excavation of the channels at whose exits they lie, due allowance being made for the finer materials carried further afield in solution and suspension. Furthermore, it is a reasonable expectation that, in the more recently glaciated parts of Britain, these deposits would show typical deltaic forms with surfaces gently inclined lakewards and steep frontal slopes. In the less recently glaciated parts such forms might be expected to occur at least at the exits of the larger channels, for they should have survived owing to their large volume and permeability.

Yet a study of the British literature on this subject shows that these conditions are rarely fulfilled. This is particularly significant when it is remembered that almost all authors have diligently sought such evidence in order to demonstrate the existence of lakes. At the lower ends of many channels, including some very large ones, no sand and gravel deposits at all have been recognised, a fact very difficult to explain if the channels emptied into lakes. At the lower ends of some other channels such deposits do exist, but they rarely display typical deltaic forms. These are the deposits that have been frequently interpreted as deltas. It is suggested that such interpretations would in many instances not

have been put forward had not their authors been looking for evidence of ice-dammed lakes. Kendall was quite clear about the lack of deltaic forms in Cleveland, saying that the deltas of that area are usually "simply patches of current-bedded sand and gravel [35]". As an example from the recent literature Twidale's paper on north Lincolnshire may be cited [36]. He mentions "numerous deposits of more or less rounded chalk gravels—presumably derived from the excavation of the channels [37]". No reference is made to delta forms. The folding map illustrating Twidale's paper shows thirty-five patches of chalk gravels and sands, but only seven in fact occur at the lower ends of channels. Half the "overflow" channels shown on the map have no deposits at all at their lower ends, while in two instances chalk gravels and sands occur by the channel intakes.

The general absence of definite deltas, along with that of shoreline features, from the great majority of the supposed ice-dammed lake-sites of Britain must surely mean that the great majority of the lakes of the dimensions postulated did not exist. If the supposed lake-sites were largely or entirely occupied by the ice itself the absence of deltas as well as many difficulties that have been encountered in explaining channels can be resolved *(see Part II)*. Where deposits appear to be absent from the lower ends of channels this is probably because the streams that cut the channels continued to flow unchecked into (or, probably less commonly, onto) the ice itself. Ill-defined spreads of sand and gravel at the lower ends of channels (where not a result of later mass-movement and weathering) may be due to the streams that cut the channels having had their velocity checked on flowing into (or, perhaps, onto) the ice, so that they began to deposit, these deposits having been subsequently redistributed and possibly mixed with ablation moraine on being let down onto the ground as the ice wasted away. Where mounds of sand and gravel occur at the lower ends of channels they may be due to deposition in crevasses and, perhaps, moulins, producing kames, or in subglacial (or, less commonly, englacial) tunnels forming eskers [38]. In some areas channels terminate in ground occupied by accordant-summitted kames, the kames sometimes having steep sides and quite extensive flat tops with kettles and partly closed depressions between. Such deposits suggest the existence of a lake that was largely choked by ice, deposition having occurred around and over blocks of dead ice. Strath Allan in Perthshire provides an example. Examples described in the literature are the highest aggradation level near Eaglesham, Renfrewshire [39] and the northern part of the Eddleston valley, Peeblesshire [40]. In yet other instances the deposits at the lower ends of channels form distinct terraces, which are occasionally pitted with kettles. These kame terraces, laid down mainly by streams, sometimes in very narrow marginal lakes, were long ago described by Jamieson [41]. Numerous examples exist in Scotland [42] and particularly fine examples in the Vale of Eden have been described and mapped by Trotter [43].

Although any of the types of depositional forms mentioned may be partly delta-bedded, this proves the existence of a lake only at the locality where a particular delta-bedded deposit occurs. That many such lakes were of very limited extent, often being merely narrow strips

of water situated between the ice and rising ground is often proved by ice-contact slopes.

The depositional forms referred to above are best preserved in the more recently glaciated parts of Britain, particularly in Scotland and north-west England. It was not in these areas, however, that Kendall's ice-dammed-lake hypothesis was initially developed. Furthermore, much of the early work that firmly established his ideas was carried out in the less recently glaciated parts of Britain. Here the patches of sand and gravel, at channel exits, that have been interpreted as deltas have lost the initial sharpness that would have suggested other origins. Further north in Britain, however, the various origins of such deposits can often be readily deduced. It is significant that, with the important exception of Charlesworth's major paper on the Highlands [44], there have been published for parts of Scotland very few, if any, maps of the type so often published for parts of England and Wales showing sequences of lakes whose extent has been deduced almost entirely from the altitudes of channel intakes and outlets. It is also perhaps significant to note that when Kendall turned his attention to East Lothian, an area with abundant sand and gravel deposits formed contemporaneously with the channels, he only once used the term "overflow channel", implying the existence of lakes, otherwise referring constantly to "glacial drainage channels [45]".

The doubts concerning bottom deposits. This topic will be considered only briefly, since Carruthers [46] has discussed it in considerable detail. Bottom deposits have not been reported from many of the ice-dammed lake-sites postulated in the literature. For example, Kendall was able to cite significant evidence in the North York Moors only from the Esk valley and much the greater part of the section headed "floor-deposits" in his 1902 paper is concerned with such deposits in areas other than the North York Moors [47]. Carruthers has pointed out that sediments have often been interpreted as lake-floor deposits because they are banded and has strongly maintained that such "banded dirts" have been laid down beneath glacier ice and are related to the "bottom melt". This view has not been favourably received in this country, but in both Sweden [48] and Canada [49] hypotheses involving the flowage of saturated till beneath glacier ice have been put forward to explain dead-ice moraine forms. These interpretations, made by workers of very considerable experience, suggest that some of Carruthers' ideas deserve more serious consideration than they have received so far. While not accepting all Carruthers' views, the present writer suggests that Carruthers has presented sufficient evidence to give rise to serious doubts regarding the orthodox interpretation of many supposed lake-floor deposits.

The nunatak problem. The orthodox ice-dammed-lake interpretation, which takes no account of drainage beneath, through or over the ice itself, encounters a serious difficulty where channels and related deposits show that nunataks developed during deglaciation. Certain writers have recognised this difficulty and have been forced to postulate subglacial, englacial or supraglacial drainage to overcome it. Kendall was aware of this difficulty, at least with regard to the Cheviot Hills, for the evidence of channels led him to conclude that ice-streams on the

northern and southern flanks of this hill-mass were probably confluent on the eastern slopes. Thus the Cheviot Hills were probably surrounded by ice at the time many of the channels were formed. Kendall wrote (with Muff [50]): "If the two glaciers were confluent, then the ultimate escape of the overflowing water of our lakes must have been either beneath the ice, as at the present day happens with the overflow from a chain of ice-dammed lakes on the Malaspina Glacier, or else over the top of the ice". Bremner [51] invoked a subglacial or englacial escape for a marginal stream that had existed on the slopes of a small nunatak in the Stonehaven area. Supraglacial drainage was mentioned by Raistrick [52] to carry away meltwaters that had cut channels on the flanks of a nunatak formed between the confluent Wharfedale and Airedale ice. Similarly, Tillotson [53] mentioned drainage through or over the ice from a nunatak formed between the confluent Vale of York and Nidderdale glaciers.

In the four papers cited drainage from the ice-free ground into, over or beneath the ice was suggested only to overcome the nunatak difficulty. Yet it is arguable that if such drainage occurred at one point in each of these areas it may well have occurred also at numerous other points in each area.

The emergence of nunataks is not peculiar to these four areas, however. Nunataks must have developed on numerous occasions in Britain and, whenever this occurred, the meltwaters must have escaped englacially, subglacially or supraglacially. Some of these nunataks may have been extensive tracts of higher ground, others were only small isolated hills. Examples of both types occur in Dwerryhouse's glacial map of the Alston Block [54], which is shown completely surrounded and partly overwhelmed by ice, forming three major and four minor nunataks. Series of ice-dammed lakes are shown on the flanks of the major nunataks. Dwerryhouse did not discuss the problem of the ultimate escape of the waters of his lakes. It should be added that the development of nunataks during deglaciation was probably far more frequent than is implied in the literature, since, except in a few instances, the significance of downwasting (as opposed to backwasting) of the ice has not been sufficiently appreciated.

Where nunatak escape-drainage occurred over the ice no evidence of the former stream-course is now recognisable, and this is probably also true of englacial escape in most instances. Where the waters escaped subglacially, however, eskers or channels were often formed. For example, the channels plunging steeply down the northern slope of the spur shown in Figure 1 carried waters from this spur (which was at that time part of a nunatak) beneath the ice occupying the Kingledoors valley. That these steeply inclined channels were cut subglacially, and not extraglacially, is shown by the fact that more than one occurs in each re-entrant, for, without the presence of ice in the re-entrants, only a single channel would have been cut, down the centre of each. An example of eskers formed by waters draining from a nunatak is the Thankerton "kame" and associated eskers, which were formed by drainage from the Tinto Hill nunatak in south Lanarkshire [55].

The conditions of drainage from nunataks suggested above are, of course, the same as those that operate on modern temperate glaciers, as

was made clear half-a-century and more ago by Russell [56], Tarr [57] Von Engeln [58] and others. For example, Russell, discussing the Malaspina Glacier, stated: "When a steep mountain spur projects into an ice-field the lines of drainage on each side converge and frequently unite at its extremity, forming a lake, from which the water usually escapes through a tunnel in the ice [59]". Tarr [60] gave a graphic description of the way in which the Kwik River plunged beneath the ice bordering the slopes of a nunatak to flow subglacially for upwards of five miles, descending 500 feet in this distance.

The problem of up-and-down channels. As with many other problems encountered by his hypothesis, Kendall was familiar with this one too. His views and those of others will be summarised in Part II. In the present context it need only be remarked that, when only a few examples of channels with up-and-down long profiles produced by erosion were known, it was not unreasonable to invoke special local circumstances such as ice-margin oscillations and reversal of lake overflow. It is now apparent that such channels are numerous and widespread. They have been described as occurring in the Cleveland Hills [61], Durham [62], Northumberland [63], Lincolnshire [64], the borders of the Lake District [65], North-East Scotland [66], Peeblesshire [67] and East Lothian [68]. Examples not described in the literature are known to occur in Midlothian, Peeblesshire, Perthshire and Kincardineshire. The existence of numerous channels of this type is difficult to explain in terms of Kendall's hypothesis and it will be suggested in Part II that a common cause of erosional profiles of this type was subglacial stream erosion under hydrostatic pressure, while some others were produced by reversal of drainage (but not in the manner envisaged by Kendall and other writers).

The significance of the absence of channels in some localities. If one adopts the interpretation put forward for the Cleveland Hills by Kendall as of general application, a major difficulty arises concerning the absence of channels. If, as Kendall believed, and many others have accepted, all channels were formed subaerially, it follows that on every spur down which the ice-margin retreated there should be a continuous parallel sequence of channels (Kendall's terminology), so positioned that immediately one channel was abandoned the next lower one in the sequence began to be excavated. Channels exhibiting this relationship do occur, of course, but an examination of many published maps of channels shows that very frequently indeed there are gaps, often very large gaps, in the sequences with no channels at all. In order to overcome this difficulty Kendall [69] suggested that some of these gaps might be explained by adjacent ice-dammed lakes having become confluent across the intervening spur-crest. This explanation might apply in certain instances, but, even if one accepts the ice-dammed-lake hypothesis, it can hardly be invoked repeatedly, especially as in many instances the form of the ground renders it highly improbable.

The difficulty ceases to exist once drainage into the ice is recognised as commonplace. Such drainage has been repeatedly described as occurring in Alaska and elsewhere on existing glaciers. De Boer made the point in concluding a short paper on a small ice-dammed lake in Norway that had been drained through an ice tunnel, saying: "Support

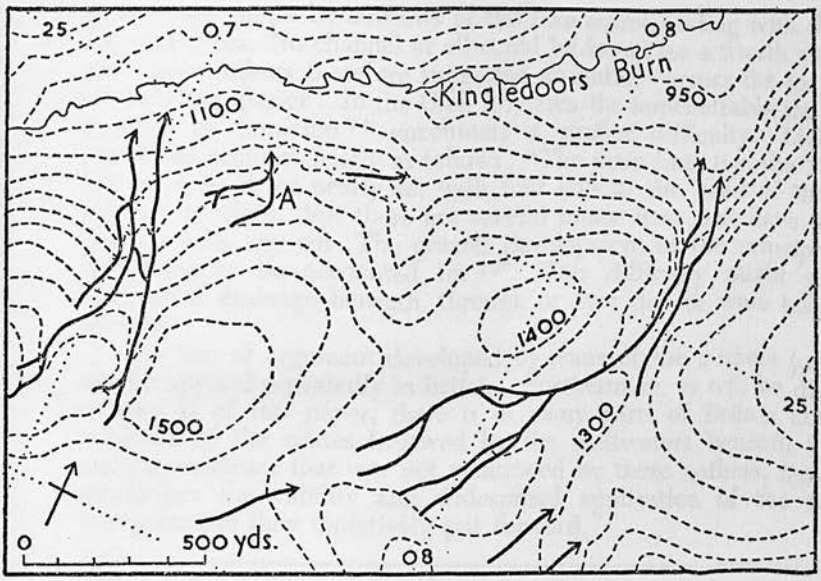


Fig. 1 Meltwater channels on part of spur between upper Tweed and Kingledoors Burn, mapped by R. J. Price. Dotted lines are kilometre National Grid. Channel A will be referred to in Part II. *Crown Copyright Resreved.*

is.... given to the suggestion that the absence, or unexpected smallness, of certain channels in an otherwise complete sequence of overflows is due to drainage into the ice front [70]". As already mentioned, Kendall was familiar with such evidence but did not apparently attach any significance to it. Charlesworth, too, has failed to apply to Britain the evidence provided by modern glaciers, for in his papers on channels he does not mention examples of drainage into or beneath the ice, although in *The Quaternary Era* he summarises this evidence and gives numerous references.

A number of workers, however, have suggested drainage into or beneath the ice to account for the absence of channels. For example, Jowett and Muff were unable to find an "overflow channel" for a lake they thought may have existed in the Hawksworth valley in the Bradford-Keighley area and therefore suggested drainage "over or through the ice-barrier [71]". Dixon and others [72], discussing the Carlisle-Silloth area, suggested subglacial drainage of a lake owing to the small size of the channels available for draining it subaerially, while Eastwood [73] suggested drainage over or through the ice owing to the absence of channels at one point in the Maryport region. Raistrick [74] suggested drainage under the ice to account for the absence of channels at one locality in Borrowdale and appears to have envisaged an important subglacial drainage system. The evidence from Rannoch Moor is particularly significant. Here several shorelines of the Glen Roy type are faintly but definitely preserved and prove the existence of a lake at several levels [75]. Yet the outflows of this lake are uncertain For example, Mathieson and Bailey found three channels that were "rather disappointing in appearance" and with obvious reluctance sug-

gested they might be outflows of the lake corresponding with three of the shorelines. No channel at all could be found for a fourth shoreline and these authors therefore suggested an outlet "across the ice of the impounding glacier". In the Glen Roy area the impenetrable ice-barrier concept of Jamieson [76] encounters a similar difficulty. Livingston posed the problem neatly as follows: "The main lines (i.e. the "roads") are on a level, or nearly so, with four cols at the head of the same number of glens. But there are several minor lines that have no connection with any col. The greater development of the principal lines will have to be accounted for [77]". This difficulty might well be resolved if drainage beneath, through or over the ice were taken into account.

The line of argument developed by some of the authors just cited can be applied repeatedly in Britain. Furthermore, as will be discussed in Part II of this paper, there is in many parts of Britain abundant evidence of the routes followed by the meltwaters beneath the ice-margin, evidence that was not mentioned by these authors, but which establishes the validity and widespread application of the type of interpretation they tentatively put forward.

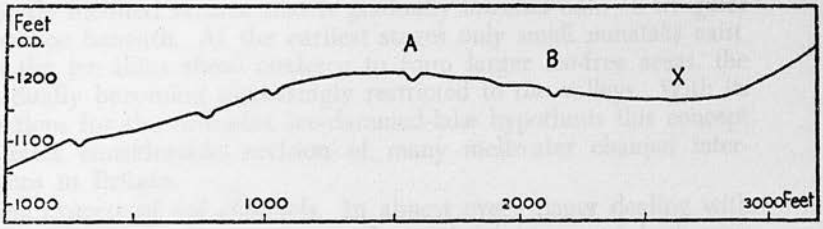


Fig. 2 Profile along crest of a spur on eastern side of Eddleston valley, Peeblesshire, showing position of five meltwater channels. A, B and X—see text.

Channels on "reverse" slopes and on minor summits on spur crests. On the interpretation put forward by Kendall channels on the slopes of spurs were formed when an ice-margin was retreating down these spur slopes. Some spur channels, however, lie in such locations that they cannot be accounted for in terms of a simple retreating ice-margin. For example, in the Lothians and Peeblesshire the writer has mapped channels running along the crests of ridges, channels that run along one side of an elongated ridge and then cross the crest-line to run along the opposite side, channels cut into the tops of minor summits on spur crests, and channels cut into the 'reverse' slopes of cols. Only channels of the two latter types, illustrated respectively by channels A and B in Figure 2, will be briefly considered here. Examples of channels on the reverse slopes of cols have been described in East Lothian [78] and the Cleveland Hills [79]. Kendall and Bailey invoked ice-margin oscillations to explain the East Lothian example, an interpretation that has been discussed elsewhere [80]. Figure 2 shows an example from Peeblesshire. It is clear that retreat of the ice-margin towards the lower ground on the left of the diagram will not explain channels A and B, while the usual explanation of ice-margin oscillations to account for anomalous channels will not suffice either. Channel A, which crosses the highest point of a minor summit on the spur crest, must have had

ice on both sides of it when it was cut. When channel B was formed the col X must have been still occupied by ice. This helps to explain the absence of a channel in the floor of the col, which on Kendall's hypothesis should contain a channel of considerable size. It may be noted that Kendall and Wroot mentioned cols without channels in the Cleveland area [81], while Best [82] also referred to an example from the same area. In neither case was any explanation offered.

The significance of downwasting. An explanation of the type put forward in the preceding paragraph for channels A and B in Figure 2 was not envisaged by Kendall, since at the time he developed his hypothesis the importance of downwasting was not recognised. In fact the significance of glacier thinning is still often insufficiently appreciated in Britain when interpreting glacial retreat features. As indicated elsewhere [83], in the considerable areas of Britian where glacier decay was by the thinning of an ice-sheet it is very unlikely that large ice-dammed lakes of the type so frequently described could ever have developed, for the supposed lake-sites would be occupied by the ice itself, its margin being closely related to the contours of the land. In order to visualise this it is suggested that the ice-surface be thought of as a gently inclined surface that is gradually lowered onto an irregular land-surface beneath. At the earliest stages only small nunataks exist, but as the ice thins these coalesce to form larger ice-free areas, the ice gradually becoming increasingly restricted to the valleys. With its implications for the orthodox ice-dammed-lake hypothesis this concept necessitates considerable revision of many meltwater channel interpretations in Britain.

Some aspects of col channels. In almost every paper dealing with meltwater channels reference is made to the existence of both very large and quite small channels. Very frequently indeed both types occur on a single spur, the very large channels often being cut into the floors of pre-existing cols and of valleys leading from them. A number of authors have pointed out that the channels on the floors of cols will naturally be deeper than other channels because they were utilised by meltwaters while the ice-margin retreated a considerable distance.

This explanation is logical if one attempts to interpret channels in terms of marginal backwasting. In terms of marginal downwasting, however, it does not apply. This is illustrated in Figure 3. On the backwasting interpretation the ice surface might occupy positions 1 to 6 during the retreat. The channel in the floor of the col would begin to be formed soon after position 2 and would continue to function until the ice had retreated to position 6. On the downwasting interpretation the ice-surface might be represented by lines A to G. In this case the col would not be ice-free until after position D. In terms of horizontal retreat the period for cutting a channel in the floor of the col would be proportional to the distance x on the backwasting interpretation but only to distance y on the other interpretation [84]. Thus on the downwasting interpretation a different explanation from the currently accepted one is required for the large channels frequently present in pre-existing cols.

This explanation is suggested by the form of many col-channels and

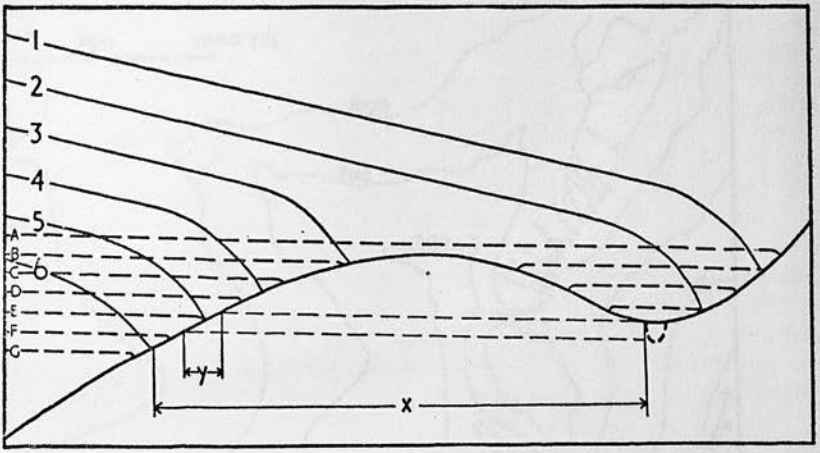


Fig. 3 Diagrammatic illustration of time available for subaerial excavation of a col channel on backwasting hypothesis (ice positions 1-6) and downwasting hypothesis (ice positions A-G). X and Y—see text.

by their relationships to other channels and depositional features. Some col-channels have up-and-down profiles that suggest a subglacial origin *(see Part II)*. Col-channels often form part of complex channel-systems that cannot be explained in terms of marginal drainage without invoking complex and improbable ice-margin oscillations. The sides of some col-channels are cut into by later channels, which are often themselves of subglacial origin. In some instances eskers lead into the heads of col-channels, thus proving that ice, not a large lake, lay immediately upstream of them, and favouring the interpretation of the col-channels as subglacial features. Several examples of this relationship are known to the writer in the Highlands and Lothians. Bremner [85] described an example in the Stonehaven area and suggested that the channel might have been partly formed subglacially.

In addition to these points the great size of many col-channels seems inconsistent with the orthodox interpretation. Repeatedly in the literature one finds adjectives such as "magnificent", "huge", "vast", "capacious", "remarkable", etc., used to describe them. Yet the adjacent channels are often relatively small. An example from the Cheviots will be discussed briefly since an explanation of the channels of a small part of this area, accompanied by an excellent map, appears in the Cheviot Hills memoir of the Geological Survey [86]. Part of this map is reproduced as Figure 4. Anderson [87] described the col-channel (X, Fig. 4) as "an impressive gorge quite 100 ft. deep". Fifteen other channels are shown on the same spur, but these channels, with one exception which is thirty feet deep, "are all small, no more than ten feet deep at most". Anderson states that the waters that cut the col-channel were provided by "an anastomosing series of shallow feeders" marked A to E in Figure 4. The difficulty encountered by this explanation is that the total cross-sectional area of these "shallow feeders" is far less than that of the col-channel, and it is difficult to believe that the relatively small streams that cut them excavated the "impressive gorge" to the north-east. In fact, the detailed evidence given by Anderson strongly

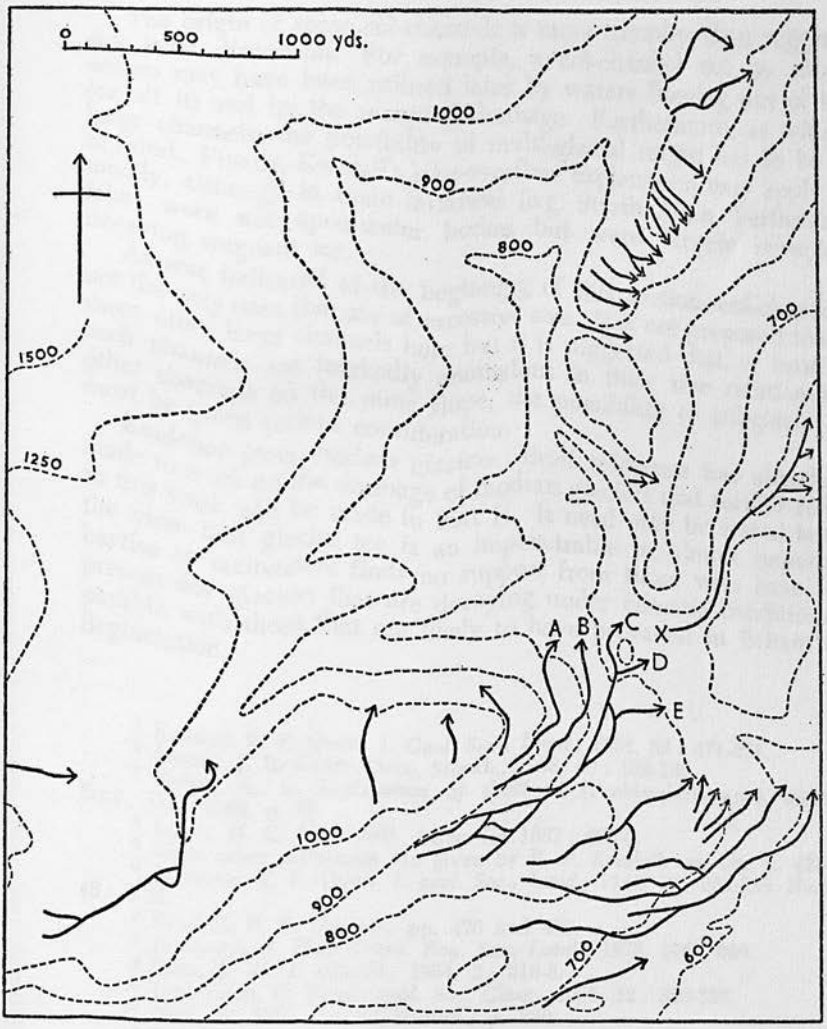


Fig. 4 Some meltwater channels on south-east slopes of Cheviot Hills. Reproduced with permission of the Controller of Her Majesty's Stationery Office from map (opp. p. 132) in *The Geology of the Cheviot Hills*, Mem. Geol. Survey, 1932.

suggests that the col-channel was excavated quite independently of the numerous small channels. These small channels appear to be of marginal, submarginal and related origins formed by relatively small volumes of water locally derived. The deep gorge is of a completely different order of size and it seems likely that it was formed by a large subglacial river that was supplied with water from a large area. The writer believes that this explanation may apply to many similar channels in other parts of Britain. Its application in part of southern Northumberland has been discussed elsewhere [88]. In that area the deep col-channels lead right down to the valley floors of the principal streams and are continued as the deep gorges occupied by these streams.

The origin of some col-channels is more complex than suggested in this brief discussion. For example, a col-channel cut by subglacial waters may have been utilised later by waters flowing out of the ice (or off it) and by the marginal drainage. Furthermore, as with other large channels, the possibility of multi-glacial origin has to be borne in mind. Finally, Kendall's lake-overflow explanation may apply occasionally, although in some instances (e.g. Strath Allan, Perthshire) the lakes were not open-water bodies but were largely occupied by decaying stagnant ice.

As was indicated at the beginning of this section, col-channels are not the only ones that are of excessive size. It is not proposed to discuss these other large channels here but it is suggested that, at least where such channels are markedly anomalous in their size relationships to other channels on the same slope, the possibility of subglacial origin must be given serious consideration.

Evidence from modern glaciers. Brief reference has already been made to work on the drainage of modern glaciers and further reference to this work will be made in Part II. It need only be stated here that the view that glacier ice is an impenetrable or almost impenetrable barrier to meltwaters finds no support from those who have studied present-day glaciers that are decaying under climatic conditions comparable with those that are likely to have prevailed in Britain during deglaciation.

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SOME ASPECTS OF GLACIAL DRAINAGE CHANNELS IN BRITAIN

Part II

J. B. Sissons

In the following pages certain types of glacial drainage channels will be described and factors affecting their origins will be considered. It is not proposed to attempt a complete classification of channels nor to discuss all types of channel, since, owing to our present inadequate knowledge, this is not advisable or practicable. Rather is it intended to show that channel origins are far more complex than has often been assumed and that origins suggested in the literature often require re-examination.

Marginal and submarginal channels. Marginal channels have been frequently described as occurring in many parts of Britain by numerous authors; submarginal channels have rarely been mentioned. In contrast, Mannerfelt has stated that in the parts of Sweden he has studied submarginal channels are the commonest type [1]. The probable reason for this apparent discrepancy in the evidence from these two areas is that in Britain a marginal origin has been assigned to virtually all channels that run along, or at a small angle with, the contours of slopes. Yet while a channel so situated shows that glacier ice stood on its downslope side, thus preventing the waters from running freely downslope, it does not show that ice was not situated on its upslope side. In other words, such a channel might well have been formed by a stream flowing along the hillside in a tunnel beneath the ice. In many instances where a marginal origin has been claimed for a channel there seems to be no reason why this explanation should not apply.

In some instances a submarginal rather than a marginal origin is suggested by the gradients of channels, for they appear too steep to have been developed along an ice-margin. Where hillside channels of varying gradients occur in juxtaposition their relationships may show that, even if the more gently inclined ones are of marginal origin, the more steeply inclined ones certainly cannot be. Such relationships are illustrated in Figure 5, which shows a part of the complex system of channels, mostly less than twenty feet deep, situated on the southern side of Strath Allan, Perthshire (See Pl. 3). Some of the highest channels here have fairly gentle gradients of about 2 per cent and may be in part marginal features. Many of the lower channels, however, as can be seen in Figure 5, have much steeper gradients and must have been cut by meltwaters flowing obliquely downslope beneath the ice. This origin is indicated not only by the channel gradients but by the sudden changes in direction of many of the channels, which show that frequently the meltwaters flowed obliquely down the hillside beneath the ice for a short distance, then turned suddenly to flow straight downslope, only to resume their oblique descent again after a few tens of yards. Such patterns seem to reflect control of the submarginal drainage in part by structures in the ice. It may be noted, too, that some of the submarginal channels are

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in places mere benches, *i.e.* one side of the channel is missing: thus the frequently made assumption that a one-sided channel is always a marginal feature is not valid [2].

Systems of submarginal and marginal channels are best developed on long, gentle or moderate slopes. This seems to be true not only in Britain (*e.g.* the systems shown in Figures 5 and 6 occur on such slopes) but also in parts of Scandinavia [3] and of Labrador [4]. Derbyshire [5] has pointed out that such channels are best developed on till-covered slopes in Labrador, and this appears to be often the case in Britain. It would seem that the steeper the slope the greater was the likelihood of meltwater streams finding their way downslope beneath the ice. In the case of marginal drainage it would be expected that, where the ice lay against a steep slope, the meltwaters would tend to erode the ice rather than the hillside. Hence marked channels are unlikely to have been formed on such slopes. Where till or other drift mantled slopes the chances of marginal drainage cutting into these slopes rather than into the ice were increased since channels could more easily become established in the unconsolidated material.

It is often difficult to prove that a channel was formed strictly marginally, rather than submarginally. One case where a marginal origin is often indicated is where channels alternate with kame terraces in aligned sequence. For example, such relationships exist in East Lothian south of Dunbar. In some instances, however, kame terraces were formed submarginally, as Holdar [6] has shown. In Figure 6 kame terrace fragments A, B, C and F and channel DE may indicate a marginal drainage resulting from the emergence at the ice-margin of the stream that built esker RSA. On the other hand, it is possible that all these features were formed beneath the ice, for it is conceivable that a depositing-stream flowing along a hillside beneath the ice might lay down its deposits in the form of a terrace. Such terraces would be the depositional counterpart of the subglacially formed one-sided channels referred to above.

Where gently inclined channels run approximately parallel with one another along slopes for considerable distances a marginal origin seems likely. That such characteristics are not in themselves proof of a marginal origin, however, can be shown by a consideration of some of the channels mapped in Figure 6. Channels GH, JK, LM and NO have gradients of about 2 per cent, are approximately parallel with one another, and are cut into a fairly gentle drift-covered slope. A marginal origin may therefore seem probable. However, between these channels cross-channels occur in places. For example, cross-channels at M, P and Q were cut by meltwaters flowing from LM to NO. The height relationships of these channels at their junctions with each other show that the cross-channels functioned successively from east to west. The last to operate, Q_2 cuts down nearly to the present floor-level of NO. This suggests that, by the time Q had been formed, NO had been excavated almost to its present depth. Thus at one and the same time meltwaters flowed in two parallel channels: LM (or part of it) and NO. A simple marginal explanation of the channels is therefore precluded. It also follows that NO was largely or entirely formed subglacially. Hence the possibility that the adjacent channels were also formed subglacially has

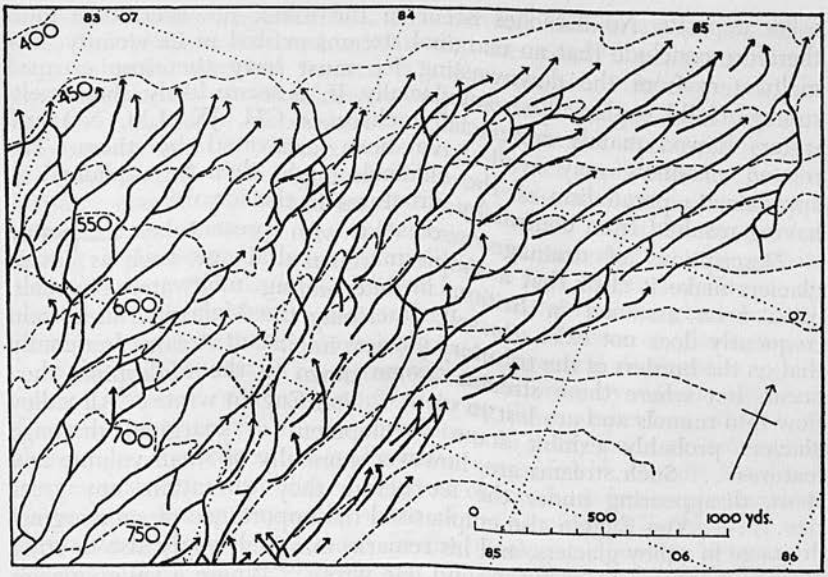


Fig. 5. Meltwater channels on the southern side of Strath Allan, Perthshire. Dotted lines are kilometre National Grid lines.

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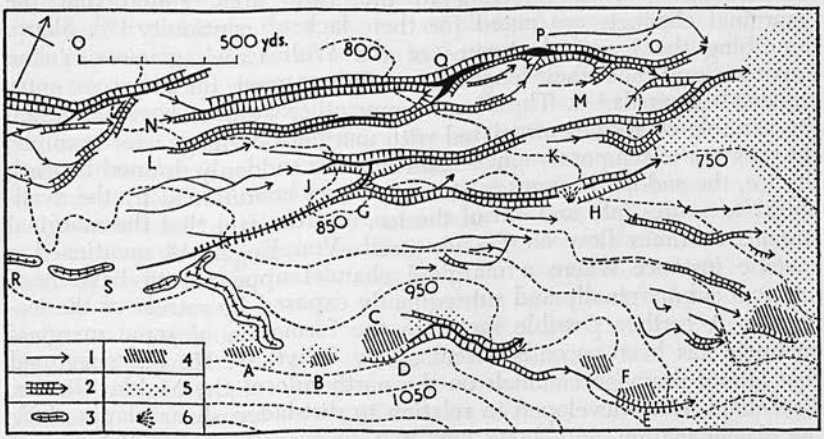


Fig. 6. Glacial drainage features at foot of northern slope of Tinto Hill, south Lanarkshire. 1. Very small meltwater channel; 2. Larger meltwater channel (no channel exceeds 25 ft. in depth); 3. Esker; 4. Kame terrace; 5. Fluvio-glacial sand and gravel, other than eskers and kame terraces; 6. Alluvial fan. Letters explained in text. *Crown Copyright reserved.*

to be considered. Evidence favouring this interpretation is provided by esker RSA. Since this esker climbs the hillside through some eighty feet and was formed subglacially, any marginal drainage that occured along the hillside in the vicinity of the esker, as the latter emerged through the downwasting ice, should have quickly breached the unconsolidated

esker deposits. No breaches occur in the esker, however. One must therefore conclude that no marginal streams existed in its vicinity. The meltwaters from the downwasting ice must have therefore escaped supraglacially, englacially or subglacially. If, as seems likely, these meltwaters flowed mainly subglacially, channels GH, JK, LM, NO and related channels may well have been excavated by them. The approximate parallelism of the channels might then be explained as having resulted from control by structures in the ice.

Descriptions of drainage associated with present-day temperate glaciers make it clear that a simple marginal drainage, such as has so often been assumed in Britain in interpreting meltwater channels, frequently does not exist. Russell, describing the Malaspina area, said that on the borders of the tributary glaciers important streams frequently occur, but where these streams "come down to the Malaspina, they flow into tunnels and are lost to view [7]". Von Engeln wrote: "All valley glaciers probably exhibit some development of marginal drainage features.... Such streams are, however, normally of small volume and short, disappearing under the ice before they can attain any great size.... [8]". Von Engeln also emphasised the importance of submarginal drainage in valley glaciers, and his remarks may well apply also to other glaciers margined by rising ground. He wrote: "Where a valley glacier has existed for a considerable period of time without experiencing any spasmodic disturbance of its mass, whether it is land-ending or tidal, nearly stagnant or in active motion, its lateral drainage will normally be submarginal [9]". Tarr, referring to the same area, stated that the "marginal channels are noted for their lack of continuity [10]". Sharp, describing the marginal streams of the Wolf Creek glaciers, Yukon Territory, said that "their length is usually not great, for they soon enter englacial channels [11]". These and many other authors have described the variable conditions associated with marginal drainage: for example, the creation of temporary small lakes that are suddenly drained beneath the ice, the sudden changes in stream courses conditioned by the availability of routes into and out of the ice, and the fact that the marginal streams at times flow on the ice itself. Von Engeln [12] mentioned a specific instance where a marginal channel appeared to have been initiated submarginally and subsequently exposed by retreat of the ice-margin. A further possible factor in the formation of some marginal channels has been recently discussed by Schytt [13]. He has suggested that lateral drainage channels on the north side of the Moltke Glacier, Greenland, were developed in relation to dirt-laden shear planes along the glacier margin, and shows how two, or even three, parallel streams may flow along and just within the margin of a glacier at the same time.

Another aspect of marginal channels that appears not to have been investigated in Britain is their possible relations to periods of active ice movement. Mannerfelt [14] emphasised the importance of submarginal and subglacial drainage during stagnation. Conversely, one might expect the absence or poor development of such drainage (or, at least, of evidence of it) and a strong development of marginal channels to characterise former periods of active ice movement. In this context Von Engeln's observations on the glaciers of the Yakutat Bay region are significant. He stated: ".... under certain conditions marginal streams

of great volume occupy such channels between the ice and the mountain side. The relations observed indicate that the typical occurrence of these large-volumed marginal streams is a phenomenon following upon the rapid advance of valley glaciers; and that they do not develop where normal conditions, dependant on long-continued, uniform climatic influences prevail [15]". If this evidence has wider application (and it has to be noted that Flint [16] found different evidence, but under different climatic conditions, in north-east Greenland) it might be possible to deduce from marginal channels periods of renewed glacial activity during deglaciation in areas such as Britain. Von Engeln's evidence also seems to carry with it the significant implication that marginal drainage should have been more common during the advance of glacier ice to its maximum extent than during periods of general deglaciation.

Subglacial chutes. This term was introduced by Mannerfelt [17] to apply to channels cut by streams plunging down slopes beneath the ice. Channels of this type are very numerous in many parts of Britain, but their origin has only occasionally been recognised [18]. Usually they have not been shown on published maps of channels. A few maps, however, show them clearly, especially some of the Geological Survey One-Inch sheets and some of the maps in the Geological Survey Memoirs [19], although their origin appears not to have been explained in these memoirs.

A simple type of subglacial chute occurs as the continuation of a marginal or submarginal channel (e.g. A Fig. 1, AB Fig. 9). One may follow the marginal or submarginal channel along the side of a hill to find that it suddenly turns sharply to plunge steeply down the hillside as a chute. In the case of a marginal channel the point at which the channel suddenly changes direction is the point at which the waters flowing along the ice-margin turned to plunge beneath the ice. In some instances the route followed by the meltwaters before they found a path down the slope is indicated by the continuation of the marginal (or submarginal) channel beyond the upper end of the chute.

Mannerfelt [20] related the formation of subglacial drainage channels and the supposedly regular spacing of marginal channels to the annual melting rhythm of the ice. He suggested that marginal drainage operated mainly in the early part of each melting season and that subglacial drainage became dominant in the later part of each melting season with the opening of the subglacial tunnels, an interpretation that seems to receive some support from Tarr's remarks on the Lucia Glacier, Alaska [21]. This annual concept has been criticised, however, by Holdar, who found an average spacing of over 20 metres between marginal channels in the part of the Swedish mountain chain he studied, a spacing far too large to be related to the annual ablation [22]. Schytt [23], as noted above, has suggested that the spacing of the marginal channels of the Moltke Glacier, Greenland, is controlled by dirt-laden shear planes. Presumably as a result of these criticisms, Mannerfelt [24] has recently expressed doubts concerning his annual interpretation of certain Swedish meltwater channels.

In the areas studied by the writer in Britain channel relationships often seem to imply that subglacial routes were available from the very beginning of the melting season. Figure 7 provides a simple illustration.

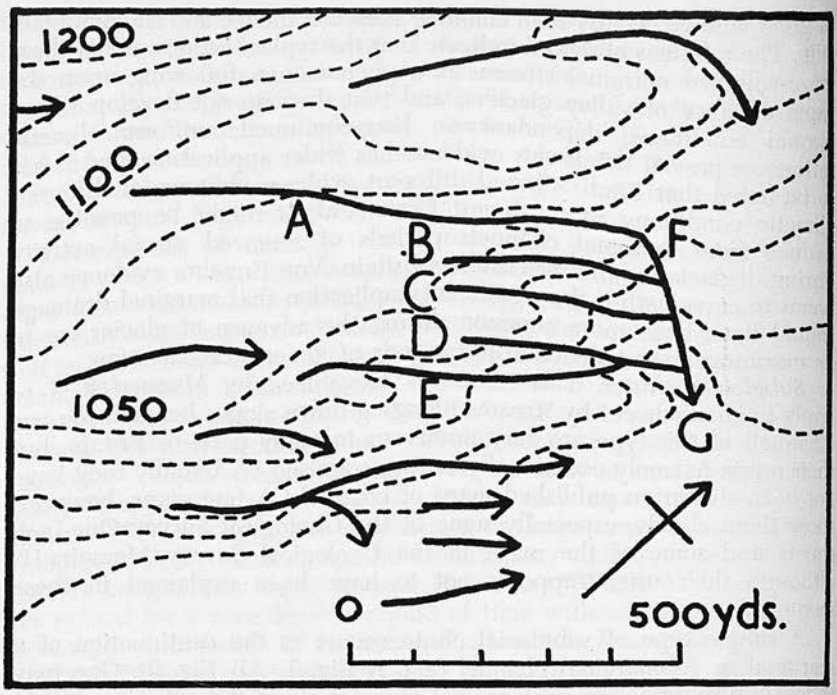


Fig. 7. Some meltwater channels to west of Eddleston valley, Peeblesshire, about 2 miles WNW of Waterheads. Letters explained in text.

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Here the channels A to E all end in subglacial chute FG and there is no trace of meltwater erosion beyond the chute. Whether or not the five channels A to E are related to the annual ablation of the ice, the fact that E lies some 50 feet lower than A indicates that more than one melting season was involved in their formation. The chute FG carried their waters throughout this period and must have functioned therefore from the beginning of each melting season.

Many subglacial chutes are unrelated to preglacial (or interglacial) stream courses (although they have quite frequently provided ready-made valleys for small misfit postglacial streams). Other chutes, however, developed along pre-existing valley floors. In such instances one may sometimes follow marginal or submarginal channels along a hillside to find that they terminate along the line of a stream descending the slope. The stream itself may be found to occupy a steep-sided trench, whose origin as a subglacial channel can be deduced from its continuity with the marginal or submarginal channels. One may presume that the development of subglacial channels along such valley floors was facilitated by the relief and by the relatively warm waters of these streams as they flowed down the ice-free parts of their valleys towards the ice-margin. Liestøl [25] mentioned the importance of streams flowing down valley-sides in keeping open a large subglacial drainage system and measured tempartures of 3 to 12° C. (c. 37-54° F.) in August,

1949, in brooks flowing down into a glacier in Norway. Von Engeln [26] also stressed the importance of the warm land-drainage and found its temperature to be around 44° F. (6.7° C.) in late July and early August in the valley of the Galiano Glacier, Alaska. It is interesting to note, too, that von Engeln described, on the north side of the Hidden Glacier, Alaska, drainage apparently identical with that described above. He wrote: "On the lower side of a rock spur there was commonly an emergence of a marginal stream from an ice cave at the shelving ice edge, and this marginal stream then continued along the ice and rock contact until it became confluent with the major stream of the valley embayment (i.e. side valley) drainage. At the point of confluence both the valley drainage and the marginal drainage disappeared under the ice through a low ice cave [27]".

In addition to occurring singly, as described above, subglacial chutes frequently occur in groups, sometimes as parts of quite complex channel systems. A common location for groups of chutes is where the ice during its downwasting extended as tongues up valleys owing to the emergence above the ice surface of the intervening spurs. In such situations the marginal and submarginal drainage, instead of following the ice-margin into each re-entrant, frequently plunged beneath the ice and cut subglacial chutes. Presumably the development of chutes in such locations was facilitated by the fact that the ice tongues extending up the valleys would tend to stagnate more readily than the main ice-mass, particularly where, as was frequently the case, such valleys lay transversely to the main direction of ice movement. An example of such chutes, cut into one flank of a broad embayment in Midlothian, is shown in Figure 8. Although some of these channels are at present occupied by streams they are not of postglacial origin. This is indicated by their relations to each other and by the fact that some of them are continuous with streamless channels that run along the generalised contours of the embayment. Some of the chutes lead into the valley of the South Esk, which stream occupies a steep-sided trench (now partly infilled in places by later-formed river terraces and alluvium) that is itself probably in part a subglacial channel that carried the waters of numerous tributary channels beneath the ice.

In some localities chutes occur in large numbers with only a limited development of marginal and submarginal drainage. An example is shown in Figure 9. That these valleys are not postglacial gullies is shown by (i) their relation in a few instances to marginal and submarginal channels, (ii) the oblique descent of some of the channels, (iii) the abrupt beginnings and endings of some channels, (iv) the fact that many of the channels stop part way down the steep slope into which they have been cut, (v) the absence from the lower ends of the channels of alluvial fans, (vi) the absence of streams from the channels, including some that exceed 50 feet in depth, there being only one exception to this (C, Figure 9, which is utilised by a stream significantly named Dry Burn).

Numerous short subglacial chutes are shown in Figure 5 and frequently occur as continuations of the short, steeply inclined, downslope segments of the submarginal channels. They represent the routes by which these channels were abandoned, this abandonment usually

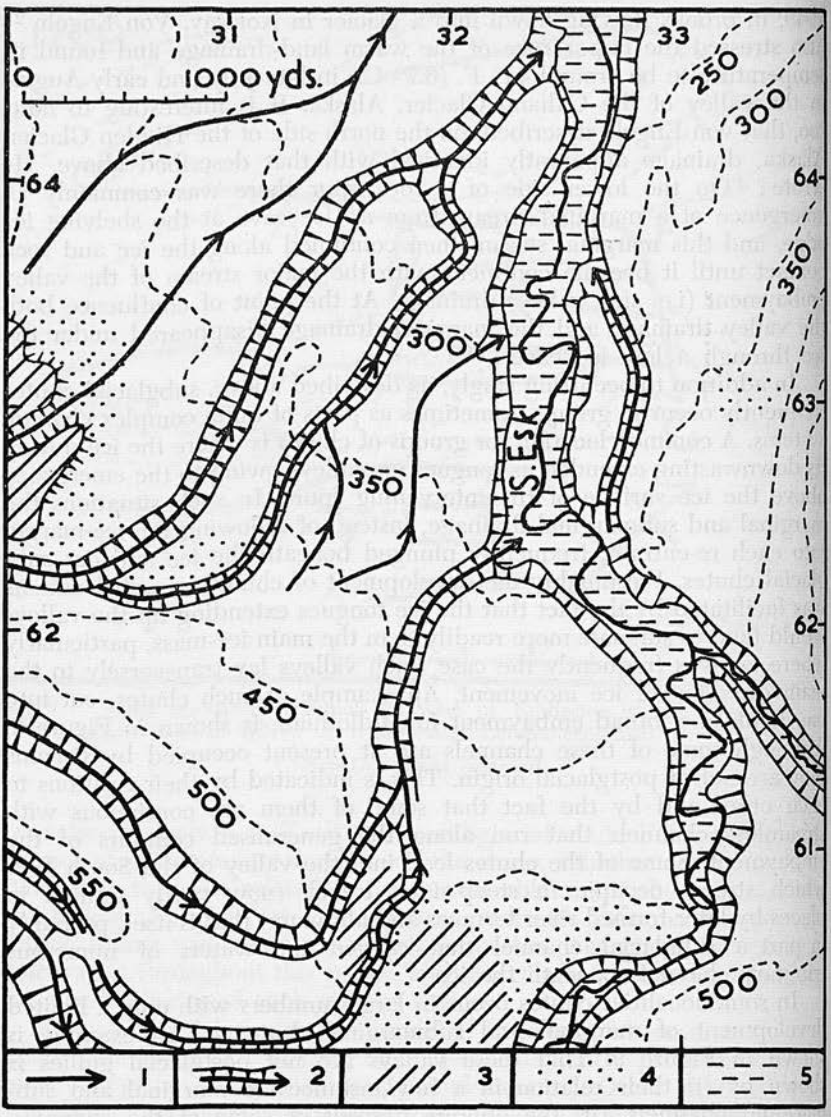


Fig. 8. Glacial drainage features in eastern Midlothian, near Newtongrange. 1. Small meltwater channel; 2. Larger meltwater channel; 3. Kame terrace; 4. Fluvio-glacial sand and gravel other than kame terraces; 5. River terraces and alluvium. National Grid kilometre lines indicated round margin of map.

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having occurred successively from east to west. One of the largest subglacial chutes occurs in the extreme east and carried the waters of five channels that join it. In the north-central part of the map complex relationships between the submarginal channels and subglacial chutes are shown. In general the map illustrates the attempts of the melt-

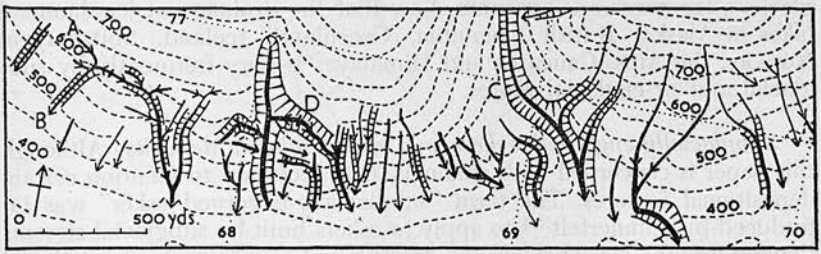


Fig. 9. Meltwater channels (mostly subglacial chutes) some three miles north-east of Fettercairn, Kincardineshire. Width of arrows proportional to width of channel floors. Larger channels shown by hachures. Letters explained in text. Dotted lines are kilometre National Grid lines. *Crown Copyright reserved.*

waters to find routes straight down the ground slope beneath the ice and the partial success of the ice in preventing this, resulting in the two dominant directions of channels shown on the map. It will be noted, however, that some channels follow intermediate directions and it is difficult to decide whether they should be classed as subglacial chutes or submarginal channels. This serves to illustrate the fact that these two types of channels grade into each other.

The effect of the relief and the ice in controlling the development of streams in the submarginal zone is also illustrated in Figure 10. The origins of most of these channels are very similar to those of the channels shown in Figure 5, even though the pattern appears at first quite different. The channels shown in Figure 10 are cut in bedrock and are mostly less than twenty feet deep. The pattern can be resolved into submarginal channels descending obliquely, but often quite rapidly, down the slopes of the embayment, and subglacial chutes going straight down the slopes, the whole system being essentially convergent owing to the pre-existing ground slopes.

One other aspect of subglacial chutes that should be mentioned is their possible significance in the draining of ice-dammed lakes. This aspect of chutes has apparently not yet been considered in Britain, although A. G. Macgregor [28] mentioned the existence of numerous dry gorges (apparently subglacial chutes) associated with the draining of an ice-dammed lake in Ayrshire, and Bremner [29] referred to an ice-dammed lake in the Findhorn valley that was drained both along the ice edge and by subglacial channels. In Sweden the idea of subglacial drainage of lakes has been developed by Gillberg [30] in his paper on the deglaciation of a part of the Swedish highlands. He has deduced the former existence of numerous marginal lakes along the mountain slopes during the downwasting of the ice-sheet. These were not large lakes, however, such as have been frequently described as formerly existing in Britain, but were mostly narrow marginal strips of water. Many of the lakes were developed on the flanks of emerging nunataks and must therefore have been drained via the ice itself. Gillberg states that the outlets of some of the lakes were subaerial but that most were subglacial, sometimes by inclined submarginal routes, at other times by subglacial chutes, the escaping waters subsequently becoming englacial. Such an interpretation is consistent with the evidence from modern

glaciers, for numerous accounts show that the drainage of ice-dammed lakes in Alaska, British Columbia, Greenland, Iceland, Spitsbergen, Norway, the Alps, Caucasus and Himalayas is very frequently by subglacial and/or englacial routes.

Subglacially-engorged eskers and some related deposits. Although this paper is concerned with channels, it is necessary to mention certain depositional features. The term "subglacially-engorged esker" was introduced by Mannerfelt [31] to apply to eskers built by subglacial streams flowing into the ice. They are the depositional counterparts of subglacial chutes. In Britain eskers have been usually explained as formed only by streams flowing out of the ice, but occasionally the origin put forward by Mannerfelt has been suggested. For example, this interpretation seems to have been favoured for some esker-like features near Oswestry by the Geological Survey officers who mapped that region [32] and by Bailey [33] for two eskers in Ayrshire. Bremner [34] mentioned a "kame" in the Aberdeen area that might have been formed in a subglacial tunnel by a stream flowing into the ice, while Pallister [35] described mounds on the lower slopes and bottoms of valleys in part of Dumfriesshire and mentioned the possibility of their having been formed by land streams washing gravels down crevasses into stagnant ice. As long ago as 1898 Garwood provided a very interesting description of how, in Spitsbergen, the draining of a small ice-dammed lake through the ice had resulted in a gravelly deposit along the floor of an ice tunnel at right-angles to the direction of the glacier flow and suggested that "we have here an instance of one possible origin of the gravelly ridges known in our country as kames and eskers [36]".

In parts of south-east Scotland, at least, subglacially-engorged eskers are fairly common, and the writer has noted some fifty or more. They are usually slightly sinuous ridges, often no more than twenty feet or so high, descending slopes and sometimes extending onto flatter ground at the base of the slopes. Many of them are no more than a few hundred yards long, although a few are much longer. The till mantle that Mannerfelt described as occurring on Swedish examples is not common, although angular and subangular cobbles and boulders sometimes occur on the eskers. Some of the eskers have no obvious relation to other glacial drainage features. A few of the eskers mapped continue meltwater channels and were clearly deposited by the streams that cut the channels. Some others lead away from large accumulations of sand and gravel, sometimes beginning at approximately the same level as the surface of an adjacent kame terrace.

While the features mentioned in the preceding paragraph would probably be accepted by most geomorphologists as eskers on the basis of their form, there exist far more numerous moundy deposits of sand and gravel that would usually be described as kames, yet which often appear to have a fairly similar origin to the subglacially-engorged eskers. Frequently such kames are closely associated with meltwater channels. For example, along the lower slopes of the western side of the north-south Eddleston valley, Peeblesshire, a series of kames extends for some four to five miles and varies in width between half a mile and nearly a mile. On the valley side immediately above the kames is a complex

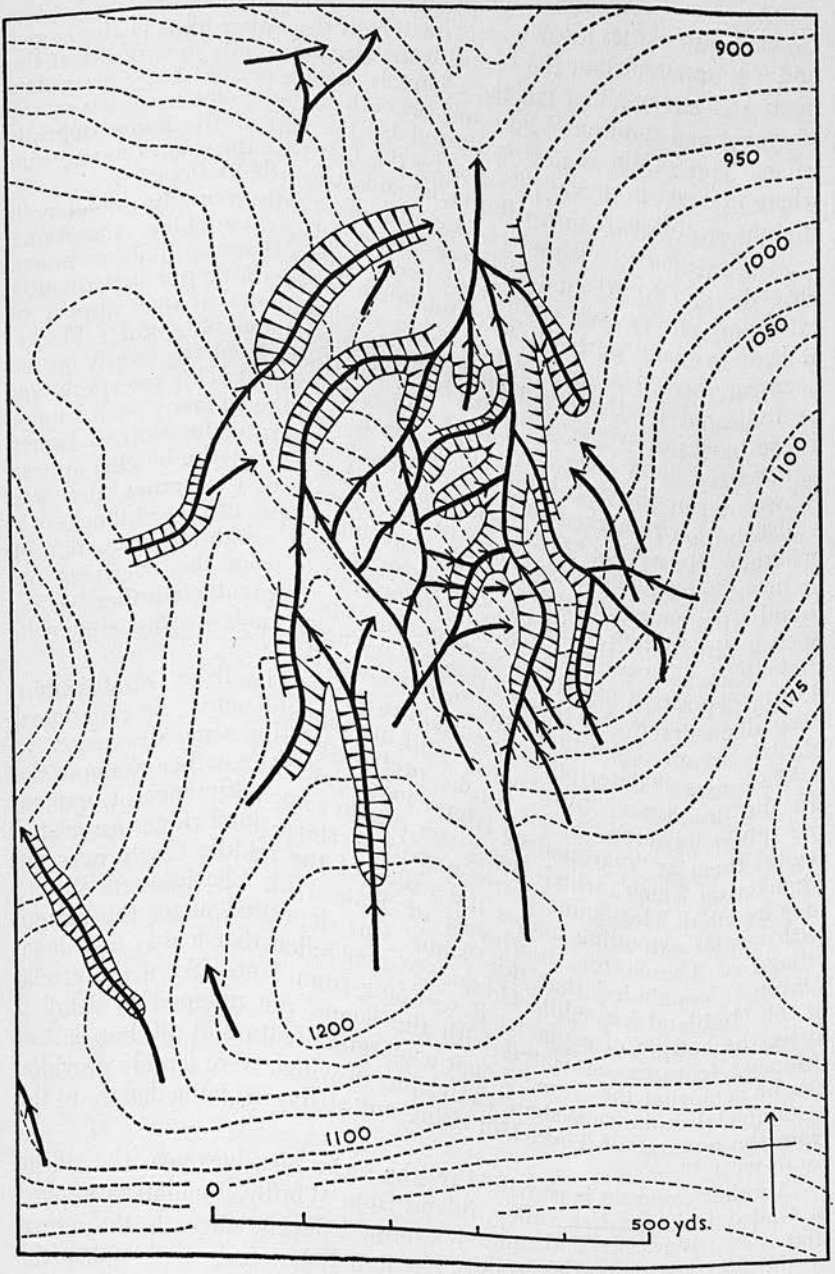


Fig. 10. Meltwater channels on part of an east-west spur on eastern side of Eddleston valley, Peeblesshire. Hachures indicate channels deeper than about 10 feet.

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system of marginal, submarginal and subglacial drainage channels. The channel system dies away approximately at the upper limit of the kames and it is apparent that the kames were formed mainly as a result of the meltwater streams that cut the channels having carried large quantities of gravel and sand onto, into and beneath the stagnant ice lying in the valley. This origin is also indicated by the dip of the kame deposits where exposed in gravel pits and by the fact that the kames mostly run straight or obliquely down the valley side towards its floor [37].

The existence of kames (or spreads of slightly moundy gravel with little definite form) on the lower slopes of valley sides, sometimes extending onto or even completely across valley floors, is quite common, at least in many parts of Scotland. Such kames have not infrequently been interpreted as end moraines, an explanation that may appear to be indicated by the occurrence on some of them of angular blocks. These blocks, however, were probably let down from the overlying ice as it wasted away after the kames had been formed. This interpretation is favoured by the distribution of blocks in relation to many such kames, a distribution remarked on by J. Geikie in referring to Scottish kames generally. He wrote : "It is remarkable.. that if they (the blocks) appear to be of infrequent occurrence in the interior of the kames, they are found often enough dotted over the tops and slopes of these hillocks [38]". Such a distribution of blocks is consistent with subglacial formation of these kames, occasional blocks having fallen from the overlying ice during deposition of the kame materials, and a greater number having been deposited when the overlying ice finally melted away after subglacial stream deposition had locally ceased.

Pallister's [39] description of certain ridges and knolls in Dumfriesshire fits the description given above and, as already noted, he mentioned the origin advanced here. Yet he preferred, after some discussion, to regard them as "marginal or lateral moraines". As another example the Thankerton 'kame', which has given rise to some difference of opinion, may be cited. This 'kame' forms a series of steep-sided ridges associated with kettles extending across the valley of the middle Clyde near the village of Thankerton, north-east of Tinto Hill. Charlesworth [40] and Linton [41] concluded that this 'kame' was deposited along the margin of the Highland ice, while Gregory [42] concluded that it was laid down along the margin of a glacier descending from Tinto. Yet if the glacial drainage features associated with the 'kame' are mapped in detail it can be seen that the 'kame' is not a marginal feature at all, but part of a subglacial drainage system, the waters of which were largely provided from the nunatak of Tinto Hill rising out of the ice immediately to the south-west [43].

Another example is provided by the series of ridges near the village of Glaisdale in the Esk valley, inland from Whitby. Kendall [44] implied that these ridges are end moraines formed during stages in the retreat of the ice down the Esk valley. He stated, however, that, unlike the Lealholm moraine further west, these ridges are composed of sand and gravel. The ridges are shown in Figure 11, as also are the well-marked terraces of sand and gravel found in the same area. The terrace deposits extending down the valley of Stonegate Beck between altitudes of 550 and 600 feet were laid down by the meltwaters that had crossed



Plate 3 *Royal Air Force, Crown Copyright Reserved.*
Part of a system of meltwater channels situated on southern side of Strath Allan, Perthshire (see Fig. 5). Channels mostly less than 20 feet deep and some are benches. South is at top of photo; scale 1 : 10,000 approx.; Nat. Grid Ref. of centre point 27/839062.

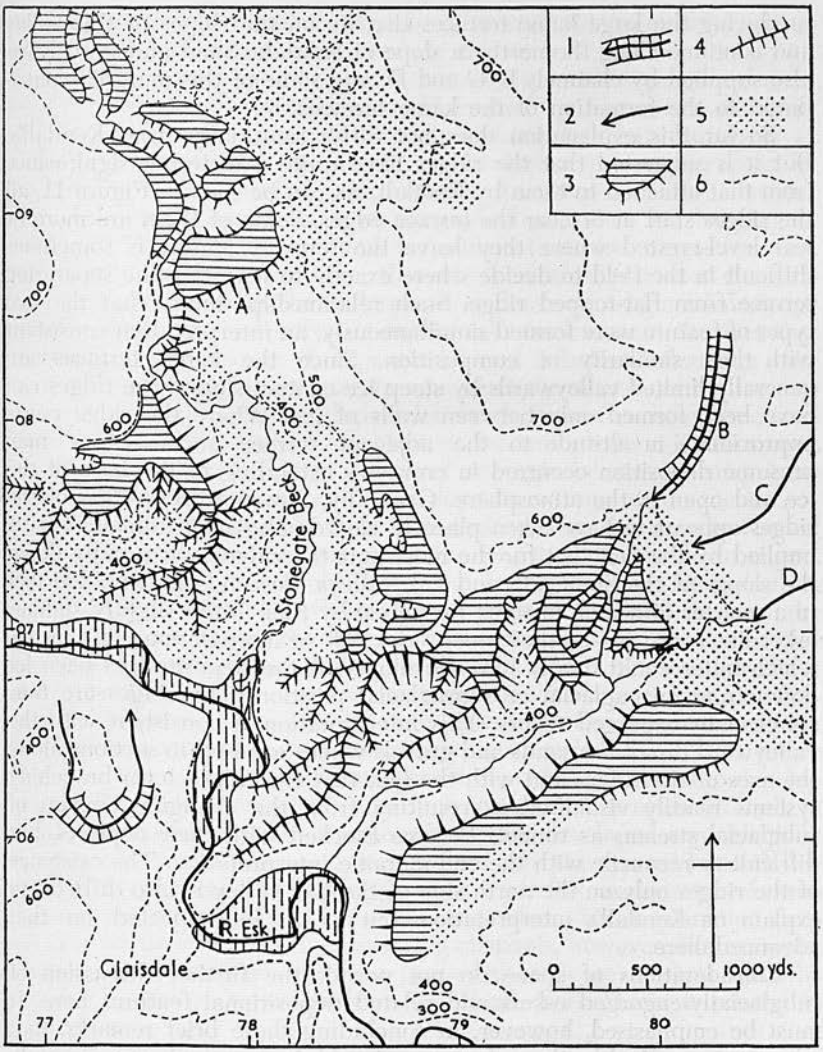


Fig. 11. Glacial drainage features in the Esk and Stonegate valleys near Glaisdale, north-east Yorkshire. 1. Large meltwater channel; 2. Small meltwater channel; 3. Fluvio-glacial terrace; 4. Subglacially-engorged esker; 5. Fluvio-glacial sand and gravel not mapped as terraces or eskers; 6. Contour. Letters explained in text. National Grid kilometre lines indicated round margin of map.

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the northern watershed of the Esk valley by the deep Tranmire Slack and related meltwater channels. As far down the Stonegate valley as point A, Figure 11, where a major ice-contact slope runs across the valley, the terrace deposits appear originally to have extended completely across the valley. Below this point deposition took place on either side of the mass of dead ice lying in the Stonegate valley,

producing the large kame terraces that lie on the slopes of this valley and continue along the northern slope of Eskdale. Sand and gravel was also supplied by channels B, C and D, and some of this material contributed to the formation of the kame terraces.

So far this explanation does not differ materially from Kendall's. But it is suggested that the ridges have quite a different significance from that attached to them by Kendall. As can be seen in Figure 11, all the ridges start at or near the terrace edges. Some of them are more or less level-crested where they leave the terraces and it is sometimes difficult in the field to decide where exactly to map the line separating terrace from flat-topped ridge. Such relationships imply that the two types of feature were formed simultaneously, an interpretation consistent with their similarity of composition. Since the kame terraces are generally limited valleywards by steep ice-contact slopes, the ridges can have been formed only between walls of ice. Where the ridge crests approximate in altitude to the adjacent terrace surfaces one may presume deposition occurred in crevasses extending to the base of the ice and open to the atmosphere. Generally, however, deposition of the ridges appears to have taken place in closed subglacial tunnels. This is implied by the fact that for the most part the ridges run rapidly down the slopes of the Stonegate and Esk valleys, descending some 200 feet in a quarter of a mile in some instances, by their often slightly sinuous esker-like form, and by their steep sides (20° or more in many instances), which latter could hardly have developed if the deposits had been let down from an englacial or supraglacial position. The ridges are thus subglacially-engorged eskers. This interpretation is consistent with the valleyward dip of the sands and gravels where exposed in sections along the axes of the ridges and with the fact that the ridges form branching systems readily visualised as resulting from the changing courses of subglacial streams as tunnels became blocked with their deposits, but difficult to reconcile with the end-moraine interpretation. The existence of the ridges only on the north side of the Esk valley is also difficult to explain on Kendall's interpretation but is to be expected on that advanced here.

Considerations of space do not permit the further discussion of subglacially-engorged eskers and related depositional features here. It must be emphasised, however, in concluding these brief remarks, that wherever practicable these features should be mapped as accurately as the channels. In favourable localities such mapping may reveal the pattern of glacier stream-flow in considerable detail and may provide the missing links between channels that have been frequently provided in the past by postulating lakes.

In-and-out channels. These channels formed one of the four types recognised by Kendall and he described them as crescentic valleys excavated in the face of a hill. He stated that they were formed "by water flowing round a projecting lobe of ice [45]", an interpretation accepted by Smith [46] and Derbyshire [47]. It is difficult, however, to envisage the existence of such lobes of ice, for the ice-margin normally conforms approximately with the contours of a slope. The fact that in-and-out channels resemble meanders suggests that they have been formed, at least in some instances, by meandering supraglacial or,

perhaps, englacial streams that cut down through the ice into the ground beneath, an explanation mentioned by Common [48].

The meandering habit of glacial streams, especially supraglacial ones, has been frequently mentioned or illustrated [49]. The origin favoured here for in-and-out channels has been stated by Sharp to apply to such channels situated on the side of a valley at present occupied by a glacier in Yukon Territory, for, referring to these channels he wrote: they "are clearly cutoff loops of entrenched meanders left hanging at both ends [50]". It should perhaps be added that the suggestion that supraglacial drainage may have, through its incision, produced channels is not inconsistent with the widespread evidence of subglacial drainage already discussed. Not only has one to take into account the fact that the relative importance of supraglacial and subglacial drainage may have varied during deglaciation in relation to the activity of the ice, but also that supraglacial and subglacial streams have often been described from immediately adjacent localities on modern glaciers. Furthermore, in some instances the development of supraglacial streams may have been favoured by a cover of fluvial deposits or moraine on the ice.

Other superimposed channels. If in-and-out channels may be formed by superimposition of streams from the ice, it is reasonable to expect that other channels too would show meandering courses. This is, in fact, often the case (*e.g.* Midlothian, East Lothian, Cleveland, Rossendale, etc.). In some instances the form of meandering channels indicates that the meanders developed during their incision into the rock (or drift) and it is unnecessary to postulate superimposition from the ice. Frequently, however, meanders are intrenched, not ingrown. Like in-and-out channels, such meandering channels are often found on slopes where the form of the ground would not permit meandering streams to develop marginally or extra-marginally. In such instances it is often difficult to avoid the conclusion that the meanders were developed supraglacially (or, perhaps, englacially) and then incised into the ground beneath the ice. This was mentioned by Hoppe [51] as a possible explanation of some Swedish channels that had occasional meanders. This explanation does not necessarily apply to all meandering channels, however. For example, Harrison and Thackeray [52] have presented evidence to show that the winding course of Newtondale in the North York Moors is related to joint planes.

Another way in which superimposed (not necessarily meandering) channels might develop was suggested by Lamplugh in the discussion of a paper on Spitsbergen by Garwood in 1899. Lamplugh asked "whether the Author had observed any instance where the streams flowing on the ice-sheets had intersected emerging ridges of solid rock [53]". In reply Garwood said he was unable to cite examples but agreed that meltwater channels might be formed in this way. This interpretation appears not to have been applied to channels cut across spurs in Britain until recently developed by Price [54] as a possible explanation of some of the channels of the upper Tweed valley. These channels cut indiscriminately across spur crests and then descend part way down the slope beyond. Price suggests that as the ice thinned the supraglacial and englacial streams became superimposed on the spur crests beneath. This drainage was thus disrupted and became subglacial (as

indicated by the channels) on the further side of each spur.

Open ice-walled channels. The term "ice-walled channel" was used by Gravenor and Kupsch [55] to include channels formed in open trenches in the ice as well as in closed tunnels. The former will be referred to here as "open ice-walled channels". So far as the writer is aware the formation of channels in open trenches has not been suggested in Britain, although the formation of certain 'kames' in such situations has [56]. Such an origin for some kames is consistent with Russell's evidence from the Malaspina Glacier. He described [57] how the Kame River emerged from a tunnel in the ice some three miles back from the border of the glacier to flow in an open ice-walled channel, the first half-mile of which was a narrow canyon with walls of dirty ice fifty or more feet high. The river was depositing large quantities of sand and gravel, partly on ice, and one may presume that, with favourable conditions, a linear kame-belt at right angles to the former glacier margin would survive after the ice wasted away. Another example of an open ice-walled channel was described by Roberts [58] in Iceland, where he found a shallow sixty-foot wide river flowing for several hundred yards between precipitous walls of dirty ice eighty to a hundred feet high.

Open ice-walled streams might develop either through collapse of the roofs of subglacial tunnels or through supraglacial streams cutting down through the ice (as was suggested above for the streams that cut some meandering channels). Russell's evidence from the Malaspina area indicates the formation of open ice-walled channels by collapse of subglacial tunnels, and this origin is envisaged by Gravenor and Kupsch [59] for some of the channels developed in association with a former ice-sheet in western Canada. The present writer has also suggested this origin for some of the huge channels formed by Great Lakes waters flowing through the Syracuse area in central New York [60]. The channels described by Gravenor and Kupsch are considered by these authors to have been cut partly in open trenches and partly in closed tunnels and in places they are related to esker systems. These channels form a complex pattern of parallel and intersecting elements. In one region two sets of channels occur at right angles, but the two sets were not everywhere used at the same time. These channels in western Canada in many respects resemble the Rinnentäler of Germany and the tunnel-dale of Denmark, where formation in closed tunnels is favoured, although Woldstedt [61] has suggested that erosion by the ice itself may have initiated the formation of some channels. The possibility that some channels in Britain have been formed by open ice-walled streams has therefore to be borne in mind.

Up-and-down channels. Many glacial drainage channels have up-and-down long profiles. Frequently a profile of this type has been obviously produced by accumulation of material on the channel floor. One of the commonest causes of this is the formation of alluvial fans by streams entering channels from their sides. Such fans were described from East Lothian by Kendall and Bailey [62] and referred to as "corroms". Some other channels possess up-and-down profiles owing to the accumulation of peat and/or mass-movement from their sides. There remains, however, a very significant number of other channels whose up-and-down profiles are due only partly or not at all to

deposition. At least six explanations have been put forward in the literature to account for such channels. These are as follows: (i) "unequal volumes of meltwater flowing in opposite directions more or less contemporaneously [63]"; (ii) postglacial stream erosion [64]; (iii) erosion of a channel whose upstream part is in drift and downstream part in rock, resulting in an uphill long profile in the more easily eroded drift [65]; (iv) erosion of a channel across a col previously lowered by a "preglacial" stream working backwards by headward erosion, resulting in the floor of the "preglacial" valley forming the uphill part of the profile [66]; (v) reversal of meltwater flow [67]; (vi) subglacial erosion by water flowing under hydrostatic pressure [68].

The first four explanations need be considered only briefly. The first one seems rather improbable and can hardly account for channels that are continuous well-marked features throughout their lengths. The second seems equally inapplicable in most instances, for postglacial streams are often absent or of small dimensions in the channels under consideration. Certainly in the example cited by Eckford [69] (Windy Gowl channel, Peeblesshire, Nat. Grid Ref. 36/1454) the half-mile uphill part of the channel could not have been excavated by the present tiny stream that occupies it: this stream has, in fact, cut a small notch a few tens of yards long into the floor of the uphill part of the channel. The third explanation applies to a particular case and cannot account for the numerous up-and-down channels where different geological conditions obtain. The fourth, although put forward by Twidale [70] as an explanation of up-and-down channel profiles, and hence listed above, is not really such. The uphill slope is simply the floor of a preglacial valley and the floor of the channel proper slopes in the same direction throughout its length. Such channels were long ago described by Kendall and Bailey [71].

The fifth explanation—reversal of meltwater flow through channels—has been by far the most commonly suggested. It was put forward originally by Kendall in 1902 [72]. In the following year he repeated the explanation, with some elaboration, apparently as a result of criticisms made by members of the Yorkshire Geological Society [73]. Subsequently, however, he appears to have become uncertain about this interpretation, for Elgee quotes a letter from Kendall on the problem of up-and-down channels as follows: "Experience gained in the last two years has made me suspicious of cases of apparent flow in opposite directions through the same channel. It is easier to assume that a transverse watershed or 'corrom' has been produced by inflow of a lateral stream, or by some similar means [74]". Much more recently Peel has discussed the reversal of drainage idea and has said that it suffers from the disadvantage "that it requires special conditions unlikely often to be fulfilled [75]".

The reversal of flow concept is certainly difficult to apply if one attempts to explain channels as lake overflows associated with an impenetrable ice-barrier, but these difficulties may disappear when the importance of drainage into the ice is recognised. For example, in the small area shown in Figure 9 the large number of subglacial chutes points to abundant drainage down the hillside beneath the ice. The mutual relations of some of the chutes show that they did not all

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Up-and-down channels. Many glacial drainage channels have up-and-down long profiles. Frequently a profile of this type has been obviously produced by accumulation of material on the channel floor. One of the commonest causes of this is the formation of alluvial fans by streams entering channels from their sides. Such fans were described from East Lothian by Kendall and Bailey [62] and referred to as "corroms". Some other channels possess up-and-down profiles owing to the accumulation of peat and/or mass-movement from their sides. There remains, however, a very significant number of other channels whose up-and-down profiles are due only partly or not at all to

deposition. At least six explanations have been put forward in the literature to account for such channels. These are as follows: (i) "unequal volumes of meltwater flowing in opposite directions more or less contemporaneously [63]"; (ii) postglacial stream erosion [64]; (iii) erosion of a channel whose upstream part is in drift and downstream part in rock, resulting in an uphill long profile in the more easily eroded drift [65]; (iv) erosion of a channel across a col previously lowered by a "preglacial" stream working backwards by headward erosion, resulting in the floor of the "preglacial" valley forming the uphill part of the profile [66]; (v) reversal of meltwater flow [67]; (vi) subglacial erosion by water flowing under hydrostatic pressure [68].

The first four explanations need be considered only briefly. The first one seems rather improbable and can hardly account for channels that are continuous well-marked features throughout their lengths. The second seems equally inapplicable in most instances, for postglacial streams are often absent or of small dimensions in the channels under consideration. Certainly in the example cited by Eckford [69] (Windy Gowl channel, Peeblesshire, Nat. Grid Ref. 36/1454) the half-mile uphill part of the channel could not have been excavated by the present tiny stream that occupies it: this stream has, in fact, cut a small notch a few tens of yards long into the floor of the uphill part of the channel. The third explanation applies to a particular case and cannot account for the numerous up-and-down channels where different geological conditions obtain. The fourth, although put forward by Twidale [70] as an explanation of up-and-down channel profiles, and hence listed above, is not really such. The uphill slope is simply the floor of a preglacial valley and the floor of the channel proper slopes in the same direction throughout its length. Such channels were long ago described by Kendall and Bailey [71].

The fifth explanation—reversal of meltwater flow through channels—has been by far the most commonly suggested. It was put forward originally by Kendall in 1902 [72]. In the following year he repeated the explanation, with some elaboration, apparently as a result of criticisms made by members of the Yorkshire Geological Society [73]. Subsequently, however, he appears to have become uncertain about this interpretation, for Elgee quotes a letter from Kendall on the problem of up-and-down channels as follows: "Experience gained in the last two years has made me suspicious of cases of apparent flow in opposite directions through the same channel. It is easier to assume that a transverse watershed or 'corrom' has been produced by inflow of a lateral stream, or by some similar means [74]". Much more recently Peel has discussed the reversal of drainage idea and has said that it suffers from the disadvantage "that it requires special conditions unlikely often to be fulfilled [75]".

The reversal of flow concept is certainly difficult to apply if one attempts to explain channels as lake overflows associated with an impenetrable ice-barrier, but these difficulties may disappear when the importance of drainage into the ice is recognised. For example, in the small area shown in Figure 9 the large number of subglacial chutes points to abundant drainage down the hillside beneath the ice. The mutual relations of some of the chutes show that they did not all

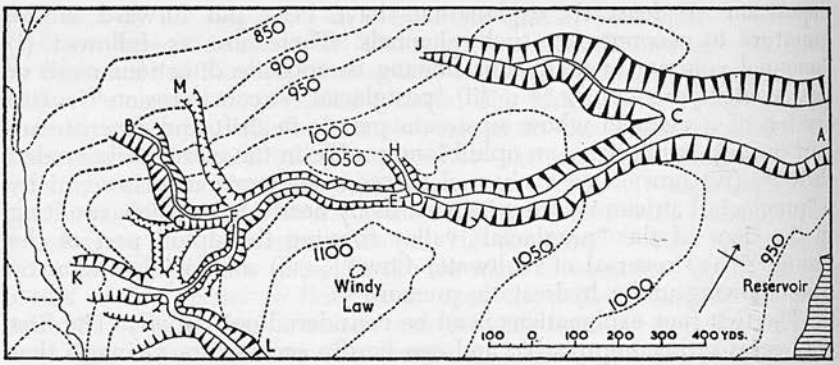


Fig. 12. An up-and-down channel (ADB and related channels on Lammermuir scarp some 4 to 5 miles SW of Gifford, East Lothion. Letters explained in text.
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function contemporaneously, and one can imagine how, with the blockage of one subglacial tunnel an adjacent one might have tapped the marginal or submarginal drainage, resulting in local reversal of this drainage and an up-and-down profile. Four up-and-down profiles do occur in the area, and at least one of them (D, Fig. 9) is very difficult to explain by any other hypothesis than reversal, for from the highest point on its floor the channel descends very steeply in both directions into a subglacial chute.

In addition to very localised reversal of flow such as that just cited, an up-and-down profile can be developed as a result of reversal of drainage on a larger scale. The example shown in Figure 12 was described by Kendall and Bailey [76]. The floor of this up-and-down channel (ADB, Fig. 12) lies at about 920 feet at its eastern end, 1,040 feet at its highest point (E) and 900 feet at its western end. Rock outcrops are frequent along the channel sides and exclude the possibility that the deposits that occupy the channel floor in places affect significantly the very marked up-and-down profile. Further up the same spur channel LK begins in a pre-existing col and descends steeply westwards, while adjacent minor channels slope westwards also. This implies that meltwaters first flowed westwards across the spur. The final stage of this westward flow is represented by EGB and EFM, the downslope parts of these two channels being excellent examples of subglacial chutes. The section DA and its tributary JC, as well as other channels and depositional evidence to the north, all indicate eastward flow. The up-and-down channel ADB is thus a composite channel, DB having been formed first and DA (or the greater part of it) later. The eastward-flowing waters that cut DA entered it via HD, which channel starts abruptly at H, indicating it was cut by a stream flowing out of (or less probably, off the surface of) the ice.

That this particular instance of reversal is not due to the fortuitous local development of escape routes into the ice but has a wider significance is strongly suggested by the fact that along the steep Lammermuir scarp for some fifteen miles to the north-east a great series of marginal and submarginal channels and kame terraces occurs, whereas

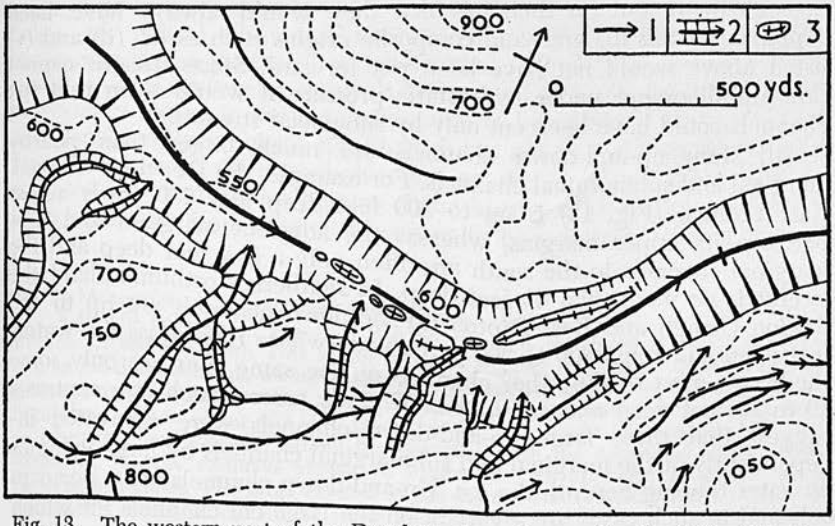


Fig. 13. The western part of the Deuchrie channel, some 5 miles SW of Dunbar, East Lothian. 1. Small channel; 2. Larger channel; 3. Esker. The floor of the Deuchrie channel rises eastwards as far as the small cross-line shown on the thickline marking the centre of the channel floor. *Crown Copyright reserved.*

to the south-west for some fifteen miles not a single feature of this type is present, there being instead a few subglacially-engorged eskers and possibly one or two subglacial channels. This contrast appears to be the result of the presence of two ice-sheets. To the north-east of channel ADB the presence of Highland ice, with a slight surface-slope towards the Lammermuirs, favoured marginal and submarginal drainage along the scarp face. To the south-west of the channel the Southern Upland ice-sheet covered the scarp and adjacent lower ground and its slope was slightly away from the scarp. The reversal of flow is probably related to differing rates of wastage of the two ice-sheets, the full significance of which has yet to be worked out.

Although reversal of flow can explain some up-and-down channels there are many others where it appears an unsatisfactory explanation. For example, up-and-down channels, sometimes of large dimensions, frequently occur as links in extensive glacial drainage-systems where all the other channels and related deposits point to glacial drainage in one general direction. It seems highly improbable that reversal of flow should have taken place in such areas, leaving as its only evidence the up-and-down channels themselves. In such instances the only adequate explanation seems to be erosion by subglacial waters flowing under hydrostatic pressure, an explanation that has been applied on the Continent, particularly to the tunneldale and Rinnentäler, which often have uphill profiles in places. Certain difficulties that may be encountered by this hypothesis have been mentioned by Peel [77] and have been discussed elsewhere [78]. Three points only will be mentioned here.

(i) Many channels with up-and-down erosional long-profiles are continuous features throughout their lengths, strongly suggesting that they were formed by streams of meltwater flowing in a single direction

through them. But for their profiles they would always have been explained in this manner, and composite origins such as (i), (ii) and (v) listed above would not have been put forward. Since streams cannot flow uphill except under hydrostatic pressure it would seem that the channels could have been cut only by subglacial streams.

(ii) Many up-and-down channels are much larger than nearby marginal and submarginal channels. For example, the Deuchrie channel, East Lothian (Fig. 13) is up to 200 feet deep and 250 yards across between its upper margins, whereas the adjacent marginal and submarginal channels to the south are often only a few feet deep and the largest is no more than 40 feet deep. In southern Northumberland the Beldon Cleugh and East Dipton up-and-down channels are up to 150 feet deep and some 200 to 300 yards wide where they cross the watersheds, yet most of the other channels on the same spur are only some 10 to 30 feet deep and a few tens of yards across. Such size contrasts suggest that these large up-and-down channels were excavated independently of the marginal and submarginal channels by large volumes of water flowing beneath the ice. Up-and-down channels often occur in cols and in such cases are a variant on the large col channels for which a subglacial origin was suggested in Part I of this paper.

(iii) Some up-and-down channels are so related to other channels and/or fluvio-glacial deposits as strongly to suggest a subglacial origin. The Deuchrie channel again serves as an example. As can be seen in Figure 13, the southern side of this channel is cut into by a number of smaller channels. These channels are subglacial chutes related to the marginal and submarginal channels further up the hillside. An esker lies on the floor of the Deuchrie channel, and its subglacial origin is proved by an exposure showing some two feet of ablation moraine containing angular blocks up to two feet across lying on top of the rounded, stream-deposited gravels of the esker proper. The Deuchrie channel must have been excavated therefore before the esker and subglacial chutes were cut. Hence it would seem that it also was excavated subglacially. The only alternative is to assume it was cut subaerially before the final deglaciation. It may well be that the channel was partly excavated prior to the final deglaciation, but this can hardly be the whole explanation, for the freshness of the channel form points to considerable erosion during the final deglaciation. Furthermore, pushing the origin of the channel back in time evades the main issue, for it does not help to explain its up-and-down profile.

CONCLUSION

It is hoped that the preceding discussion of certain types and systems of meltwater channels will help towards a better understanding of these features. The author is well aware that much has been omitted: for example, little reference has been made to the complications resulting from multiple glaciation, the concept of the englacial water-table has been mentioned only briefly, the possible implications of extensive drainage into the ice in terms of large subglacial and englacial river systems have been barely touched upon, and the depositional features produced by glacial drainage have been given insufficient attention. These omissions are in part simply a result of considerations of

space, but in part they too result from the inadequacy of our knowledge of glacial drainage features in Britain. Despite frequent statements to the contrary, we understand very little of the origin and significance of meltwater channels and related deposits. A great quantity of evidence stored in the literature needs to be re-interpreted and much additional evidence whose significance has not as yet been appreciated must be recorded. It is probable that as this is done there will be revealed a very different picture of the history of deglaciation in many parts of Britain from that which seems to be generally held at present.

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[2] It will be seen in Figure 5 that 26 channels terminate at about the same altitude (c. 450-475 feet). This might seem to suggest that these channels drained into an ice-dammed lake, an interpretation that might seem to be supported by the existence at the eastern end of Strath Allan of a large meltwater channel cut down from about 450 to about 400 feet, by which the lake might have overflowed. Such an interpretation, however, cannot be reconciled with the subglacial origin already suggested for the channels shown in Figure 5. The apparent anomaly can be resolved as follows. The large channel at the eastern end of Strath Allan carried the meltwaters eastwards out of the strath and controlled the level of the water in the strath itself to the west. But this water did not exist as an open lake. Numerous steep-sided kames, many of them with flattish tops at about 400 feet O.D., together with associated kettles, show that the floor of the strath was, in the later stages of deglaciation, occupied by masses of dead ice around which much of the fluvioglacial material was laid down, the upper limit of deposition being controlled by the large channel. At a slightly earlier (and partly contemporaneous) stage, when ice still covered the slopes shown in Figure 5, the meltwaters flowed down beneath the ice and eroded the channels, but erosion could not extend below the level of the large outlet channel at the eastern end of Strath Allan, for this controlled the level of water in the tunnels, crevasses and other openings within the ice itself. This level may be referred to as the 'englacial water-table'. The influence of the water-table on meltwater drainage in the Eddleston valley, Peeblesshire, has been discussed elsewhere (*Geogr. Ann., Stockh.*, 1958). The idea has been applied for some time in Scandinavia and plays a major part in Gjessing's (1960) interpretation of deglaciation in part of central Norway.

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8

A Subglacial Drainage System by the Tinto Hills, Lanarkshire

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ABSTRACT

A system of meltwater channels and related eskers and other deposits situated by the Tinto Hills in south Lanarkshire is described and interpreted as largely of subglacial origin. The Thankerton ' kame ' is reinterpreted as an esker system and Charlesworth's Lake Clyde is discussed. Evidence is given for the existence of stagnant ice in the Tinto area during the formation of the glacial drainage features and the application of the stagnation hypothesis to a large part of the Southern Uplands is briefly discussed.

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I. PRINCIPAL RELIEF FEATURES

The Tinto Hills are a steep-sided east-west ridge formed mainly of felsite. The ridge exceeds 2300 ft. at its highest point, the summit of Tinto (Fig. 1). The broad valley of the northward-flowing Clyde swings round the eastern end of the Tinto Hills and a broad tongue of lowland extends westwards from the Clyde valley along the northern side of the ridge. The low drift-covered ground of the Clyde valley lies between altitudes of about 600 and 800 ft., ending at the latter altitude in a sharp concave break of slope at the foot of the Tinto Hills. Along the northern side of the hills a belt of drift-covered ground with moderate to gentle slopes, its upper limit at about 1000 ft., lies between the steep rock slopes of

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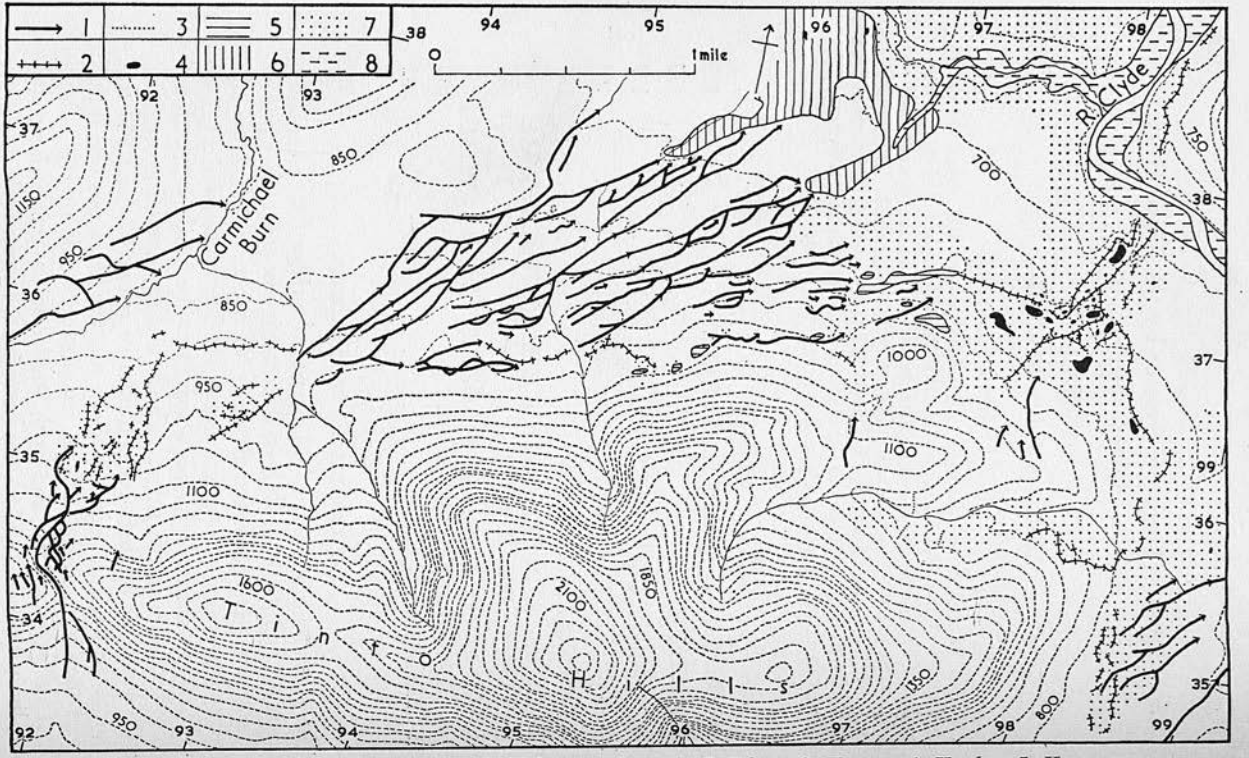


FIG. 1. The Tinto glacial drainage system. 1. Meltwater channel. 2. Esker. 3. Kame. 4. Kettle. 5. Kame terrace. 6. Flat spread of fluvioglacial sand and gravel. 7. Other areas of fluvioglacial sand and gravel. 8. Alluvium.

the Tinto ridge and the flattish central part of the lowland situated at about 700 to 850 ft. It is in this intermediate zone of moderate to gentle slopes that meltwater channels are best developed.

II. Methods

The area was first studied stereoscopically on aerial photographs at approximately 1:10,000 scale, all identifiable glacial retreat features and drift boundaries being marked on the photographs. The marked photographs were used as a base for field mapping, which was carried out on a transparent overlay. The completed field maps were subsequently transferred to Ordnance Survey six-inch sheets with the aid of a Grant Projector, various precautions being taken to reduce the effects of distortion in the photographs. This method of mapping permits accurate location of features.

Accuracy cannot be claimed, however, for some of the limits of fluvioglacial sand and gravel areas shown on the accompanying maps. Where sections are not available it is often difficult to differentiate between till and stratified drift since the till often contains numerous rounded and partly-rounded stones from the Old Red conglomerate, while the stratified drift contains much felsite, which remains angular or subangular even in eskers. Furthermore, much of the stratified drift was laid down subglacially and was covered in places by ablation moraine when the ice above subsequently wasted away, so that the limits of the stratified drift, even where exposures exist, have sometimes to be mapped arbitrarily owing to the complex boundaries. These limitations of the drift mapping do not affect the deductions made in the following pages.

III. The Western Area

Fig. 1 shows the glacial drainage features of the whole area and Figs. 2, 3 and 4 show them in greater detail. The features will be considered from west to east.

The main glacial drainage forms of the western area (Fig. 2) are a system of channels crossing the Tinto ridge by a col and an esker system that continues the channels northwards. The largest channel lies in the floor of the col and is cut as deep as 60 ft. in rock. On the slopes of the col on either side of the main channel are other channels, ranging from 30 ft. deep to small steps a few feet broad. In addition to the features marked on the map, faint benches and grooves cross the ridge in the vicinity of the col. These features may have been partly produced by meltwaters, but since this is not certain they are not shown on Figs. 1 and 2.

The greatest development of channels is on the northern slopes of the ridge, suggesting that the channels were cut by northward-flowing waters. This direction of flow is confirmed by the fact that the channels are continued by eskers, the bedding of which

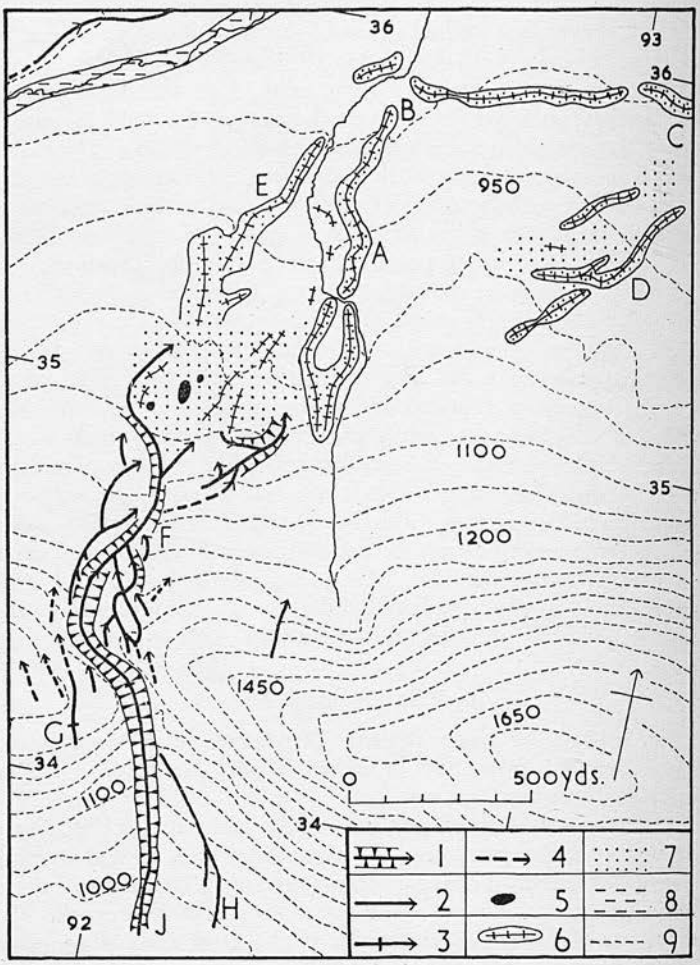


FIG. 2. Glacial drainage features of the western part of the Tinto area. 1. Larger meltwater channel. 2. Other meltwater channel. 3. Up-and-down channel, cross-line indicating highest point on channel floor. 4. One-sided channel. 5. Kettle. 6. Esker. 7. Other areas of fluvioglacial sand and gravel. 8. Alluvium. 9. Contour.

indicates northward flow. The location of the channels in a col with the deepest channel in the bottom of the col may seem to suggest that they were formed by the overflowing waters of a lake held up by ice on the south side of the col. This interpretation is unsatisfactory, however. Not only is there no evidence of lake shorelines, bottom deposits or deltas, but the complex anastomosing system of channels in the col could hardly have been produced by a lake overflow. The channels are much more satisfactorily explained as having been formed subglacially. The channels that branch off from the main channel (and sometimes rejoin it farther north) probably resulted from collapse of the ice-tunnel in which the main channel was cut, causing temporary blockage of that channel and diversion of the meltwaters flowing through it. The small independent channels and benches on the slopes of the col were probably cut by independent subglacial streams. That they are not marginal features is indicated by their steep gradients and, in one instance (G, Fig. 2), by an up-and-down long profile.

The main channel also has a long profile of this type. From the highest point on its floor at 1246 ft. (O.S. spot height) the channel descends northwards to just above the 1000-ft. contour. On the southern slope of the ridge it can be followed at least as far down as the 1000-ft. contour (J, Fig. 2). This part of the channel is neither as deep nor as sharply defined as the part on the northern slope. Nevertheless it is a well-marked feature 20 to 25 ft. deep and about 50 yd. across occupied only by a tiny stream. One way in which the southern part of the channel may have been formed is by erosion by meltwaters flowing uphill under high hydrostatic pressure. This explanation has the merit of according with the abundant evidence of northward-flowing subglacial streams to the north of the watershed. On the other hand, it has the disadvantage that it requires a subglacial stream to have eroded while climbing uphill through a vertical distance of 250 ft. An alternative explanation is local reversal of the subglacial drainage. This would allow the isolated channel H to be explained as due to the southward drainage also. However, apart from channel H and the up-and-down channels, there appears to be no other evidence that can be cited in favour of temporary glacial drainage to the south. The writer therefore remains uncertain about the origin of the southern part of channel JF.

The col channel system ends northwards in an area of sand and gravel. The upper (southern) part of this deposit has a fairly smooth surface that slopes towards the north and resembles an alluvial fan. Northwards this feature merges into low mounds, which

increase in size and sharpness to form eskers. Esker A is the sharpest feature, being about 20 ft. high with steep sides, winding course and slightly-undulating crest. Esker E is broader and more subdued, a difference partly accounted for by the presence of a minor road along its crest. Between these two features are several short esker ridges.

If, as suggested above, the channels to the south of the deposits were formed subglacially, then the deposits themselves must be of subglacial origin also, for their position and orientation show that they were laid down by the streams that cut the channels. Independent evidence for a subglacial origin is provided by the steep sides and well-preserved bedding of esker A. Furthermore, angular stones up to one foot across rest on this esker and at one point a section showed till overlying current-bedded sands. The whole system of deposits closely resembles systems in central Norway described by Gjessing (1960) and shown by him to have been formed subglacially.

The streams that cut the channels in the col could have functioned only as long as the ice surface was at least as high as the bottom of the col, that is, at least 1250 ft. O.D. The related deposits extend down to about 870 ft. O.D. Therefore the subglacial streams that deposited the eskers must have penetrated at least 350 to 400 ft. beneath the surface of the ice down the northern slope of the Tinto ridge. It will be noted that the kettles and partly-closed depressions associated with the eskers and subglacial fan must also have been formed at the base of the ice far below its surface.

The two principal eskers converge on point B and here turn sharply eastwards as a single low ridge (BC) running along the 900-ft. contour. The reason for this sudden change of direction is uncertain. The form of the surrounding ground does not explain it, for this is an almost featureless, till-covered, gentle slope. Presumably the meltwaters found a major line of weakness in the base of the ice and exploited it. A few hundred yards farther south a small group of eskers (D) represents another route followed by the meltwaters that cut the channels through the col.

IV. The Central Area

Eskers BC and D are continued eastwards by the system of channels shown in Fig. 3. All these channels are less than 25 ft. deep and some of them are only a few feet deep. The long profiles of almost all the channels slope down towards the east-north-east and a common gradient for the more gently inclined ones is 2 per

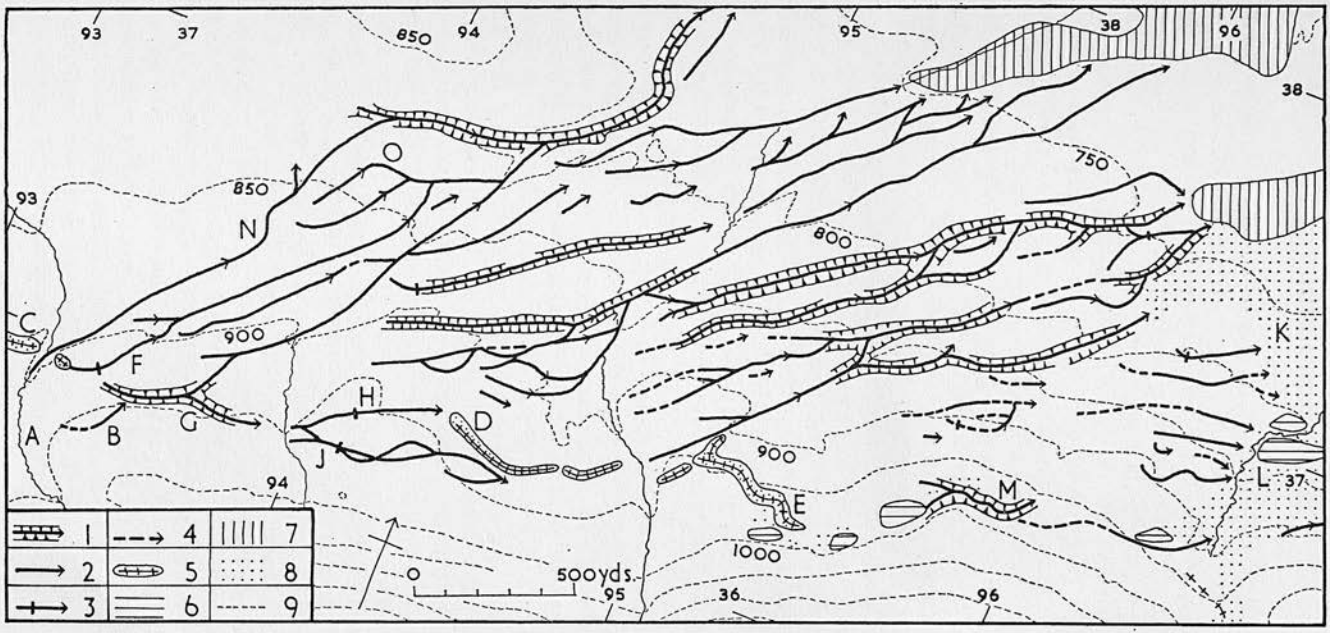


FIG. 3. Glacial drainage features of the central part of the Tinto area. 1. Larger meltwater channel. 2. Other meltwater channel. 3. Up-and-down channel, cross-line indicating highest point on channel floor. 4. One-sided channel. 5. Esker. 6. Kame terrace. 7. Flat spread of fluvioglacial sand and gravel. 8. Other areas of fluvioglacial sand and gravel. 9. Contour.

cent. Much of the channel system consists of nearly parallel channels linked in places by cross-channels. This approximate parallelism of the main channels may seem to suggest that they were formed along a receding ice-margin. This view might seem to be supported by the fact that in places some of the channels are merely benches, the missing side of such channels having been the ice itself. Channels of this type have often been used to deduce former ice-margin positions. That the marginal interpretation is untenable, however, can be shown in various ways.

For example, the fact that the system as a whole spreads out from a single point, the eastern end of esker C, suggests that the channels were formed by meltwaters that flowed along, or near, the line of the esker. (Two exceptions will be mentioned below.) Since this esker is a subglacial feature, it is difficult to see how the channels that continue it can be other than subglacial also. Furthermore, throughout the time the esker stream flowed, the ice surface at the col only a mile away must have been at least 350 ft. higher than point C. Therefore, unless the ice surface had an abnormally steep gradient, at least the western part of the channelled area shown in Fig. 3 must have been ice-covered at the time the channels were cut.

Another reason for rejecting the marginal interpretation is that it would require the ice-margin to have remained stationary at point C (Fig. 3) while it retreated approximately a mile to the north at the eastern end of the channelled area. This retreat would have been associated with a thinning of the ice of some 300 ft. (*cf.* contours on Fig. 3). Thinning of this magnitude could hardly have failed to affect the ice-margin at point C, causing it to retreat there also. Among other evidence indicating a subglacial origin for some of the channels is the existence of an esker (DE) continuing one of them and the occurrence of six channels with erosional up-and-down long profiles.

Several of these anomalous channels occur in the south-western part of the area shown in Fig. 3 (F, G, H, J). Channel G is in line with esker C and begins 100 yd. east of the end of the esker. This channel is about 15 ft. deep with a nearly level floor as far as the point where it bifurcates. From here the southern branch becomes gradually shallower and its floor rises eastwards until it dies away 600 yd. east of the end of esker C. After a gap of 100 yd. channels H and J begin. Both these channels ascend the western slope of a low spur for about 200 yd. before descending its eastern slope. Where channel H dies away, esker DE begins. The total eastward increase in altitude of the channel floors from the beginning of channel G to the highest point in channels H and J is about 40 ft.

This sequence of glacial drainage channels, comprising esker C, channels G and H, and esker DE, clearly shows that up-and-down channels can be formed subglacially. One cannot postulate reversal of flow, for the glacial drainage system as a whole shows that meltwaters flowed in the same general direction for several miles, and the only evidence of meltwater flow in the opposite direction, if the reversal hypothesis were adopted, would be the 'uphill' parts of the up-and-down channels. On the subglacial interpretation an ample head of water would have been available to force the subglacial waters uphill, since the floor of the large col channel where it crosses the watershed of the Tinto Hills is more than 250 ft. above the highest point on the up-and-down channel floor.

The eastern part of esker DE makes a sharp turn to climb some 80 ft. up the hillside. It is continued by small patches of gravel whose flattish accordant tops, lying just below the 1000-ft. contour, suggest the remains of a kame terrace. This line of deposits then leads into a channel (M), which runs along the hillside for a short distance before turning sharply down the slope and dying out. A branch of channel M continues eastwards to end in the extensive sand and gravel area shown in the eastern part of Fig. 3. The kame terrace fragments and the part of channel M running along the hillside may have been formed along the ice-margin. The uphill part of esker DE would then mark the route by which the subglacial stream climbed the hillside to emerge at the ice-margin, while the sharp bend in channel M would mark one of the routes by which the waters flowed back into the ice. Although this explanation is attractive, it will be noted that an entirely subglacial interpretation is feasible also. The kame terrace fragments and channel M could have been formed by the stream that built esker DE as it flowed along the hillside beneath the ice.

In general, the higher, more southerly channels shown in Fig. 3 were formed first and were successively replaced by lower, more northerly routes. This can be deduced at points where channels join or branch, for the relative amounts of incision indicate their relative ages. The abandonment of a channel for a slightly lower, sub-parallel one frequently took place by the short cross-channels, and it appears that in some instances two adjacent subparallel channels functioned contemporaneously.

As mentioned above, two exceptions occur to the generalisation that the channels were cut by waters that had flowed along or near the line of esker C. One of these, channel B, Fig. 3, is a very faint feature and was formed by the waters that deposited esker system

D shown in Fig. 2. The other exception is channel N and certain other channels, such as O, that appear to have been cut wholly or partly by waters that had passed along part of channel N. This channel leads away from the bottom of the small incised valley of the Cleuch Burn (A, Fig. 3) where this burn cuts through esker C. Since the burn flows at right-angles to the esker where it breaches it and has cut down some 15 to 25 ft. below the top of the esker, the waters that cut channel N must have come down the Cleuch Burn valley. Channel N and related channels thus represent subglacial courses of the Cleuch Burn.

V. The Eastern Area

In the eastern part of the area shown in Fig. 3 the meltwaters ceased to erode channels and instead deposited large quantities of sand and gravel. The more northerly channels end in an extensive spread of these deposits, which is almost flat except where subsequently dissected by streams. Farther south around K (Fig. 3) the fluvioglacial deposits mantle the slopes immediately east of the ends of the channels, but have little definite form. Still farther south the deposits form two terraces. One of these, terrace L, is up to 70 yd. broad and was clearly formed by the streams that cut the channels immediately west of it. The terrace slopes down gently towards the east. After about 600 yd. it gradually changes to a low ridge, which becomes higher as it is followed eastwards, until it forms a well-marked esker (M, Fig. 4), about 15 ft. high, bordered by kettles. After continuing for about half a mile the esker turns abruptly through a right angle to form part of the conspicuous series of sand-and-gravel ridges that crosses the Clyde valley at this point.

These, and the other sand-and-gravel ridges shown in Fig. 4 are referred to in the literature as the Thankerton Kame. As can be seen from Fig. 4, the part TO consists of several ridges associated with kettles. The ridges vary from low features a few feet high to steep-sided features 25 ft. or more high. The system is interrupted by the Clyde, north of which it is resumed as a single ridge up to 25 ft. high (Figs. 1 and 4, R). Traced south-westwards the ridge complex TO changes to a single, low, broad feature (TN, Fig. 4) that dies away as it is followed up the slope. Another ridge, US, joins the main complex from the south. This winding ridge is mostly about 15 ft. high and, apart from four short breaks, extends continuously for a mile to the south. At U the ridge changes direction and wanders up the side of a valley leading down from

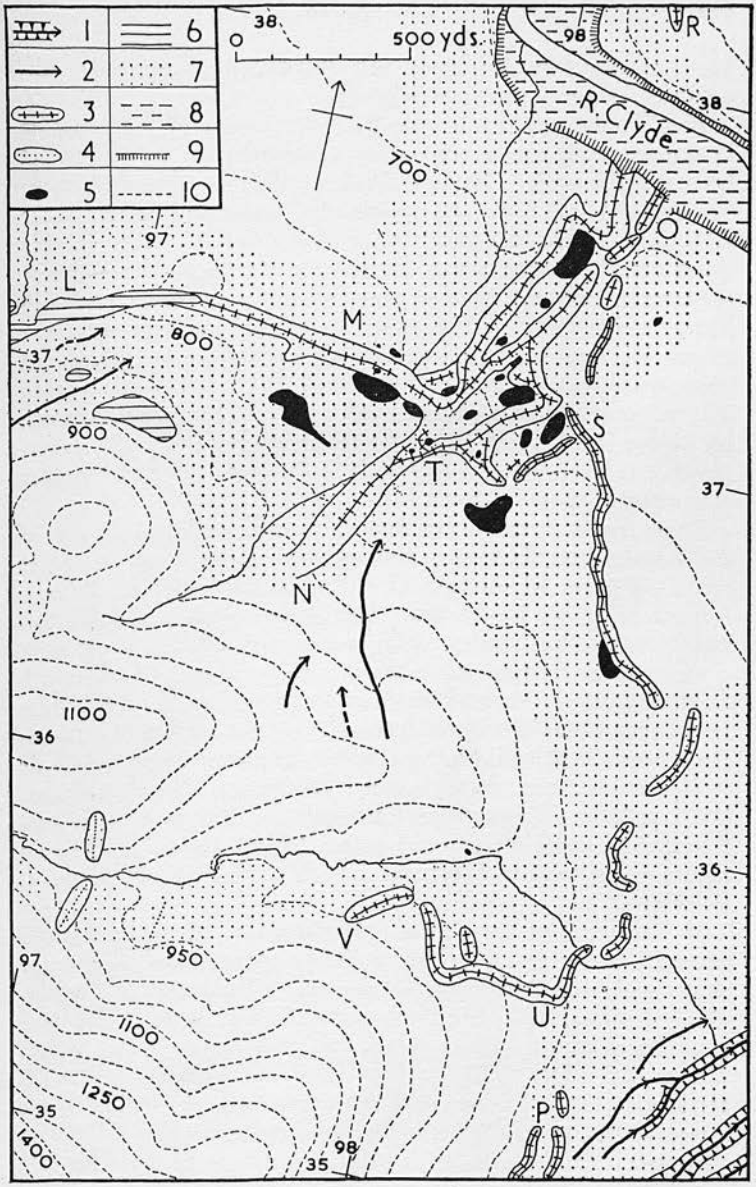


FIG. 4. The Thankerton esker system. 1. Larger meltwater channel. 2. Other meltwater channel. 3. Esker. 4. Kame. 5. Kettle. 6. Kame terrace. 7. Other areas of fluvioglacial sand and gravel. 8. Alluvium. 9. Bluff bordering flood-plain. 10. Contour.

Tinto, as far as point V (Fig. 4). The line of SU is continued southwards by two more sinuous ridges (P).

The Thankerton ' kame ' has been the subject of some dispute in the literature and a satisfactory explanation of it has not yet been put forward. Gregory (1915, p. 475) concluded that the ' kame ' was built along the margin of a glacier that descended the Kirk Burn valley (the valley containing ridge VU, Fig. 4). He stated that the course of the ' kame ' is crescentic and lies opposite the mouth of this valley. It is very difficult indeed to reconcile this interpretation with the ridges shown on Fig. 4, even if one ignores ridges VU and R (as Gregory did). Furthermore, Gregory's statement that the ridges are unstratified (1915, p. 475) is incorrect. The writer agrees with Charlesworth's view (1926b, p. 32), accepted by Linton (1933, p. 554) that the ridge US is an esker. Charlesworth remarked that this feature is ' a linear ridge of whose esker origin there can be no doubt.'

These two authors did not apply this explanation to the rest of the ridge system, however, both of them stating that the ridges extending from M through O and northwards beyond R were formed along an ice-margin. The only reason given for this interpretation (by Charlesworth) is that it is difficult to picture a glacial stream forming an esker in this position. Charlesworth deduced from the down-valley concavity in plan of ridge system MOR that these ridges were formed along the margin of ice from the Scottish Highlands during a halt in its retreat down the Clyde valley, an interpretation with which Linton was 'in fullest accord' (1933, p. 553). The present writer finds this interpretation unsatisfactory and considers that the ridges MOR, as well as NT, are eskers. The principal reasons for this reinterpretation are as follows:—

(i) The form of the ridges is characteristic of well-developed eskers in all respects. The section TO comprises mainly steep-sided ridges with slopes exceeding 20 degrees. The kettles and partly-closed depressions (the latter not marked in Fig. 4) show that these steep slopes are ice-contact slopes, thus proving that ice lay on *both* sides of each ridge. The occasional bends in the ridge system MTO are characteristic of eskers, and the single ridge R, continuing the system north of the Clyde, is slightly sinuous with an undulating crest-line. The latter ridge is identical in form to the ridge US, which both Charlesworth and Linton unhesitatingly interpreted as an esker. Again, as is characteristic of eskers, the highest points of the ridges frequently occur at bends. The maximum height of the whole feature, some 45 ft., occurs at a right-angle bend.

(ii) The ice-marginal interpretation encounters the serious difficulty that it requires a large number of closely-spaced streams to have issued from the ice and deposited their loads so as to produce the continuous ridges. On that hypothesis one would expect a series of fans or deltas to have been formed, rather than long, narrow ridges parallel with the ice-margin.

(iii) The writer has not found any Highland erratics in the Thankerton 'kame' and it is significant that neither Charlesworth nor Linton reported the occurrence of any. On the other hand, large quantities of felsite occur, as noted by Gregory (1915, p. 475). If the 'kame' were produced by streams draining Highland ice this felsite must have been derived mainly from the limited outcrop of felsite on Cairngryfe Hill some three miles to the north-west of the 'kame,' a source that seems inadequate. If, however, the 'kame' is, in fact, an esker system, the abundance of felsite is readily accounted for as having been derived from the great mass of felsite forming the Tinto Hills. The drift in which the channels shown in Fig. 3 are cut is largely composed of fragments of this rock.

(iv) The pattern of ridges shown on Fig. 4 seems to be satisfactorily explicable only in terms of a subglacial river system. Thus ridge M, as already mentioned, is an esker continuing terrace L (formed subglacially) and some of the meltwater channels shown in Fig. 3. Another esker (NT) comes from the south-west. Since this esker dies away gradually as it is traced up the slope towards a small valley leading down from Tinto, it is reasonable to conclude that it was formed by a stream flowing down this valley. In the next valley to the south another esker was formed in a similar manner. Here esker VU leads down from the mouth of the valley and joins esker PU coming from the south. The Thankerton 'kame' is thus an esker system that shows clearly the courses of several subglacial streams that flowed from the slopes of Tinto and along its flanks to the low ground of the Clyde valley.

VI. Local Significance of the Tinto Glacial Drainage System

The preceding discussion of glacial drainage features by the Tinto Hills has a bearing on certain local glacial problems, which will now be considered.

Direction of ice-movement. Ice masses from several different sources have figured in discussions of the Tinto area. Originally it was supposed that only Southern Upland ice affected the area (*e.g.* J. Geikie 1894). Gregory (1915) added the concept of a glacier

descending from Tinto and Charlesworth (1926b) added the Highland ice. As shown above the evidence is inconsistent with the presence of Highland ice in this area at the time the Thankerton eskers and related glacial drainage features were formed. The concept of a Tinto glacier has been effectively dismissed by Linton (1933). There remains, therefore, only the Southern Upland ice. Gregory considered this ice to have taken the form of a glacier pushing northwards down the Clyde valley as far as the Carstairs 'kame.' This view is not consistent with the evidence provided by striae or the transport of erratics. A. Geikie (1873, p. 42) stated that along the flanks of the Tinto Hills striae show that the ice moved from south-west to north-east. Felsite erratics derived from Tinto extend north-eastwards past the southern end of the Pentlands into Midlothian (McCall and Goodlet 1952). This evidence indicates a general south-west-to-north-east ice movement, as J. Geikie (1894) concluded. The general north-eastward trend of the Tinto subglacial drainage system accords with this evidence, for subglacial stream systems tend to be aligned parallel with the direction of ice-movement (or, if developed under stagnant ice, with the direction of former ice-movement).

Lake Clyde. As mentioned above, Charlesworth (1926b) interpreted part of the Thankerton esker system as an end-moraine related to the retreating Highland ice. He believed that the Carstairs 'kame,' some six miles to the north, marked another stage in the retreat of this ice. This interpretation inevitably led him to conclude that a large lake existed in the middle Clyde valley, overflowing by the Biggar gap. A serious difficulty encountered by this interpretation is that, although the bedding and relatively fine grade of the moundy deposits south of the main Carstairs ridge system point to the existence of a lake, the altitude of the deposits does not accord with the concept of a lake overflowing through the Biggar gap. The floor of this gap is at an altitude of about 670 ft. on the Clyde–Tweed drainage divide, but the summits of most of the sandy mounds in the Carstairs area range in altitude from above 750 ft. down to slightly below 700 ft. It might seem that one way round this difficulty is to postulate erosion of the Biggar gap by meltwaters contemporaneously with deposition in the Carstairs area, so that the level of Lake Clyde was gradually lowered. This explanation would be unsatisfactory, however, for there is no evidence of such erosion in the Biggar gap. An esker extending almost to the floor of the gap shows that significant erosion has not taken place. In fact Linton, who accepted Charlesworth's Lake Clyde, stated that meltwaters aggraded the floor of

the Biggar gap, thus increasing the difficulty of reconciling the altitude of the Carstairs deposits with that of the gap.

It seems that the only way of resolving this difficulty is to assume that the Biggar gap was not open to meltwaters at the time the Carstairs deposits (or most of them) were laid down. This is consistent with the conclusion reached above that the ice in the middle Clyde area came from the Southern Uplands.

Three aspects of the glacial drainage evidence in the Thankerton area point to a further conclusion:

(i) None of the meltwater channels in this area extends below an altitude of 700 ft., although several terminate a little above this altitude ;

(ii) The altitude of the Thankerton esker system, as a whole, declines northwards and the feature ends at about 700 ft.;

(iii) An extensive area of fluvioglacial deposits, partly formed by the meltwaters that cut some of the channels shown in Fig. 3, lies north and west of Thankerton village (Figs. 1 and 3). These deposits, which in their eastern part consist largely of sand, are usually flat-topped at about 700 ft. and include the extensive, almost flat area known as Thankerton Moor. This evidence points to the existence of a lake at an altitude of about 700 ft. in this part of the middle Clyde valley.

Similarity of altitude suggests that this was the same lake as that in which the lower mounds south of the Carstairs ridge system were laid down.[1] During the formation of the higher mounds of the Carstairs area the lake overflowed through the major channel that leads from the Clyde drainage to that of the Tweed near Garvald House (Nat. Grid Ref. NT 095488). This flat-floored depression, which is followed by the canalised Garvald Burn–Tarth Water and extends for more than seven miles, clearly carried a large volume of water. It is not clear, however, how long this channel continued to function since its floor level on the Clyde–Tweed divide appears to have been built up to its present altitude of about 735 ft. by later deposition by the South Medwin Water and Garvald Burn. If the channel infill is of considerable thickness the channel could have continued to serve as the lake overflow while the lake was lowered to the 700-ft. level indicated by the evidence near Thankerton. Even if this route was not available, however, the Biggar gap cannot be invoked as an alternative, subglacial, escape route for the waters of the 700-ft. lake, since the Thankerton esker system shows that

[1] That this lake was partly occupied by masses of stagnant ice, at least in the Carstairs area, is indicated by the moundy form of the deposits laid down in it there.

the subglacial drainage of the Thankerton area was northwards at the time this lake existed. The lake must have been drained therefore by a route lying to the west or north of the area considered here, for no other point to the south and east is at a sufficiently low altitude.

Further discussion of this problem is beyond the scope of this paper, for it must be linked with a reinterpretation of the Carstairs 'kame.' For the same reason it is not possible to decide whether or not the Biggar gap functioned as an overflow for the Clyde lake after the Southern Upland ice had vacated it. The gap certainly carried the waters of the upper Clyde, however, for a well-defined river terrace, some 10 to 15 ft. above the Clyde flood-plain and according approximately in level with the floor of the gap, extends up the Clyde valley from the entrance to the Biggar gap, but is absent from the Clyde valley below this point. It is not known when the Clyde ceased to flow through the Biggar gap to resume its present course. A. Geikie has pointed out that in times of high flood some of the waters of the Clyde still pass through the gap into the Tweed (Geikie 1901, p. 378).

The Tinto nunatak. No attempt has been made to reconstruct positions of the ice-margin in the Tinto area, owing to the absence of definite marginal drainage features. The height of the Tinto Hills above the surrounding ground leaves little doubt, however, that they formed a nunatak rising well above the surface of the ice at the time the glacial drainage features discussed above were formed. Eskers VU and NT in Fig. 4 and channel N in Fig. 3 show that some, at least, of the streams that ran down the slopes of this nunatak went beneath the surrounding ice.

VII. Regional Significance of the Tinto Glacial Drainage System

The various eskers that are so well-preserved in the Tinto region could not have survived unless the ice around them had become stagnant by the time they were formed. This is especially true of the eskers that lie perpendicular to the direction of former ice-movement. The eskers related to the col channel at the western end of the Tinto ridge (Fig. 2) indicate that this stagnant condition had developed when the ice was still several hundred feet thick in their vicinity. As pointed out above, in order that these eskers could be formed the ice surface must have been at least as high as the col and hence at least 350 to 400 ft. above the lowest parts of the eskers. It may have been considerably higher.

The concept of stagnation has long been applied to thin ice occupying low ground in Scotland, but it does not appear to have been applied to thicker ice in upland areas. In the Southern Uplands the deglaciation has been described by Charlesworth (1926a and b) as taking the form of active (or periodically active) valley glaciers retreating up their valleys towards their sources in the high ground. Within the valleys of the higher parts of the Southern Uplands the former existence of active valley glaciers is demonstrated by steep-sided morainic mounds and ridges, which abound in some of these valleys. Outside these higher areas, however, as J. Geikie (1894) pointed out, clear end-moraines seem to be absent: instead the till covers the valley floors and mantles many of the hill slopes as an extensive sheet with few or no minor constructional forms. Gregory (1927) offered the unacceptable explanation that the smooth sheets of till are due to deposition beneath the sea during a widespread submergence. J. Geikie, however, had stated much earlier that the till sheets were the deposits of an ice-sheet, in contrast with the moraines of the higher hill areas, which were formed by valley glaciers associated with a subsequent return of glacial conditions to a less extensive area. The reason for the absence of end-moraines outside the limits of the valley glaciation seems to have puzzled Geikie, however, as the following quotations from the third edition of his *Great Ice Age* show. On p. 213 he wrote: ' We cannot indeed, escape from the conviction that the last extensive *mer de glace* must eventually have broken up, as it were, into . . . a series of valley glaciers.' Returning to the same topic, he stated on p. 258: ' From all that is known of the phenomena attendant on the decay of the last general ice-sheet, we cannot doubt that the retreat of the valley glaciers into which that ice-sheet was resolved was marked by pauses.' But he continued: ' None of these, however, would appear to have been protracted,' and on the following page wrote: ' The whole aspect of the valleys, and the manner in which morainic and fluvio-glacial detritus are distributed, lead, in short, to the conclusion that the dissolution of the last general *mer de glace* and of its succeeding local glaciers was more or less continuous—and that any pauses or temporary re-advances of the ice were of inconsiderable importance.' These quotations reveal the difficulty that Geikie faced: on the one hand, the concepts of glacial decay held at the time he wrote forced him to conclude that the ice-sheet was resolved into valley glaciers as it decayed and that these glaciers were periodically active. On the other hand, he could find no clear end-moraines to demonstrate this activity.

Since Geikie wrote, the concept of widespread stagnation and

decay *in situ* of an ice-sheet has been developed, for example by Flint (1929) in North America and Mannerfelt (1945) in Scandinavia. There seems to be no reason why this concept should not be applicable to the last ice-sheet to cover the Southern Uplands. This interpretation is opposed to that of Charlesworth (1926a and b), for he has mapped various retreat stages of Southern Upland glaciers deduced from 'moraines.' Beyond limits approximating to Geikie's valley glaciation, however, these 'moraines' appear to be fluvioglacial deposits. One of them, at Thankerton, has been discussed above and shown to be an esker. The so-called kame-moraine of the Eddleston valley, Peeblesshire, has been shown elsewhere to have been formed by streams washing material into dead ice (Sissons 1958). The 'moraine' in the Nith valley north of Dumfries is a mass of fluvioglacial deposits pitted with numerous kettles and partly closed depressions (Stone 1959). Features such as these point to the existence, not of active, but of dead ice.

The absence of end-moraines and the existence of wide spreads of almost featureless till mantling hill slopes accord with the view that the last ice-sheet to cover the Southern Uplands decayed *in situ* without significant movement. The occurrence of eskers winding down slopes and of numerous subglacial channels, some of the latter descending slopes for several hundred feet, is also consistent with the concept of more or less stagnant ice. Even in valleys near the source areas of the ice-sheet one finds such evidence below the limits of the valley glaciation, as, for example, in the Yarrow valley, where eskers lead down to the valley floor almost at right-angles to the direction of former ice-movement.

The glacial drainage evidence of the Tinto area, which has formed the main subject of this paper, is similar to evidence found widely in the Southern Uplands. Such evidence points to the existence of well-developed subglacial drainage systems and suggests that the last ice-sheet to cover the Southern Uplands eventually became more or less stagnant and decayed *in situ* over wide areas.

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The Central and Eastern Parts of the Lammermuir-Stranraer Moraine

By J. B. Sissons

Abstract

The Lammermuir-Stranraer moraine was first described by Charlesworth and interpreted by him as marking an important readvance. The central and eastern parts of this feature are discussed. It is concluded that it is not a true moraine in these areas and that the readvance supposed to have formed it did not occur.

I. Introduction

In 1926 J. K. Charlesworth described a kame-moraine extending along much of the northern edge of the Southern Uplands from St. Abb's Head in the east to Stranraer in the west. He named this feature the Lammermuir-Stranraer moraine and interpreted it as marking the limit of a major readvance of Highland ice. In another paper, published at the same time, he correlated a readvance identified in the western part of the Southern Uplands (the Kirkcowan stage) with the kame-moraine. The term Lammermuir-Stranraer moraine has also come to be applied to the combined limits of Southern Upland and Highland ice at this stage.

The moraine has been subsequently accepted by most writers concerned with glacial retreat stages in Scotland (and in Britain as a whole) as an important stage and has been variously correlated with stages in continental Europe. It appears, for example, in maps by Wills (1951) and Synge (1952) of the major retreat stages in Britain. Manley (1951) discussed the climate at the time the Lammermuir-Stranraer moraine was formed. Movius (1942) referred to it in his " Irish Stone Age " while Lacaille (1948) accepted it in his paper on the deglaciation of Scotland in relation to early man's environment. It appears in maps and text in works on pollen analysis such as Godwin's " History of the British Flora " (1956) and Donner's paper on Late-glacial retreat stages in Scotland (1957). Among important foreign books in which it appears are Flint's " Glacial and Pleistocene Geology " (1957) and Woldstedt's " Das Eiszeitalter " (1958). Recently Charlesworth has repeated his interpretation in summary form in " The Quaternary Era " (1957, pp. 1204–5).

In view of such authoritative and widespread acceptance of the Lammermuir-Stranraer stage one almost hesitates to point out that the conclusions of almost all local workers other than Charlesworth are opposed to the existence of such a stage. Some of the evidence on which these conclusions are based was published before the appearance of Charlesworth's paper on the Lammermuir-Stranraer readvance and in that paper he discussed this evidence and interpreted it differently.

Nevertheless, subsequent work has often demonstrated the validity of the conclusions of the early workers.

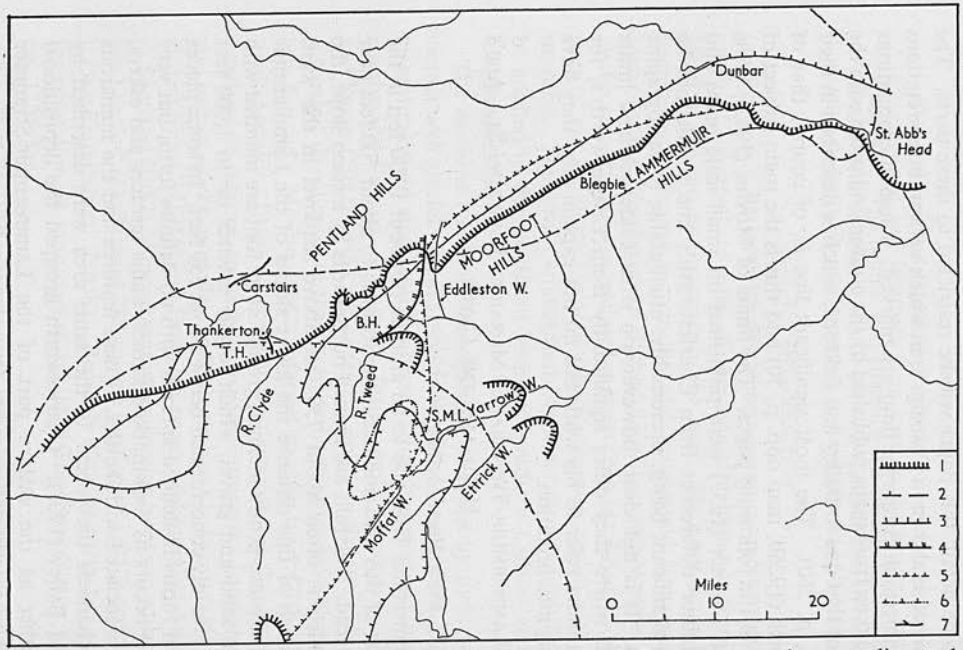
Since Charlesworth (1926b) described the moraine as best developed in its central and eastern parts, and since the present writer has studied these in some detail, this paper will be restricted to these parts. The area concerned is shown in Text-fig. 1, in which various interpretations of the Lammermuir-Stranraer limit are marked. These interpretations have been taken from maps published by the authors indicated with the exception of that due to Bailey and Eckford, which is described in their text (1956, p. 260). The most significant line is, of course, that of Charlesworth (1926b, map opp. p. 50) and this is the main subject of discussion in the following pages. The limits of Movius (1942), Wills (1951), and Donner (1957) were published on small-scale maps and therefore minor differences from Charlesworth's line should not be considered significant being, presumably, attributable to cartographic limitations. It is not clear, however, on what evidence these limits were based where they differ significantly from Charlesworth's, for all three authors refer to his work but do not explain why they have modified his interpretation.

II. The Lammermuir-Stranraer Moraine Between St. Abb's Head and the Clyde

(*a*) *From St. Abb's Head to Blegbie*

The Lammermuir Hills rise to an altitude of about 1,700 feet in the western part of this area and decline eastwards to about 700 feet near St. Abb's Head. The hills are limited northwards by a steep slope, the base of which is often about 750 feet above sea-level in the west. Through much of this distance the lower slopes of the Lammermuir scarp and the more-gently-inclined ground at its foot are mantled with fluvioglacial sand and gravel, which locally extends up to 1,000 feet or so, but is mainly concentrated below about 750 feet. In some places this material forms ill-defined spreads with no definite form, but very frequently it occurs as well-marked kames, kame terraces and eskers, sometimes associated with kettles. These features and the numerous meltwater channels that occur in the same area were described by Kendall and Bailey (1908). Charlesworth grouped the depositional forms together as the eastern part of the Lammermuir-Stranraer moraine. This interpretation meets several difficulties, including the following.

(i) The moraine is almost entirely composed of fluvioglacial material. Where till does occur in the moraine it appears to be restricted to small patches on ice-contact slopes and small areas of ablation moraine of limited thickness. (The morainic mounds with blocks of Old Red Sandstone referred to by Kendall and Bailey (1908, p. 19) are actually



TEXT-FIG. 1.—The eastern and central parts of the Lammermuir-Stranraer moraine according to 1, Charlesworth (1926b, map opp. p. 50); 2, Movius (1942, map p. 52); 3, Wills (1951, map p. 62); 4, Bailey and Eckford (1956, p. 260); 5, Donner (1957, map p. 223). 6, Generalized limits of the valley glaciation in the area between the Yarrow and upper Tweed, partly after Young (1864). 7, The Carstairs and Thankerton "kames". T.H. = Tinto Hills. B.H. = Broughton Heights. S.M.L. = St. Mary's Loch.

an esker, the gravels of which are in places covered by up to 2 feet of angular ablation moraine.) No constructional mounds composed largely or entirely of till have yet been found. This evidence does not support the interpretation of the features as an end-moraine formed by a major readvance.

(ii) The view that the belt of fluvioglacial features is a moraine implies that these features are of approximately the same age throughout the length of the belt. Detailed mapping and interpretation of these features (and of the meltwater channels) shows, however, that they were not formed contemporaneously, but are older in the east than the west (Sissons, 1958b, p. 73). The ice-margin lay at a slight angle with the fluvioglacial belt during the deposition of the latter, as is shown, for example, by well-marked kame terraces alternating with channels in aligned sequence.

(iii) Although Charlesworth stated that the Lammermuir-Stranraer moraine " marks the limit of a readvance of uncertain yet probably considerable magnitude " (1926b, p. 44), he also stated that this moraine lies to the north of the line of maximum readvance in the Lammermuir area (1926b, p. 28). This seems a rather serious anomaly in his interpretation. His conclusions were based on Kendall and Bailey's (1908) valid evidence that the meltwater channels imply that the last ice to occupy East Lothian reached an altitude of over 1,000 feet on the western part of the Lammermuir scarp, but that the main concentration of fluvioglacial deposits is around the scarp foot, especially below about 750 feet. This concentration owes its existence and location to the concentration of meltwater flow in the depression between the scarp face and the ice mass wasting back from it. Into this depression flowed streams from the Lammermuir scarp and from the wasting ice. At times the marginal zone of the ice was buried beneath the fluvioglacial deposits resulting in the kame-and-kettle topography that is locally well developed. The influence of the pre-existing relief on deposition is shown, not only by the abundance of the deposits along the scarp foot for much of its length, but also by the local concentration of material in pre-existing valleys, at the base of steep slopes and at points where streams descending from the Lammermuirs were checked in their rapid progress by the ice mass and caused to deposit their heavy loads. The abundance of fluvioglacial deposits in this area in fact seems to suggest, not a readvance of the ice, but a period of rapid glacier decay.

It is therefore concluded that the fluvioglacial belt following the foot of the Lammermuir scarp is not a moraine and is not related to a halt or readvance of the ice-margin.

Whether or not the higher line marking the upper limit of definite glacial retreat features on the Lammermuir scarp is significant has yet to be ascertained. This line falls gently eastwards from an altitude of

about 1,200 feet in the western part of the area now being considered. It is tempting to conclude that the last major readvance of Highland ice failed to surmount the Lammermuirs and that the upper limit of definite retreat features marks its maximum extent. Hollingworth (1952), however, has pointed out that conclusions of this type may not be valid, stating that " It is generally accepted that marginal drainage will only operate below the snow line ". On the Lammermuir scarp additional factors preventing the formation of channels and related deposits on the upper slopes might have been the steepness of these slopes and the small available drainage areas of streams from the ice-free area. The problem might be solved by studying solifluction deposits, for it is a reasonable expectation, especially in an upland area such as the Lammermuirs, that such deposits would be strongly developed outside the limit of a readvance, but not at the surface within it (having been destroyed by the readvance or buried by its deposits). Such studies have not been carried out in the Lammermuirs and hence the significance of the upper limit of definite retreat features remains uncertain.

(*b*) *From Blegbie to the Eddleston Valley*

For some 17 miles from Blegbie farm to the Eddleston valley the Lammermuir-Moorfoot plateau (*c.* 1,200–2,000 feet) ends in a nearly-straight scarp facing north-west. The ground near the scarp foot is usually an area of subdued relief at between 800 and 1,000 feet altitude. Abundant fluvioglacial deposits occur in this latter area, in places forming complex areas of kames, eskers, and kettles. These deposits and forms continue southwards into the Eddleston valley. All these deposits were considered by Charlesworth to form part of the Lammermuir-Stranraer moraine produced when the margin of the Highland ice lay along the foot of the Lammermuir-Moorfoot scarp with a tongue extending southwards down the Eddleston valley (Text-fig. 1).

Among the difficulties encountered by this interpretation are some already mentioned in discussing area (a), namely, the absence of till moraines and the abundance of fluvioglacial material, evidence favouring rapid deglaciation rather than a period of readvance. A more serious difficulty is met with in this area, however, and will be considered first in relation to the Eddleston valley.

In this valley there is clear evidence that the last glacier ice came, not from the highlands, but from the Southern Uplands. For example, some 200 meltwater channels mapped by the writer in the area drained by the Eddleston Water (1958a, map p. 162) show that meltwaters flowed in a general south–north direction on both sides of the valley, turning north-eastwards to the north of the Moorfoot scarp. The inclination of the channels as a group reflects the direction of slope of

the ice surface and indicates that in the Eddleston valley area the ice came from the Southern Uplands.

Although striae are rare in this area, the only two marked on Geological Survey One-Inch sheet 24 indicate ice movement from south-west to north-east. Further evidence is provided by several eskers, up to a mile long, that occur immediately north-east of the northern part of the Eddleston valley. It is generally accepted that eskers (excluding Mannerfelt's (1945) subglacially-engorged type) often indicate the latest direction of former ice movement. The eskers near the Eddleston valley were formed by streams flowing south-west to north-east and thus imply that the last ice movement was in this direction.

Another line of evidence is provided by indicator stones. In fluvio-glacial gravels in the Eddleston valley felsite from the Tinto Hills, which lie 20 miles to the south-west, has been found (Bailey and Eckford, 1956, p. 255). Furthermore, these gravels are free from coal fragments. The valley is cut in Ordovician and Silurian rocks, but Carboniferous rocks outcrop extensively immediately to the north and north-west of its head. Had the gravels been laid down by streams draining Highland ice they could hardly have failed to contain coal fragments. That the deposits were laid down by streams draining Southern Upland ice is also demonstrated by their bedding, which shows that these streams flowed northwards along the valley side or into the ice towards the valley axis.

There is thus ample evidence that the last ice to occupy the Eddleston valley came from the Southern Uplands. How far to the north this ice extended is not definitely established, but McCall and Goodlet's (1952) study of indicator stones strongly suggests that it reached at least as far as a point 8 miles due north of the head of the valley. One would, therefore expect that it would have also extended along the Moorfoot-Lammermuir scarp-foot zone for a considerable distance north-east of the Eddleston valley. The meltwater evidence favours this view.

In a zone up to several miles wide to the north-west of the scarp a considerable proportion of the meltwater features appear to have been formed subglacially. Proof of subglacial drainage is provided by a number of eskers, the most common direction of the longer ones being south-west to north-east. Several of these eskers can be traced into meltwater channels, strongly suggesting that these channels are of subglacial origin also. There are also several channels with up-and-down long profiles probably formed by meltwaters flowing subglacially under hydrostatic pressure. Throughout this area there is a marked contrast with the corresponding belt at the foot of the Lammermuir scarp north-east of Blegbie. In the latter area kame terraces and channels running along north- and north-west-facing slopes are numerous and

point to an important marginal and submarginal drainage developed in relation to the wasting Highland ice. In the Eddleston-Blegbie belt, however, kame terraces and channels running along slopes are relatively infrequent and most slopes facing north and north-west have no such features at all. On the other hand, several channels occur on slopes facing south and south-east. Furthermore, the south-eastern edge of the main gravel concentration is often some distance from the scarp foot, unlike the gravel belt in area (a), thus again suggesting a difference in origin.

On the scarp slope itself there is an even greater contrast between the parts north-east and south-west of Blegbie, In the former area a great number of gently-inclined channels and kame terraces run along the middle and lower parts of the slope. South-west of Blegbie, however, over a distance of 14 miles there are neither kame terraces nor channels along the scarp slope.

These marked contrasts can be explained by the former presence of Southern Upland ice in the area south-west of Blegbie. The absence of marginal and submarginal features on the scarp slope and on many other slopes facing north-west and north is then accounted for by the ice surface having been inclined away from the scarp (probably at a slight angle). The drainage of this ice mass was partly subglacial, as shown by evidence referred to above. The clear preservation of the eskers with their numerous minor bends shows that, at least in the later stages of deglaciation, the ice was stagnant. Stagnation is also evidenced by the numerous irregular masses of sand and gravel, often bounded by ice-contact slopes. It was these and other fluvioglacial deposits (including the eskers) that Charlesworth grouped together as a portion of the Lammermuir-Stranraer moraine formed along the margin of the Highland ice.

(*c*) *The area between the Pentland Hills and the Broughton Heights*

In the relatively low ground between these two upland areas are numerous fluvioglacial deposits with typical ice-contact forms included in the Lammermuir-Stranraer moraine by Charlesworth. Eckford (1952) has shown that these deposits were laid down in association with Southern Upland ice. As in area (b), the form and distribution of the deposits are indicative of dead ice.

(*d*) *The middle Clyde area*

In this area are two important linear accumulations of fluvioglacial material. The more northerly and larger is well known as the " Carstairs kame " and the other is referred to in the literature as the " Thankerton kame ". Charlesworth (1926b) interpreted each of these " kames " as an end-moraine formed along the margin of the Highland ice.

The Thankerton " kame " has been recently discussed in detail (Sissons, 1961) and will be considered only briefly therefore. It consists of several steep-sided fluvioglacial ridges situated in the valley of the Clyde east of the Tinto Hills (Text-fig. 1). Although differing as regards some aspects of its origin, Charlesworth (1926b) and other writers have all agreed that these ridges were formed partly or wholly along an ice-margin. There is clear evidence, however, that all the ridges are eskers. The principal evidence is as follows. (i) The ridges have steep ice-contact slopes on both sides and are in places margined by kettles, showing that ice was present on both sides of them when they were formed. (ii) The ridges have typical esker form with winding courses, and undulating crest lines, the highest points often occurring at bends. (iii) Two of the principal ridges can be traced back into valleys leading down from the Tinto Hills and the third main ridge can be followed back into a system of meltwater channels and other fluvioglacial features running along the northern side of these hills. (iv) The high proportion of felsite in the ridges shows that they were formed by streams coming from the felsite mass of the Tinto Hills, not from Highland ice that had come up the Clyde valley. The Thankerton ridges therefore do not indicate the former position of the margin of any ice mass. That the last ice to occupy the region in which they occur came from the Southern Uplands is indicated by striae (Geikie, 1873, p. 42), the distribution of Tinto felsite (McCall and Goodlet, 1952) and the numerous meltwater channels and fluvioglacial deposits of the surrounding region.

The Carstairs " kame " (Text-fig. 1) comprises a system of steep-sided fluvioglacial ridges, in places over 50 feet high, with a general south-west to north-east trend. The main ridge system extends over a distance of about 4½ miles and varies in width from a few tens of yards to about half a mile. The ridges split and rejoin one another but generally trend approximately parallel with the system as a whole. Between the ridges are peat-filled, often-closed depressions. South of the ridge system is an area of broad mounds, often with smooth slopes of gentle to moderate inclination. This latter area is dominantly sandy whereas the ridges comprise material ranging from sand to boulders over 2 feet across.

The composition and bedding of the mounds indicate deposition in a lake or lakes, as concluded by Gregory (1915, p. 473), Charlesworth (1926b) and Macgregor (1927, p. 498). The moundy form of these deposits and the irregular, sometimes-closed depressions between indicate deposition in association with dead ice.

The ridge system has been the subject of some dispute in the literature. Charlesworth (1926b) concluded that it was deposited by streams draining Highland ice. The majority opinion, however, favours an

origin in relation to Southern Upland ice (e.g. Geikie, 1894, p. 187 ; Gregory, 1915; Macgregor, 1927). There seems to be no reason for adopting Charlesworth's minority interpretation: he himself referred (1926b, p. 35) to the " apparently overwhelming evidence " against it. Particularly significant is the fact that almost all the material either has been derived or could have been derived from a southerly or south-westerly source. For example, Gregory (1915, p. 471) found in one section 30 per cent of greywackes and other Southern Upland sedimentary rocks but only 2 per cent of Highland rocks. In order to overcome this difficulty Charlesworth suggested various possibilities, the principal one being that the presence of Southern Upland erratics is the result of Highland ice having advanced over ground previously covered by Southern Upland ice. It seems much more likely, however, that the high proportion of Southern Upland erratics is due to Southern Upland ice having advanced over ground earlier occupied by Highland ice. Further east Eckford (1952) and McCall and Goodlet (1952) have shown that this in fact occurred (sections reveal Southern Upland deposits overlying Highland ones) and a similar sequence of events is therefore expectable in the Carstairs area.

Charlesworth and Gregory both concluded that the Carstairs ridge system was formed along an ice margin by streams draining from the ice. This conclusion, however, cannot be reconciled with the form of the feature. Streams draining from an ice mass would have produced either alluvial fans sloping away from the ice or flat-topped deltas. Yet the Carstairs ridges have narrow, undulating crest-lines and are steep on both sides. These steep slopes and the numerous associated kettles show that ice lay on both sides of each ridge when it was formed. This was recognized by Macgregor, for he suggested (1927, p. 498) that the ridges " may have been laid down between ice-walls ". Macgregor also suggested, however, that the ridges were laid down between dead Southern Upland ice to the north and active Southern Upland ice to the south. This interpretation seems rather improbable: in fact the numerous kettles and the frequent bends in the well-defined ice-contact slopes of the ridges indicate that the ice on both sides of the ridge system was stagnant. Macgregor's interpretation also implies that the system was formed in an open ice-walled channel. This is probably partly true, but at least in the north-eastern part a subglacial origin seems more probable. Here there is one main ridge for two miles with a smaller ridge (joined to it at both ends) over a third of this distance. These two ridges, with their steep sides, undulating crests and winding courses, appear to be typical subglacially-formed eskers. Referring to the Carstairs ridges as a whole, Gregory (1915, p. 473) remarked how they " strikingly resemble some typical Swedish eskers " and one can also compare them with some of the Canadian eskers with their

numerous sub-parallel ridges and associated steep-sided kettles. The Carstairs ridges are, in fact, a complex esker and, like the Thankerton esker system further south, form part of the extensive glacial drainage system of the Southern Upland ice sheet, which, over a large area extending from the borders of East Lothian southwards to the source of the Tweed and westwards beyond the Tinto Hills, trends towards a direction between north-east and north.

Thus the concept of a Lammermuir-Stranraer moraine formed along the margin of Highland ice finds no support in the middle Clyde area. The two sets of features in this area interpreted by Charlesworth as end-moraines formed by this ice are both eskers (or esker systems) formed in association with Southern Upland ice.

III. The Lammermuir-Stranraer Stage in the Tweed Basin

As mentioned above, Charlesworth (1926a, 1926b) described and mapped the limits of the Southern Upland ice at the Lammermuir-Stranraer stage. In the vicinity of the Solway he showed this limit as a continuous line, but in the Tweed basin he indicated only the termini of three valley glaciers (1926b, map opp. p. 50). In his text he stated that these glaciers occupied the valleys of the Yarrow and Ettrick Waters and of the upper Tweed. He said that at the same time corrie glaciers existed in the Lammermuirs. In none of these areas, however, are there moraines or other evidence to support this interpretation.

Movius (1942) interpreted the Scottish Readvance described by Trotter (1929, pp. 600–602) in the Carlisle area, Charlesworth's Lammermuir-Stranraer Readvance, and the Perth Readvance of Simpson (1933) as representing a single stage. Other writers, however have considered these three to be distinct and successive stages. Movius's reasons for differing so radically from other interpretations are not given. Furthermore, he stated (p. 53): " Actually the limit of this ice from the Southern Uplands has not been traced north of Brampton in Cumberland . . .". The line shown on his map does not correspond with any evidence in the Southern Uplands indicating the limits of a readvance.

A novel feature of a map included by Wills in his Palaeogeographic Atlas (1951, p. 62) is the delimitation of a nunatak area of considerable extent at the Lammermuir-Stranraer stage in the Southern Uplands. The present writer is not aware of any evidence that suggests the former existence of such a nunatak. Wills' Lammermuir-Stranraer limits in the Southern Uplands correspond approximately with moraines only in part of the area between the Yarrow Water and upper Tweed. These moraines are referred to below.

In a fairly recent paper Bailey and Eckford (1956) suggested a modi-

fication of Charlesworth's interpretation in the Eddleston valley and adjacent parts of the Tweed valley. They concluded that the edge of the Southern Upland ice lay along the western slope of the Eddleston valley and continued south-westwards into the Tweed valley (Text-fig. 1). This interpretation, which has been recently republished in summary form (1960) is opposed to that of the present writer (1958a). The main evidence used by Bailey and Eckford is the fluvioglacial deposits. In the Eddleston valley these deposits are concentrated on the western side (in the area between their ice-margin shown in Text-fig. 1 and the Eddleston Water). Bailey and Eckford interpreted the deposits as formed by proglacial streams that flowed from the ice down the western side of the Eddleston valley. The forms of these deposits show conclusively, however, that they were laid down in intimate association with dead ice, for they occur as kames, eskers, kame terraces and kettles. These forms and their bedding show that the deposits were washed into the ice, not out of it (Sissons, 1958a). This is also demonstrated by about 100 meltwater channels that occur on the valley side immediately above the deposits. These channels, none of which was mentioned by Bailey and Eckford, mostly slope down northwards or towards the axis of the valley and could have been cut only when ice occupied the valley. The relations of the channels to the deposits show that the streams that cut the former also laid down the latter. In the Tweed valley also the evidence of meltwater channels and fluvioglacial deposits is not in accord with Bailey and Eckford's interpretation (R. J. Price, unpublished).

The most recent interpretation of the position of the Southern Upland ice-margin at the Lammermuir-Stranraer stage is shown in a map by Donner (1957, p. 223). In the Eddleston valley his line corresponds in part with Bailey and Eckford's line already discussed (Text-fig. 1). Further south it crosses the Tweed at an altitude of about 600 feet, rises to about 2,000 feet on the Tweed/Yarrow watershed and then descends to 800 feet near St. Mary's Loch. Such an ice-margin position seems very improbable and study of the area reveals no evidence to justify it. Southwards, however, Donner's line does correspond in part with a series of readvance moraines. Some of these moraines, which are composed mainly of till, often with large angular blocks, were described by Young in 1864. The area in which moraines occur is shown in generalized form in Text-fig. 1. As Young clearly demonstrated, these moraines were produced by local valley glaciers (it is not possible to show these individually in Text-fig. 1) and this group of moraines is restricted to the valleys of the high hill area between the Yarrow Water and upper Tweed. Similar moraine areas, not shown in Text-fig. 1, occur further west in the Southern Uplands in and around areas of high ground. These moraines, however, represent a valley glaciation

(or glaciations) of limited extent and cannot be integrated to form part of an extensive ice-sheet as implied by Donner's map.

IV. Conclusion

It has been shown that the central and eastern parts of the Lammermuir-Stranraer moraine are composed almost everywhere of fluvioglacial deposits formed in association with dead ice. Over a considerable length of the northern edge of the Southern Uplands the last glacier ice came from the Southern Uplands, not from the Highlands as Charlesworth maintained. The only true moraines occur in the higher parts of the Southern Uplands, but these represent a local valley glaciation or glaciations. Only in the Lammermuir area did the last ice come from the Highlands and it is not yet known if the limits of a readvance of this ice were reached in this area.

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A re-interpretation of the Literature on Late-glacial Shorelines in Scotland with Particular reference to the Forth Area

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Abstract

Some problems revealed by the literature on Scottish late-glacial raised beaches are indicated and it is suggested that these problems cannot be resolved in terms of the widely-held concept of so-called 100-ft. and 50-ft. beaches. Reasons are given for believing that some late-glacial shorelines in Scotland have considerably steeper gradients than has hitherto been recognised. Particular reference is made to the Forth area, where shoreline altitudes as recorded in the literature reveal anomalies that appear not to be explicable in terms of current ideas. The relation of shorelines to glacial stages is briefly considered and modification of the limit of Simpson's Perth Readvance in the Forth area suggested. Some implications of the hypothesis presented are mentioned.

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I. The Orthodox Interpretation

It has been accepted since before the beginning of this century that there are several distinct raised beaches in Scotland, although

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the interpretation that prevails at present was criticised as long ago as 1906 by Jamieson. A 100-ft. beach is repeatedly referred to in the literature and one at 50 ft. or thereabouts is quite frequently mentioned. A 25-ft. beach is widely recognised and in some areas a lower beach, the 15-ft., is also identified. The two higher beaches are interpreted as of late-glacial age and the two lower as of post-glacial age.

It has long been recognised that these beaches are warped owing to unequal isostatic uplift. The 100-ft. beach reaches an altitude of 145 ft. in the Stirling area according to Dinham and Haldane (1932, p. 212) and 135 ft. according to Read (1959, p. 60). Donner (1959) suggested that the area of maximum isostatic uplift of this beach is probably around Callander. The beach is described as sloping outwards from this area and is believed to descend to sea-level in Caithness and to be absent from the Orkneys, Shetlands and Outer Hebrides (*e.g.* Wright 1937, map p. 369). It is considered to slope southwards in Kintyre and the Firth of Clyde (*e.g.* Donner 1959) and towards the east or south-east in the Forth area (*e.g.* Lacaille 1954, map p. 40).

The evidence for the 50-ft. beach is rather indefinite and areal limits for it have not been described in the literature. One of the few areas in which it has been clearly located is Kintyre and Loch Fyne, where it has been described by Donner (1959). The 50-ft. beach is generally considered to mark a halt in the emergence following the formation of the 100-ft. beach.

The 25-ft. beach has a wider distribution than that claimed for the 100-ft. beach. For example, it has been traced along the Lancashire coast (Gresswell 1953, 1958) and is reported in North Wales (Stephens 1957, p. 144). Gresswell has measured its shoreline as 17 ft. O.D. in S.W. Lancashire. It is said to reach 40 ft. or slightly more in the area between Loch Linnhe and the Firth of Clyde (McCallien 1937, p. 198) and 45 to 49 ft. in the Stirling-Lake of Menteith area (Dinham and Haldane 1932). The 25-ft. beach is separated from the two earlier beaches by a period of lower sea-level indicated by peats, tree remains, etc. near or below present sea-level, as Jamieson recognised nearly a century ago (Jamieson 1865).

The 15-ft. beach is less widely mapped than the 25-ft. beach. It is believed to mark a halt in the emergence from the 25-ft. submergence.

This pattern of sea-level change is conveniently summarised in a diagram by Donner (1959, p. 20), which includes a correlation with the British pollen zones and with carbon 14 dates.

II. Discussion of the Orthodox Interpretation

In the following pages the terms '100-ft. beach', '25-ft. beach', etc. will be used as they are normally used in the literature. Thus the term '25-ft. beach' will be used for the feature that is said to be at about 40 ft. in the Loch Linnhe area and 17 ft. in S.W. Lancashire regardless of its local altitude. Altitudes will be expressed in terms of Ordnance Datum.

The 100-ft. beach is remarkable in that it is usually not sharply delimited at its inner margin. This fact is referred to in many of the Geological Survey memoirs and Wright in his summary (1937, p. 372) referred to 'its extremely feeble development, except in certain districts'. Where it is well developed this is usually due to special factors such as contemporaneous supplies of fluvioglacial material or exposure to strong waves, as in parts of western Scotland. In view of the widespread nature of this submergence the normally feeble development of its shoreline seems rather strange, for one might reasonably expect such a major event to have resulted in a marked feature. One should also add that in some places the upper limit of the so-called 100-ft. sea is marked by a thinning out of the marine sands and clays rather than by a distinct erosional feature.

It is also interesting to note that many of the arctic or semi-arctic shell-bearing deposits associated in the literature with the 100-ft. sea are far below this altitude. In part this can be explained by marine clays having been laid down on the floors of drowned river valleys and also by the fact that a considerable number of sections have been revealed by marine erosion along the present shoreline or by excavations near sea-level. Yet often it cannot be proved that the low-level shell-bearing deposits accumulated when the sea-level was at 100 ft. This was recognised by Bailey in his discussion of an arctic shell bed by Loch Sween in Argyllshire. He said that this low-lying deposit, which is covered in part by deposits of the 25ft. beach, has, in its upper part, 'a distinctly littoral facies'. Bailey went on to say: 'It seems likely that the shell bed is of the age of the 100-foot beach, but the connection, as is so often the case, is far from clear' (Bailey 1911, p. 129). This lack of connection between shell-bearing deposits and the 100-ft. sea is often ignored and in many instances there seems to be no reason why the sea at the time such deposits were accumulated should not have been at a quite different level. One instance where this has been suggested is in F.W. Anderson's (1947) discussion of the Paisley clays.

The altitudinal evidence has also, to some extent, been made to accord with the standard beach pattern of 100, 50, 25 and 15 ft. For example, beaches at altitudes between about 40 and 70 ft. have been quite frequently referred to as ' the 50-ft. beach ', while those between about 80 and 130 ft. have often been called ' the 100-ft. beach '. Wright appears to have had such unsatisfactory correlations in mind when he referred to the feature ' which commonly goes by the name of the " 50-foot " beach ' and said that it is not obvious ' that what has been termed the 50-foot beach is everywhere the same shore-line ' (1937, p. 375). Furthermore, in some parts of Scotland there is published evidence that the simple pattern is not applicable. Thus in the Edinburgh area beaches are said to occur at five levels: 125-130, 100, 75, 50 and 25 ft. (Peach *et al.* 1910, p. 335).

Another problem presented by the higher Scottish raised beaches is their rapid disappearance or poor development in the peripheral areas of their occurrence. For example, in the Edinburgh area beaches are described as occurring up to 130 ft., but some 25 miles to the east at Dunbar there is no clear feature above the 25-ft. beach (Clough *et al.* 1910, p. 183). Farther south, however, the Tweed has definite terraces, the highest of which is said to be 60 to 80 ft. above sea-level (Pringle 1935, p. 93). Beaches up to 100 or 110 ft. occur in Ayrshire (Richey *et al.* 1930, p. 342), Arran (Tyrrell 1928, pp. 264-265) and Kintyre (Donner 1959) and beaches above the 25-ft. continue to Stranraer (A. Geikie 1869, p. 25; A. and J. Geikie 1869, p. 15; A. Geikie 1873, p. 24). Yet along the Solway coast of Scotland there are only scattered occurrences of beaches above the 25-ft. This is rather surprising since the 25-ft. and related fluvial material form extensive flat spreads infilling the heads of bays such as Wigtown Bay, so that one might reasonably expect the higher beaches also to be represented by extensive deposits in these protected localities.

Another perturbing aspect of the interpretation of late-glacial beaches is revealed when one compares the recent work of Charlesworth (1955) and Donner (1959). Charlesworth correlated his glacial stage M, equivalent to the Loch Lomond Readvance of Simpson (1933) with the 100-ft. sea. Donner, however, correlated Simpson's Perth Readvance with the 100-ft. sea and the Loch Lomond Readvance with the 50-ft. sea. Charlesworth correlated his next stage, N, with the 50-ft. sea. Such a major difference between recent correlations causes one to have serious doubts about the basis of such correlations.

III. The Amount and Rate of Isostatic Recovery

In order to attempt to resolve some of these problems it is necessary first to obtain an idea of the amount and rate of isostatic uplift in Scotland. The full amount of uplift cannot be deduced since part of it occurred before raised beaches were formed in Scotland. Only a minimal figure can be obtained and this varies with distance from the centre of uplift. Let us therefore consider a part of the Forth region, the area around Stirling. It might seem that, because the highest beach here reaches about 140 ft. (145 ft. according to Dinham 1927, and Dinham and Haldane 1932, and 135 ft. according to Read 1959) that this is the amount of post-beach isostatic uplift. However, one has to take into account the fact that when this and other high beaches in Scotland were being formed world sea-level was much lower than now. The 140-ft. beach at Stirling was probably formed at the time of Simpson's Perth Readvance (see below), which is shown by Donner (1958) to have occurred at the end of the Zone I period of the pollen zonation. The end of this zone is fixed at about 12,000 B.P. (before present) by carbon 14 datings. At this time, according to Fairbridge's recent detailed synthesis (Fairbridge 1961, p. 155), world sea-level was probably about 160 ft. below its present level. There has thus been about 300 ft. of isostatic uplift at Stirling since the 140-ft. beach was formed.

In both Scandinavia and North America it has been shown that when the earlier beaches were being formed isostatic uplift was rapid, but that later it slowed down considerably (Charlesworth 1957, p. 1327). It is a reasonable presumption that the same pattern of uplift prevailed in Scotland. This can be demonstrated for the Stirling area as follows. Simpson (1933) showed that his second major readvance, the Loch Lomond Readvance, extended to just beyond Lake of Menteith. Here the moraine rests on marine clay at 65 ft. O.D. During the retreat surface moraine was deposited down to 50 ft. O.D., so that the sea could not have been above this latter altitude near Lake of Menteith at that time (Simpson 1933, p. 644). Ignoring the relatively small difference in amount of isostatic recovery between Lake of Menteith and Stirling, an uplift of at least 90 ft. is thus indicated for the Stirling area between the times the ice began to retreat from the Perth Readvance maximum and from the Loch Lomond Readvance maximum. Donner (1958) has shown that the latter readvance occurred during the Zone III period. The end of this period is well fixed by carbon datings as about 10,300 B.P. At this time

Fairbridge (1961) places the world sea-level at about 130 ft. below its present level. It had thus risen about 30 ft. since the Perth Readvance. Thus in the Stirling area about 120 ft. of isostatic uplift had taken place in about 1,700 years. This gives an average rate of isostatic recovery of about 7 ft. a century.

The next datable event is the formation of the 25-ft. beach, which is represented by a feature at 45 ft. at Stirling (Dinham 1927; Dinham and Haldane 1932). The age of this beach appears to vary at different places in Britain (Wright 1937, p. 423; Stephens 1957, pp. 144-145) and it is not yet accurately dated in the Forth area. However, lack of knowledge of the precise age of the feature does not materially affect the present approximate determination of rate of uplift and we may therefore take 5,300 B.P., the mean of Fairbridge's dates (1961, pp. 161-163) for the Middle Littorina transgression, as a convenient date. The world sea-level at this time in stable areas was some 10 to 15 ft. above its present level (Fairbridge 1961, pp. 161-163) and had therefore risen about 140-145 ft. since the ice began to retreat from the Loch Lomond maximum. Thus the average rate of isostatic recovery in the Stirling area in the 5,000 years preceding the formation of the 45-ft. beach was slightly under 3 ft. a century.

Since the 45-ft. beach was formed the amount of isostatic recovery in the Stirling area has been about 25 ft. This figure is less than the altitude of the beach because allowance has to be made for a eustatic fall of sea-level of about 10-15 ft. since the beach was formed and for the fact that the altitude of the beach is referred to Ordnance Datum. The average rate of isostatic uplift at Stirling since the beach at 45 ft. was formed therefore appears to be slightly less than half a foot a century.

There is considerable doubt about the nature of present uplift in Scotland. Valentin (1953) deduced from tide-gauge records that Scotland is still rising isostatically. On the other hand, Hafemann (1954) concluded that isostatic recovery ceased about 1900 years ago. Comparison of the first and second geodetic levellings of the Ordnance Survey indicates that Scotland was still rising very recently, however (Valentin 1953, p. 304), and the newly-completed third geodetic levelling reveals the same trend (L. J. Harris, personal communication). However, in detail, the pattern of uplift revealed by levelling is complex and is not explicable solely by isostatic recovery.

Although uncertainty exists concerning the nature of present movements, a graph can be constructed to show the general pattern of uplift at Stirling over the last 12,000 years (Fig. 1).

Even though the graph can be regarded as only a first approximation and will certainly need modification when more accurate evidence becomes available, there can be no doubt about the rapid rate of uplift in late-glacial times and the relatively slow rate since the 25-ft. beach was formed. It is a reasonable assumption that the

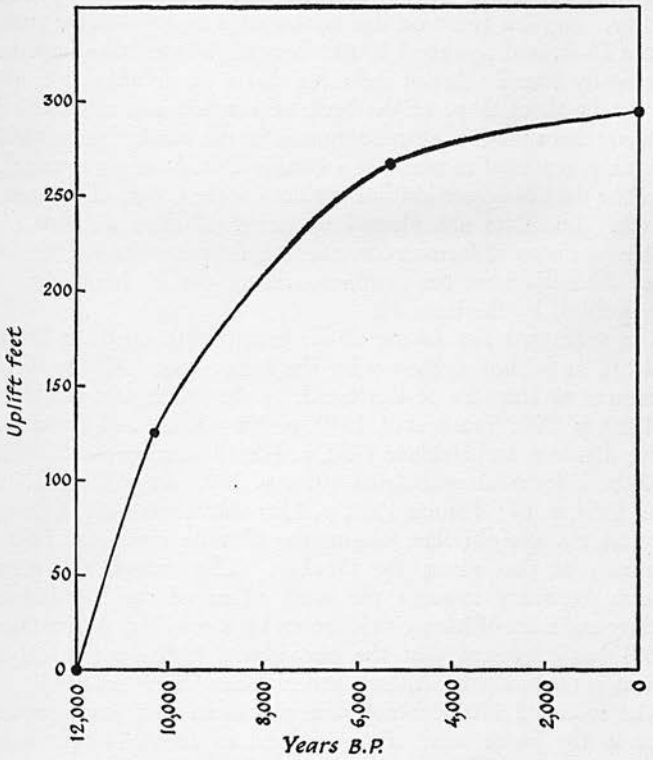


FIG. 1. Isostatic uplift at Stirling since 12,000 B.P.

uplift curve had a similar form elsewhere in Scotland. One implication of these data is that the older beaches should be more steeply inclined than the younger ones. Thus a beach formed at the time of the Perth Readvance should have a steeper slope than one formed at the time of the Loch Lomond Readvance. This in turn should have a steeper slope than one formed in post-glacial times. With these considerations in mind one may now examine some of the evidence at present available regarding raised beaches in the Forth area.

IV. The Highest and Lowest Shorelines of the Forth Area

Along the shores of the Firth of Forth and up the river to the Stirling area a considerable amount of information on raised beaches is available, particularly in memoirs of the Geological Survey. Fig. 2 is based on this information and shows the altitude of the 25-ft. and so-called 100-ft. beaches where these have been specifically stated. A dot indicates that a single altitude is given for the break of slope at the back of a beach and a vertical line indicates that a range of altitude is given for the break of slope. Where a beach is reported to occur at a certain altitude along a stretch of coastline the dot or vertical line is placed at the centre of that coastal stretch. Localities are plotted in terms of their distance from Callander since this forms a convenient reference point for the Forth area. Shorelines on the northern side of the Firth of Forth are distinguished by the letter N.

The westward rise of the 25-ft. beach from 20 ft. at Dunbar to 45 ft. at Stirling is shown by the lower line. All the specific references to altitudes of this beach in the Forth area (Clough *et al.* 1910, p. 184; Peach *et al.* 1910, p. 336; Allan and Knox 1932, p. 76; Dinham and Haldane 1932, p. 212; Macgregor and Haldane 1933, p. 104; Allan and Knox 1934, p. 190; Knox 1954, p. 120; Read 1959, p. 62; Francis 1961, p. 132; Morrison 1961) fall on or very near a straight line joining the altitude given for Lake of Menteith to that given for Dunbar. This reflects the greater isostatic recovery towards the west. Part of the line between Stirling and Lake of Menteith is shown broken in Fig. 2 since Durno (1958) has suggested that the post-glacial transgression did not extend as far inland as Dinham and Haldane (1932) believed.

The so-called 100-ft. beach does not show a constant eastward slope in the Forth area. From a level of about 140 ft. around Stirling it soon declines to about 120-130 ft., but then remains at this altitude along the north shore of the Firth of Forth to Fife Ness, the easternmost point of Fife (Haldane and Allan 1931, p. 140; Dinham and Haldane 1932, p. 212; Macgregor and Haldane 1933, p. 104; Allan and Knox 1934, p. 189; Read 1959, p. 60; Francis 1961, p. 132; Morrison 1961). On the south side of the Forth a shoreline at 120-130 ft. is described in the Edinburgh area (Peach *et al.* 1910, pp. 335-336). Farther east on this side of the Forth reliable information concerning this feature is not available.

This feature at 140 to 120-130 ft. cannot be reconciled with the slope of the 25-ft. beach in terms of present concepts. The higher

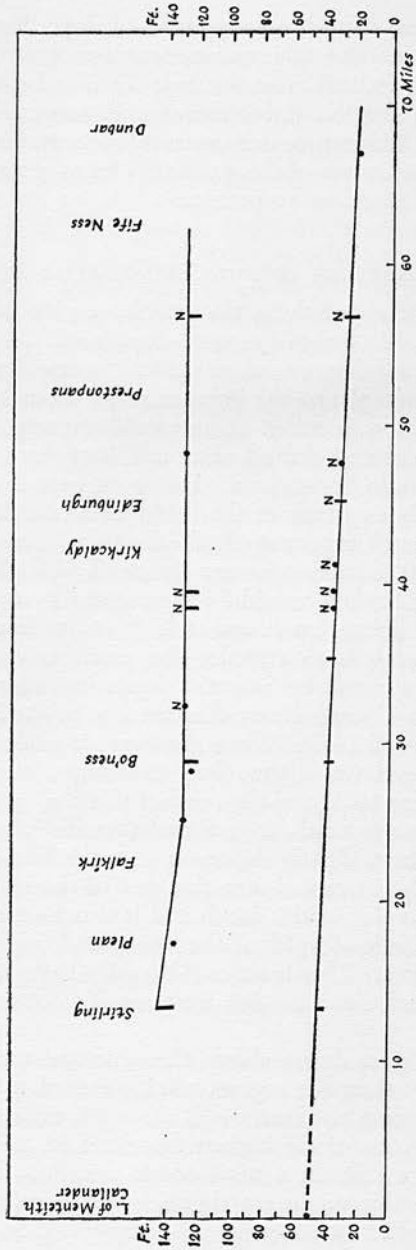


FIG. 2. The slopes of the 25-ft. and so-called 100ft. beaches in the Forth area.

feature should have a much greater eastward slope than the lower one, for a large amount of isostatic recovery occurred between the times the two features were formed. One is therefore forced to the conclusion that the higher shoreline shown in Fig. 2 is not a true shoreline. This can be demonstrated in several ways, which also provide an indication of the gradients one might expect some of the late-glacial shorelines to possess.

V. Possible Gradients of some Late-Glacial Shorelines

One method of approaching this problem is by analogy with Scandinavia. F. M. Synge (personal communication) finds the maximum slope of late-glacial shorelines in southern Finland, measured at right-angles to the isobases, to be about 1·7 metres a kilometre (about 9 ft. a mile). One would not expect slopes of this gradient in eastern Scotland, since in this country the ice was much thinner than in Scandinavia. However, even if the Scottish ice-sheet was only a quarter of the thickness of the Scandinavian ice-sheet, one could expect to find late-glacial shorelines with slopes up to 2¼ ft. a mile in eastern Scotland. On this basis the 140-ft. feature at Stirling would be represented by a shoreline at 65-70 ft. in the Edinburgh area and at 15 ft. at Dunbar.

Another approach is to consider the probable extent of the area in Britain affected by isostatic depression and subsequent recovery. In the Liverpool area the shoreline of the 25-ft. beach is at 17 ft. (Gresswell 1953). When allowance is made for the fact that world sea-level was higher than now when this beach was formed (Fairbridge 1961) and for the fact that the figure of 17 ft. relates to high-water mark, it is found that the Liverpool area appears to have been slightly depressed since the formation of the 25-ft. beach. The Liverpool area thus lies in the vicinity of the limit of isostatic uplift of this beach and it is not unreasonable to assume that the limit of uplift of the late-glacial beaches was also in this general area. This limit could hardly have been situated much farther south since the last ice-sheet did not extend much farther south.

Around Stirling, as shown above, there has been some 300 ft. of isostatic recovery since the highest marine incursion there. The distance from Stirling to Liverpool is about 200 miles. Therefore the average gradients of the highest shorelines are unlikely to be less than 1½ ft. a mile in a north-south direction from central Scotland. Even this minimum figure is incompatible with the concept of a 100-ft. beach as generally accepted for Scotland. For

example, a shoreline at 100 ft. at the mouth of the Clyde (*say* Gourock) would be contemporaneous with a sea-level 15 ft. below the present level at Stranraer. If the same gradient of uplift applies in an eastward direction down the Forth the Stirling 140 ft. beach would equate with a sea-level at about 90 ft. at Edinburgh and about 55 ft. at Dunbar.

It seems very probable, however, that the gradient of uplift would be considerably greater down the Forth than in a north-south direction from Stirling. It is generally accepted that the last ice-sheet reached mid-Shropshire on the west side of England but extended no farther south than mid-Yorkshire in the Vale of York. This reflects the fact that the main ice-accumulation regions were in the west of Britain, for which reason one would not expect the Scottish ice to have extended very far into the present site of the North Sea. This view is favoured by the presence in Buchan of an area of 800 square miles, much of which is low-lying ground, that remained ice-free during the last glaciation (Synge 1956, pp. 132 and 139–140). This view is also represented in various published maps of the limits of the last ice-sheet (*e.g.* maps reproduced in Charlesworth 1957, pp. 957 and 1073). The distance eastwards from Stirling to the limit of the isostatically-affected area was therefore probably considerably less than 200 miles and the late-glacial shoreline gradients of the Forth consequently steeper than $1\frac{1}{2}$ feet a mile.

The distribution of the 25-ft. beach suggests a similar conclusion. Although this feature has been identified in north Wales (Stephens 1957, p. 144) it seems not to have been traced into England on the eastern side of Britain. It is possible, however, that the feature extends some distance south of Berwick and has yet to be mapped. Even if this is the case there is still a great difference between its southern limits on the eastern and western sides of Britain. This in turn implies that the limit of the isostatically-uplifted area lies much farther north on the eastern side of Britain. This would result in a closer spacing of the isobases eastwards from Stirling than southwards and hence in shoreline gradients greater than $1\frac{1}{2}$ ft. a mile.

An indication of the possible eastward gradient of some of the late-glacial beaches in the Forth area may be obtained by considering Wright's zero isobase for the 25-ft. beach (1937, p. 369). This lies out in the North Sea about 130 miles east of Stirling. Assuming this isobase is roughly correct and that it corresponds with the limit of uplift for the 140-ft. beach at Stirling, an average eastward slope of about $2\frac{1}{3}$ ft. a mile for the Stirling feature is indicated.

This steeper gradient in an eastward direction than in a southward one accords with the slope of the 25-ft. beach, which descends more rapidly down the Forth than it does in a southward direction along the coast of south-west Scotland and north-west England. A gradient of $2\frac{1}{3}$ ft. a mile for the Stirling feature would result in a contemporaneous sea-level of about 65 ft. at Edinburgh and 10 ft. at Dunbar.

These calculations of late-glacial beach gradients are clearly only rough approximations. Refinements have not been attempted since the available evidence does not warrant them. But, unless the data used are completely invalid, the conclusion cannot be avoided that the late-glacial shorelines have much steeper slopes than has yet been recognised, and that the so-called 100-ft. and 50-ft. beaches, so widely accepted, cannot exist.

VI. The Relation of the So-called 100-ft. Beach to Ice-Margins

It has been stated in the literature that the so-called 100-ft. beach was formed in immediate proximity to glacier ice in many parts of Scotland. For example, in East Lothian it has been concluded that ice existed in the Aberlady-Tynemouth depression when the beach was being formed in the adjacent coastal areas (Clough *et al.* 1910, pp. 181-182). Near St. Andrews Rice (1960 pp. 121-122) has correlated a sea-level at about 100-120 ft. with glacier decay in that area. Near Perth Simpson (1933, pp. 639-640) correlated his Perth Readvance with a sea-level at about 90 ft. Synge (1956, p. 140) correlated the contemporaneous Aberdeen Readvance with a sea-level at 80 ft. or more. Synge also associated the decay of the Strathmore glacier with a sea-level at about 100 ft. Around Elgin fluvioglacial erosional and depositional features indicate a sea-level at about 100-120 ft. when the ice wasted away there. On the central Fife coast, near Buckhaven, it is reported that the inner margin of the beach at 120-130 ft. is obscured by fluvioglacial sand and gravel, thus suggesting the proximity of glacier ice when the beach was being formed (Allan and Knox 1933, p. 76). Farther up the Forth in the vicinity of Stirling the highest beach merges into outwash (Dinham 1927). Along the west coast of Scotland and on the coasts of some of the inner islands the so-called 100-ft. beach has been correlated with ice-margin positions at numerous points (*e.g.* Charlesworth 1955; Wright 1937). If this beach is of the same age everywhere, as is frequently implied, one is forced to conclude that the ice-margin was at all the points

referred to (and many others not mentioned) at the same time. This is clearly not possible. The work of Simpson (1933) and Synge (1956) in fact shows that the so-called 100-ft. beach is not everywhere the same age for they have demonstrated that the sea-levels at about 90 and 80 ft. at Perth and Aberdeen respectively are related to a glacial readvance.

These difficulties of relating the so-called 100-ft. beach to glacier retreats and readvances are largely overcome when it is realised that the beach is a time-transgressive feature. During the long period of glacier decay in Scotland, interrupted by occasional readvances, the sea flooded into coastal areas as the ice vacated them. The upper limit of marine submergence over a considerable part of Scotland is the so-called 100-ft. beach. Owing to differential uplift its age varies along its length so that it is progressively younger towards the centre of the isostatically-affected area. The upper limit of the feature is often vague because it does not mark a major submergence of long duration, but merely the temporary maximum level of the sea at any point. In a few places the level of maximum submergence is well marked, but this is due usually to storm-ridges, which can be formed in a very short time, or to abundant fluvioglacial material.

VII. A Possible Late-Glacial Shoreline in the Forth Area

The preceding discussion explains why the upper ' beach ' line in the Forth area shown in Fig. 2 does not slope more steeply than the lower one. It also shows that true late-glacial shorelines should intersect the upper and lower lines of Fig. 2 at distinct angles. The evidence in the literature is not sufficiently detailed to establish the altitudes of these late-glacial shorelines, except possibly in one instance. In eastern Fife alone the coastal stretch between St. Andrews and Kincraig Point Morrison (1961) has recognised a shoreline at 90-100 ft., which he shows to be quite distinct from the maximum submergence limit of 120-130 ft. Morrison measured the altitude of the 90-100 ft. shoreline at 22 points. Some 7 or 8 miles west of Kincraig Point in the East Wemyss-Buckhaven area Allan and Knox stated that ' a distinct notch is cut in boulder clay at a level of about 112 ft. above O.D.' (1932, p. 76). This feature was later described (Knox 1954, p. 119) for a rather larger area, including the East Wemyss-Buckhaven coast and continuing south-westwards almost to Kirkcaldy, as forming a nick at 110-120 ft. Along the northern coast of the Forth southwest from Kirkcaldy a shoreline at 110-120 ft. is not mentioned in Francis's recent

description (1961), but one at 120-130 ft. is described. Taken together this evidence suggests the existence of a shoreline that slopes down from 120-130 ft. in the area southwest of Kirkcaldy to 90-100 ft. in the area east of Kincraig Point, a suggestion that is strengthened by the fact that in the accounts cited no reference is made to shorelines at similar altitudes that might permit different correlations.

On the south side of the Forth the evidence available at present is insufficient to permit even tentative correlations to be made. It is perhaps significant, however, that in the Edinburgh area it has been stated that a feature exists west of the city at approximately 100 ft., but that the '75- and 50-ft. terraces' occur only east of Granton (Clough *et al.* 1910, pp. 335-336).

VIII. The Upper Marine Limit and the Perth Readvance in the Stirling Area

One aspect of Fig. 2 that has not so far been referred to is the termination of the high sea-level in the Stirling area. This is described by Dinham (1927, pp. 488-489) as follows: 'while the 100-ft. beach, when followed up the valley of the Carron, merges into the high-level stream terraces of that river, nothing similar occurs at Stirling; the platform comes to a sudden end and cannot be traced beyond Cambusbarron and Bridge of Allan.' Such a marked drop in the upper limit of marine submergence is normally found to occur behind a moraine or other evidence indicating an important halt in glacial retreat or a readvance of the ice. Thus Simpson (1933, pp. 640-641) concluded that a lowering of sea-level of at least 40 ft. occurred when the ice was at the Perth maximum and immediately afterwards. Synge (1956, p. 134) recorded a drop of sea-level of about 75 ft. at this time in the Aberdeen area. Such marked falls of sea-level are due partly to isostatic uplift of the land and partly to a temporary reversal of the eustatic rise in sea-level caused by temporary growth of glaciers in the world as a whole.

One might suggest that the narrow gap in which Stirling lies, itself partly blocked by rocky hills, would result in a retreating ice-margin stabilising itself at this position for a considerable time, during which there was a marked fall of sea-level. This seems to have been Dinham's view, for he suggested that a considerable fall of sea-level occurred 'at this stage in the retreat of the Forth glacier' (1927, p. 489). Thus he did not envisage a readvance of the ice. It seems very probable, however, that such a marked descent of the upper marine limit as occurs in the Stirling area is the result of an important readvance. It also seems very likely

that such a readvance would be equivalent to the Perth Readvance of Simpson, especially as nowhere else in the Forth area below Stirling is there evidence of a fall of sea-level of the type that occurred at the time of the Perth Readvance.

Simpson did not suggest this correlation, presumably because he believed the ice in the Forth valley extended much farther down the valley than Stirling at the time of the Perth Readvance. One of Simpson's reasons for this view was the existence of meltwater channels at Glenfarg and Newburgh that carried waters across the Ochils from south to north (1933, p. 638). From this he deduced that the ice stood higher on the southern side of the Ochils than on the northern side and, consequently, that it must have extended much farther east than did the ice in the Tay valley at the same period. It is suggested that this reasoning is not relevant, for there appears to be no evidence that the two meltwater channels referred to were cut at the time of the Perth Readvance.

Simpson's other reason for taking the Perth Readvance far beyond Stirling was that ice from the west flowed up Strath Allan. He argued that this could have occurred only if the region east and south of Stirling was occupied by a still more powerful mass of glacier ice. The writer believes that the flow of ice up Strath Allan is not a serious problem and that the limited width of the gap at Stirling, coupled with the fact that the ice terminus in the Stirling area was situated in the sea and thus subject to melting by sea-water (and possibly to calving), was sufficient to stabilise the ice-margin of the Perth Readvance in the Stirling area.

IX. Some Implications of the Preceding Discussion

If the conclusions reached above concerning late-glacial shorelines are valid it follows that the terms '100-ft. beach' and '50-ft. beach' as normally used in the literature should be abandoned. The gradients tentatively suggested for some of the late-glacial shorelines also suggest some revision of the current concepts of phases of emergence and submergence may be required. For example, at the time that raised beaches were being formed in some coastal areas sea-level in others may have been lower than it is now. The evidence of the marine deposits with arctic or semi-arctic fauna needs to be reconsidered and the fact that so many of these deposits occur at low levels now has an additional explanation. Furthermore, it has to be recognised that the crossing of shorelines of different ages may occur. At any point the gradients of older shorelines will be steeper than those of younger ones, considerable differences in gradient being expectable between the

late-glacial and post-glacial shorelines. Finally, it has to be remembered that the so-called 100-ft. and 50-ft. beaches have been used to correlate glacial stages in different parts of Scotland (Charlesworth 1955; Donner 1958, 1959). Charlesworth (1955, p. 904) stated that, for correlation purposes ' the value of these lines cannot be over-emphasised.' The validity of such correlations is therefore very doubtful.

X. Conclusion

The type of interpretation outlined in the preceding pages is not new, for such interpretations have long prevailed in Scandinavia. It may be that the pattern revealed by detailed study in Scandinavia is quite different from that which will emerge in Scotland. The present writer does not think so, but only research more detailed than that undertaken hitherto in Scotland will reveal the truth.

XI. Acknowledgement

The writer wishes to acknowledge his indebtedness to Mr. F. M. Synge for a valuable discussion of the shorelines of southern Finland.

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SCOTTISH RAISED SHORELINE HEIGHTS WITH PARTICULAR REFERENCE TO THE FORTH VALLEY

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Introduction

It has long been known that during and since the decay of the last ice sheet the level of the sea has been higher than at present around the coast of much of Scotland for considerable periods. This is shown especially by the raised beaches.

If these raised beaches were always clearly and continuously developed and were everywhere horizontal it would be a simple task to determine the former levels of the sea. However, the beaches are often found only at intervals along the coast and are sometimes imperfectly preserved. In addition, warping has occurred owing to the differential uplift of Scotland as it recovered from its depression beneath the weight of the last ice-sheet. A problem that these beaches present, therefore, is that of correlating the various remnants that survive.

To many workers in this field this problem does not exist: it has long been solved. For the last eighty years or so the Scottish raised beaches have been grouped as the "100-ft.", "50-ft.", and "25-ft." beaches. This three-fold sequence, sometimes with a beach at about 15 ft. added, has received, and still receives, wide acceptance. For example, it has formed the basis of three accounts published very recently (Mitchell et al., 1962, pp. 123—125; Earp, Francis and Read, 1962; Donner, 1963).

Surprisingly, although this orthodox interpretation refers to the beaches in terms of their altitude, few accurate height measurements have been published. In some published accounts the heights assigned to beaches appear to be estimates related to map contours, and this is the basis of the heights of 100, 50 and 25 ft. assigned to the three main features. In some other accounts reference is made to instrumental heighting, usually by aneroid barometer, sextant or hand level, the datum plane usually being high-water mark or the barnacle line. Although such measurements can be of considerable value in a provisonal study (e. g. Bailey et al., 1924) it has to be recognised that neither the datum plane nor instrument is absolutely reliable and errors of several feet may be present in the heights assigned to shorelines. For example, Donner, in his recent (1963) study of Scottish raised beaches, mentions discrepancies of up to 2 metres between aneroid readings at the same point, and in some areas was unable to locate the barnacle line to which most of his heights are related. The only fully satisfactory method of determining raised shoreline heights is by precise levelling based on Ordnance Survey bench-marks. Yet the literature contains no more than a few dozen heights determined in this manner in the whole of Scotland. In brief, therefore, the Scottish raised beach sequence as currently accepted is largely based on unsatisfactory height information.

Methods

In view of these uncertainties the writer has commenced a detailed study of the raised beaches of a limited area: the south side of the Forth between Dunbar and the western limit of the carse-lands near Gartmore. This work is based on mapping on the scale of six inches to the mile of all identifiable raised beaches and related features such as outwash plains and meltwater channels. Heights are being obtained by precise levelling based on Ordnance Survey bench-marks. It is hoped to integrate this evidence with the stratigraphy, palynological studies and the results of radiocarbon dating. Although this work is still far from complete it is hoped that some of the height information so far obtained may be of interest, especially as this indicates the need for some revision of generally-accepted views and is inconsistent with some recently-published evidence.

The heights of raised shorelines given below all relate approximately to the concave break of slope at the back of raised beach deposits. The

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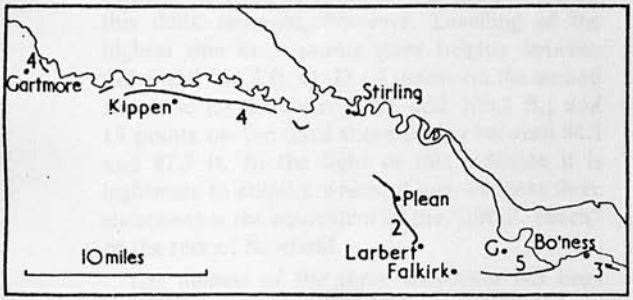


Fig. 1. The western part of the Forth basin showing places referred to in the text and the location of the shorelines and fluvioglacial spreads whose heights are shown in Figures 2 to 5. G = Grangemouth.

precise junction of beach with backslope is almost always obscured by material that has moved down the backslope. For this reason measurements on post-glacial beaches were made sufficiently far on to the beach to avoid the downwashed material, usually between 20 and 80 metres distant from the actual shoreline. The difference between the true heights of a post-glacial shoreline and the heights measured in this way is probably no more than a few inches, however, for the post-glacial beaches of the Forth west of the Grangemouth area (the area considered here) are represented by the extensive carse-lands with slopes that are usually only a few feet a mile in directions perpendicular to the shorelines. Most measurements on late-glacial shorelines were made in similar positions. In a few instances, however, the late-glacial beach remnants are of limited width or have suffered considerable modification by mass-movement or stream erosion. Measurements on such remnants are therefore less accurate than the remainder and this is reflected in the greater scatter of heights for the late-glacial shoreline shown in Figure 2 as compared with the scatter of measurements for the post-glacial shorelines shown in Figures 4 and 5. Smooth-surfaced deposits of fluvioglacial material were levelled in the same manner as the raised beaches.

Where sufficient height information has been obtained for raised shorelines or fluvioglacial spreads, regression lines have been calculated and are shown on the appropriate diagrams, along with relevant data. It may be noted that all the regression lines have a high degree of significance (better than .001).

Some measurements on late-glacial shorelines

Jamieson (1906) stated that the concept of the 100-ft. raised beach appears to have been introduced in 1879 when the Geological Survey published a memoir including the Falkirk area (Geikie et al., 1879). North of Falkirk, around Larbert and Stenhousemuir, is an extensive remnant of the raised delta of the River Carron with maximum dimensions of 3 by 2½ kms. This is shown on the latest published Geological Survey maps as one raised beach: the "100-ft.

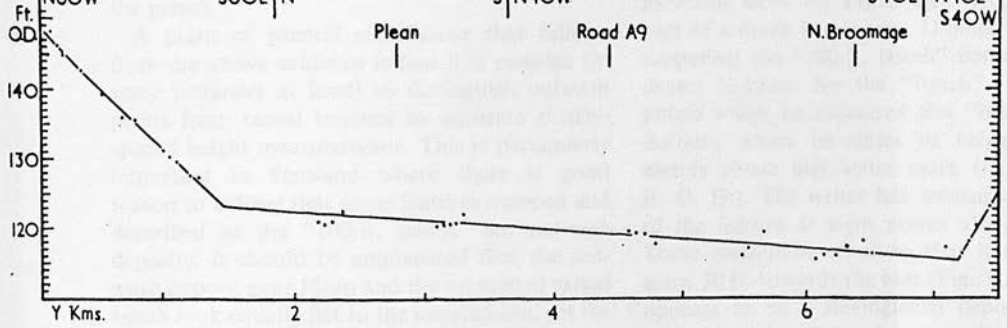


Fig. 2. The highest raised shoreline and related outwash spreads in the Plean-Larbert area. Outwash deposit near Plean: r = — 0.996, significance better than .001 (d. f. n-2 = 10); regression equation, Y = 8.59 ± 0.005—0.06 X; fiducial limits for a probability of 0.99 are b ± 0.015. Raised shoreline: r = — 0.86, significance better than.001 (d. f. n-2 = 23); regression equation, Y = 89.14 ± 0.09—0.71 X; fiducial limits for a probability of 0.99 are b ± 0.25. Outwash deposit near Larbert: line estimated.

beach". At least three shorelines are present on this delta remnant, however. Levelling of the highest one at 9 points gave heights between 115.6 and 118.2 ft. O. D.: 5 points on the second shoreline lay between 107.8 and 109.2 ft.; and 13 points on the third shoreline lay between 84.3 and 87.7 ft. In the light of this evidence it is legitimate to enquire which, if any, of these three shorelines is the equivalent of the "100-ft. beach" of the rest of Scotland.

The highest of the three shorelines has been traced more or less continuously northwards to near Plean. Including the 9 heights mentioned above, a total of 25 heights has been obtained on this feature in a distance of 5 kms. These heights all lie between 115.6 and 122.6 ft., the higher ones occurring towards the north and the lower ones towards the south (Fig. 2). The slope of the shoreline calculated from these data is about 1.4 ft. a km. (0.4 m./km.). North-west of Plean the raised beach merges into an extensive area of deposits which, where not dissected, appears flat to the naked eye. The upper layers of these deposits are much coarser than the raised beach deposits and towards their north-western end are composed of very coarse gravel. Here the deposits are continued by meltwater channels and ice-contact features. This strongly suggests that the apparently-flat area of coarse deposits is a glacial outwash plain. This is confirmed by levelling, which indicates a slope of about 18 ft. a km. (5½ m./km.) towards the south-east (Fig. 2). As indicated in Figure 2, the beach also merges into outwash near Larbert, this outwash having been derived from glacier ice occupying the Carron valley to the west of Larbert (maps and discussion of the glacial evidence of this area will be found in Sissons, in the press).

A point of general significance that follows from the above evidence is that it is possible (in some instances at least) to distinguish outwash plains from raised beaches by accurate closely-spaced height measurements. This is particularly important in Scotland where there is good reason to believe that some features mapped and described as the "100-ft. beach" are outwash deposits. It should be emphasised that the outwash deposit near Plean and the associated raised beach look equally flat to the unaided eye, yet the slope of the one is approximately twelve times that of the other.

Failure to distinguish the two types of feature may help to account for some recent differences of opinion on the late-glacial raised beaches of the middle Forth area. In a brief provisional account the present writer stated that the shoreline rises to approximately 130 ft. near Plean (Sissons, 1962, p. 223). This figure was based on measurements by aneroid barometer and has now been rejected along with more than 300 other aneroid measurements on the Forth raised beaches. Earp, Francis and Read (1962) published a diagram including a shoreline situated between 125 and 135 ft. in the Larbert-Plean area. This diagram cannot be reconciled with the writer's levelled heights of the same feature, none of which exceeds 123 ft. One may suggest that some of the heights given by Earp et al. are on outwash deposits and are therefore too high. Their statement that the raised beach is essentially horizontal is unacceptable for it is clearly inconsistent with the data shown in Figure 2. Yet, as long ago as 1879, in the original description of part of the middle Forth area by the Geological Survey, Peach (in Geikie et al.) clearly described the highest raised beach as sloping and gave estimated heights that closely approximate to those measured by the writer.

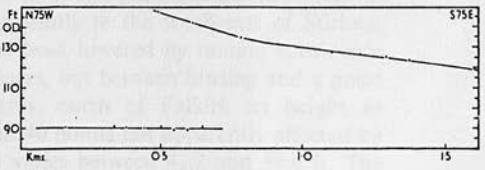


Fig. 3. A fluvioglacial deposit (upper line) and a raised shoreline (lower line) near Bo'ness: lines estimated.

The most recent contribution to the study of the Forth raised beaches is that of Donner (1963), in whose work the Forth area is considered as part of a much larger area. Donner has strongly supported the "100-ft. beach" concept and has drawn isobases for the "beach". One of the points where he measured this "beach" is near Bo'ness, where he states its height to be 38 metres above high-water mark (i.e. about 134 ft. O. D.). The writer has measured the height of the feature at eight points along its length. These measurements show that it slopes down some 30 ft. towards the east (Fig. 3). The feature appears to be a fluvioglacial deposit and was probably laid down by a meltwater stream flowing eastwards along the shallow valley in which the deposit begins. It should be added that about 250 metres to the north of the feature

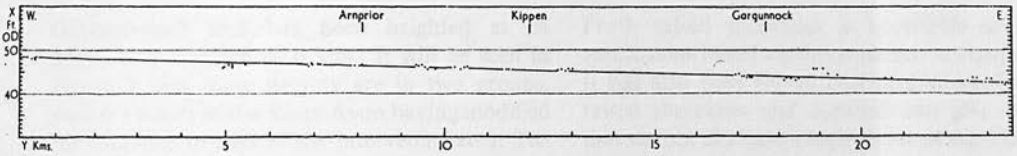


Fig. 4. The highest post-glacial shoreline west of Stirling: r = — 0.92, significance better than .001 (d. f. n-2 = 58); regression equation, Y = 187.39 ± 0.21—3.86 X; fiducial limits for a probability of 0.99 are b ± 0.57.

heighted by Donner is a very clear, 600-metre-long, late-glacial raised beach at about 90 ft. (Fig. 3) not mentioned by him. One therefore hesitates to attach any significance to the isobases for the "100-ft. beach" drawn by Donner for the Forth area.

Some measurements on post-glacial shorelines

The post-glacial raised beaches of the Forth valley westwards from the Grangemouth-Bo'ness area are represented by the extensive carse-lands. Most or all of these carse deposits have been usually referred to as the "25-ft. beach", although almost all authors from Jamieson (1865) to Donner (1963) have recognised that this feature is above 25 ft. and that it slopes eastwards down the Forth valley. The precise heights assigned to the feature have varied in different accounts. However, apparent precision has been recently given to its recorded height by Earp, Francis and Read (1962). They show a diagram of heights for the feature, including 19 heights on the south side of the Forth, all of them between 43 and 50 ft. O. D. In this diagram the shoreline heights are projected on to an east-west line and a band of shading representing the beach is shown as horizontal. In the text they state that the feature "shows virtually no east-to-west slope in a distance of 15 miles [24 kms.] from Plean westwards to Kippen" (1962, p. 217).

The writer's measurements on this highest post-glacial shoreline cannot be reconciled with the evidence or conclusions of Earp, Francis and Read, but they agree with and give precision to the conclusion of most earlier workers that the shoreline is tilted. Between the western end of the carse and Stirling the shoreline trends approximately west-east (Fig. 1). In this 22 km. distance its altitude has been determined by precise levelling at 60 points. In Figure 4 these heights are projected on to a west-east line and the regression line calculated from them is shown. 10 % of the measured heights differ from this line by between 1 and 2 ft. and the remaining 90 % differ from it by less than one foot. The regression line, representing the shoreline, slopes from about 48 ft. at the western end of the carse to about 42 ft. near Stirling, a gradient of about 0.25 ft. a km. (0.08 m./km.). The shoreline continues to slope down gently to the south-east of Stirling. Here it has been lowered by mining subsidence in a few places, but between Stirling and a point about 6 kms. north of Falkirk its height as measured at 40 points not apparently affected by subsidence varies between 42.2 and 39.6 ft. The shoreline continues to descend eastwards from Falkirk at least as far as Dunbar.

A second post-glacial shoreline at about 30 ft. was identified by Read (1959). It has been heighted at only about 20 points so far, most of which lie between 27 and 32 ft. It appears to slope eastwards down the Forth valley.

A third post-glacial shoreline is present near

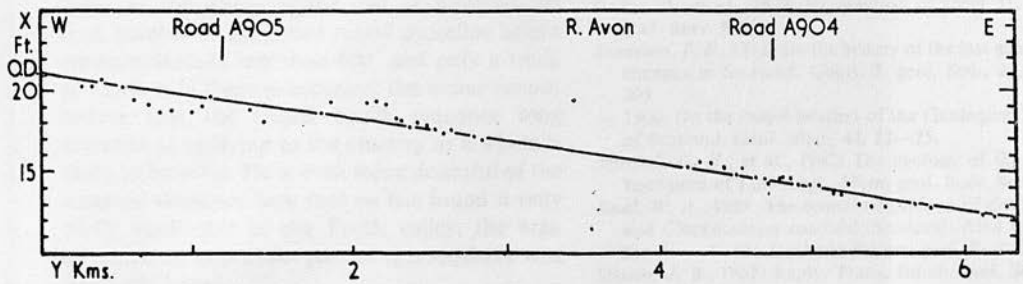


Fig. 5. The lowest apparent raised shoreline near Grangemouth: r = — 0.98, significance better than 001 (d. f. n-2 = 52); regression equation Y = 14.97 ± 0.02—0.71 X; fiducial limits for a probability of 0.99 are b ± 0.05.

Grangemouth and has been heighted at 54 points in a distance of 6 kms. It will be seen in Figure 5 that these heights are in two groups, mainly a result of the River Avon having modified the shoreline in part of the intervening area. The surprising result of these measurements is that they indicate a slope of about 1½ ft. a km. (0.45 m./km.), descending from about 21 ft. to about 12 ft. in the 6 km *s* One might expect that this shoreline, the lowest and therefore presumably the youngest of those so far identified, would have a more gentle slope than any of the higher shorelines, owing to its having been least influenced by isostatic warping. The reason for the anomaly is uncertain at present. The relatively-steep slope of the feature may have been produced by a regular fall of sea-level and/or rise of the land. Alternatively, there may have been several slight oscillations of sea-level that produced several shorelines separated by such small height intervals that it has not been possible to separate them by the method used. Whatever, the cause of this anomaly it serves to emphasise the fact that far less is known about the raised shorelines of Scotland than is generally assumed.

Conclusion

Six raised shorelines—three late-glacial ones and three post-glacial ones—have been mentioned above as existing in the Forth valley westwards from the Grangemouth area. There is, in addition, evidence of other late-glacial shorelines, but this has not been presented since these shorelines have been heighted at only a few points so far. Although the writer has now made over 600 precise height measurements he still does not know the total number of shorelines nor how some of the shoreline remnants that have been measured correlate with others within the area. Since, in the whole of the rest of Scotland the total number of published raised shoreline height measurements is less than 600, and only a small proportion of these is accurate, the writer cannot believe that the raised beach sequence long accepted as applying to the country as a whole is likely to be valid. He is even more doubtful of the accepted sequence now that he has found it only partly applicable in the Forth valley, the area on which an important part of this sequence was originally based.

In this paper it has been shown that the most recently-published height information on the Forth raised shorelines is unreliable and the conclusions based on this evidence unacceptable. It has also been shown that precise levelling of raised shorelines and outwash can give results that statistically have a high degree of significance. It is suggested that the main conclusion to be drawn from this paper is that it is essential that, in the study of Scottish raised shorelines, preconceived notions of shoreline relationships established in the distant past be rejected and that a new start be made based initially on detailed morphological mapping and numerous precise height measurements. When this task has been completed, and the results integrated with stratigraphic and palynological evidence and radiocarbon datings, it is probable that there will be revealed a shoreline sequence very different in some respects from that widely accepted at present.

Acknowledgements

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Appendix

Levelled heights in feet O. D. (with National Grid references) on raised shorelines and fluvioglacial features used in the construction of Figures 2—5. Heights are in order from left to right in relation to each diagram.

Figure 2. South-western margin of outwash plain situated north-west of Plean.

148.2	NS/8190.8831	135.8	NS/8247.8789
146.3	NS/8194.8829	131.2	NS/8268.8784
145.0	NS/8199.8827	130.3	NS/8273.8783
142.9	NS/8209.8813	129.8	NS/8279.8782
139.2	NS/8222.8802	127.9	NS/8288.8776
136.0	NS/8235.8789	127.4	NS/8295.8774

Figure 2. Raised shoreline extending from near Plean to near Larbert.

121.1	NS/8356.8702	116.1	NS/8471.8406
120.8	NS/8359.8696	117.1	NS/8474.8402
121.1	NS/8362.8690	119.5	NS/8484.7993
122.6	NS/8364.8682	115.6	NS/8518.8367
120.2	NS/8355.8619	116.1	NS/8523.8362
120.5	NS/8354.8614	116.0	NS/8530.8356
120.7	NS/8354.8609	117.3	NS/8534.8347
120.8	NS/8354.8605	118.2	NS/8541.8337
122.0	NS/8352.8600	116.4	NS/8546.8332
119.8	NS/8430.8489	117.3	NS/8555.8289
119.4	NS/8342.8483	116.5	NS/8548.8281
119.2	NS/8453.8476	118.2	NS/8540.8280
117.8	NS/8437.8469		

Figure 2. North-western margin of outwash plain near Larbert.

120.7	NS/8534.8278	123.3	NS/8526.8271

Note: the two last heights in the preceding section may belong to this feature.

Figure 3. Fluvioglacial feature near Bo'ness.

147.6	NT/0207.8078	127.5	NT/0267.8064
142.1	NT/0218.8075	125.4	NT/0285.8057
134.6	NT/0235.8073	123.5	NT/0293.8054
128.9	NT/0248.8070	119.1	NT/0312.8041

Figure 3. Raised shoreline near Bo'ness.

90.6	NT/0181.8109	90.7	NT/0214.8100
91.0	NT/0198.8105	89.0	NT/0218.8099
90.2	NT/0202.8104	88.9	NT/0223.9098
90.0	NT/0208.8102	90.4	NT/0228.8097

Figure 4. Raised shoreline from near Gartmore to near Stirling.

48.0	NS/5336.9786	45.1	NS/6587.9529
48.2	NS/5343.9804	45.7	NS/6593.9530
47.9	NS/5345.9777	45.7	NS/6602.9529
48.4	NS/5349.9810	44.1	NS/6997.9476
45.9	NS/5792.9520	44.2	NS/7005.9475
46.0	NS/5797.9516	43.8	NS/7012.9474
46.6	NS/5806.9510	43.9	NS/7019.9474
46.4	NS/5814.9508	43.4	NS/7025.9474
46.0	NS/5819.9511	43.5	NS/7031.9475
46.7	NS/5842.9532	43.5	NS/7036.9476
46.1	NS/5988.9614	42.9	NS/7114.9466
46.9	NS/5993.9606	43.4	NS/7120.9469
46.4	NS/5997.9621	43.2	NS/7130.9469
46.8	NS/6011.9622	42.9	NS/7141.9469
46.8	NS/6014.9612	43.4	NS/7150.9466
46.3	NS/6092.9509	45.5	NS/7187.9455
46.6	NS/6233.9558	45.5	NS/7193.9451
46.3	NS/6240.9562	45.7	NS/7208.9440
46.5	NS/6248.9568	44.3	NS/7220.9435
46.0	NS/6254.9571	44.1	NS/7242.9431
45.6	NS/6262.9572	43.3	NS/7359.9400
45.8	NS/6279.9573	43.5	NS/7372.9398
46.0	NS/6323.9560	42.9	NS/7554.9253
45.7	NS/6339.9557	43.6	NS/7567.9253
46.0	NS/6346.9555	43.4	NS/7576.9252
46.1	NS/6352.9554	43.2	NS/7589.9249
45.5	NS/6444.9552	42.5	NS/7596.9247
45.5	NS/6451.9551	43.2	NS/7610.9242
45.3	NS/6568.9528	42.3	NS/7638.9264
45.4	NS/6576.9527	41.9	NS/7641.9256

Figure 5. Raised shoreline near Grangemouth.

20.3	NS/9130.8013	17.0	NS/9384.7971
20.3	NS/9136.8012	15.4	NS/9518.7993
20.7	NS/9143.8009	15.6	NS/9526.7996
20.3	NS/9156.8002	15.6	NS/9534.7998
19.5	NS/9164.8002	15.3	NS/9543.8002
19.2	NS/9173.8001	14.9	NS/9552.8002
18.7	NS/9180.8000	15.3	NS/9555.8004
18.8	NS/9187.7997	14.6	NS/9564.8007
18.7	NS/9201.7994	14.6	NS/9570.8011
19.1	NS/9207.7994	14.3	NS/9572.8010
19.6	NS/9212.7994	14.6	NS/9579.8014
18.9	NS/9236.7994	14.7	NS/9582.8014
19.1	NS/9245.7993	14.6	NS/9587.8015
18.6	NS/9281.7975	14.2	NS/9592.8016
19.4	NS/9287.7974	14.2	NS/9599.8017
19.3	NS/9310.7970	14.2	NS/9605.8018
19.4	NS/9316.7970	13.7	NS/9615.8018
19.2	NS/9322.7969	13.6	NS/9620.8018
18.4	NS/9328.7970	14.2	NS/9625.8018
18.2	NS/9332.7972	13.0	NS/9632.8020
17.8	NS/9337.7974	12.9	NS/9640.8021
18.1	NS/9342.7977	12.9	NS/9649.8025
17.9	NS/9348.7980	13.0	NS/9668.8032
17.6	NS/9353.7981	12.8	NS/9678.8034
17.4	NS/9359.7980	12.4	NS/9712.8037
17.6	NS/9364.7977	12.2	NS/9720.8037
17.2	NS/9378.7973	12.8	NS/9728.8040

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THE GLACIAL DRAINAGE SYSTEM AROUND CARLOPS, PEEBLESSHIRE

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AROUND the small village of Carlops, situated thirteen miles south-west of Edinburgh by the Midlothian–Peeblesshire boundary, is the most impressive system of meltwater channels in the Edinburgh area. The main purpose of this paper is to attempt to explain the origin and development of this system and of the related fluvioglacial deposits. Brief reference will also be made to the glacial drainage features of the surrounding district.

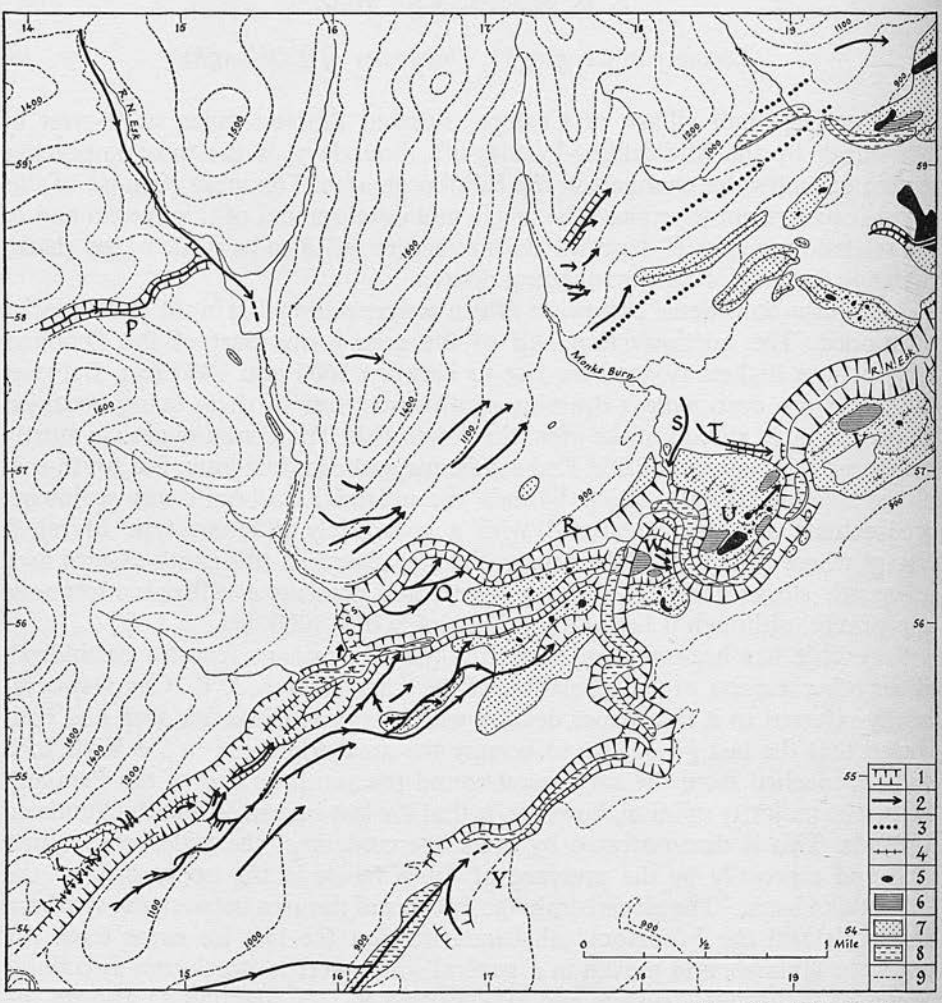
The area considered comprises fifteen square miles lying in the two counties mentioned. The north-western half of the area forms part of the Pentland Hills, whose highest ridges here rise to between 1600 and 1800 feet. Between the ridges are deep valleys draining southwards, one of them being occupied by the principal stream of the area, the North Esk. The Pentland area is mostly composed of lavas and of Old Red sandstone and conglomerate, but south-east of the Pentland Fault, which bounds the uplands, Carboniferous sediments predominate and are associated with a completely different type of relief. Except where dissected by the deep meltwater channels, this south-eastern area has gentle slopes and very shallow valleys and presents a rather monotonous appearance, although it lies at an altitude of 800 to 1000 feet.

Very little has been written about the glacial drainage features or, indeed, about other aspects of the glaciation of the Carlops area. J. K. Charlesworth[1] briefly referred to it in a paper dealing with a very much larger area and concluded that the last glacier ice to occupy the area originated in the Highlands and approached from the south-west round the southern end of the Pentland Hills. The majority opinion, however, is that the last ice came from the Southern Uplands. This is demonstrated by the stone content of the widespread upper drift and especially by the presence of Tinto felsite in the upper drift of the Midlothian basin.[2] The glacial drainage features of the area between the southern Pentlands and the Moorfoots also indicate that the last ice came from the Southern Uplands and moved in a general south-west to north-east direction.[3] Scores of meltwater channels and eskers trend in this direction as also do the few known striae.[4] Within the Carlops area itself this south-west to north-east trend of the glacial drainage features is dominant and further evidence of this direction of ice-movement is provided by several ice-moulded rock and till ridges at the foot of the Pentlands (Fig. 1).

Although R. J. A. Eckford included the meltwater channels south-west of Carlops shown in Figure 2 in a fairly lengthy study of the glacial drainage features of the adjacent West Linton–Dolphinton area, he avoided discussing this system of channels in detail. He said that 'there seem to be impressions of

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more than one ice-melt phase'[5] and concluded that the principal channel functioned as the overflow of a lake situated in the Lyne valley to the south-west. This explanation does not account for the complex system shown in Figure 2.



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FIGURE 1—Glacial drainage features around Carlops.

1. Larger meltwater channel. 2. Smaller meltwater channel. 3. Ice-moulded ridge. 4. Esker. 5. Kettle. 6. Flattish area of fluvioglacial material. 7. Other areas of fluvioglacial material. 8. Alluvium. 9. Contour.

Fieldwork

Fieldwork was preceded by stereoscopic study of aerial photographs on a scale of 1 : 10,000, identifiable forms and drift boundaries being copied on to

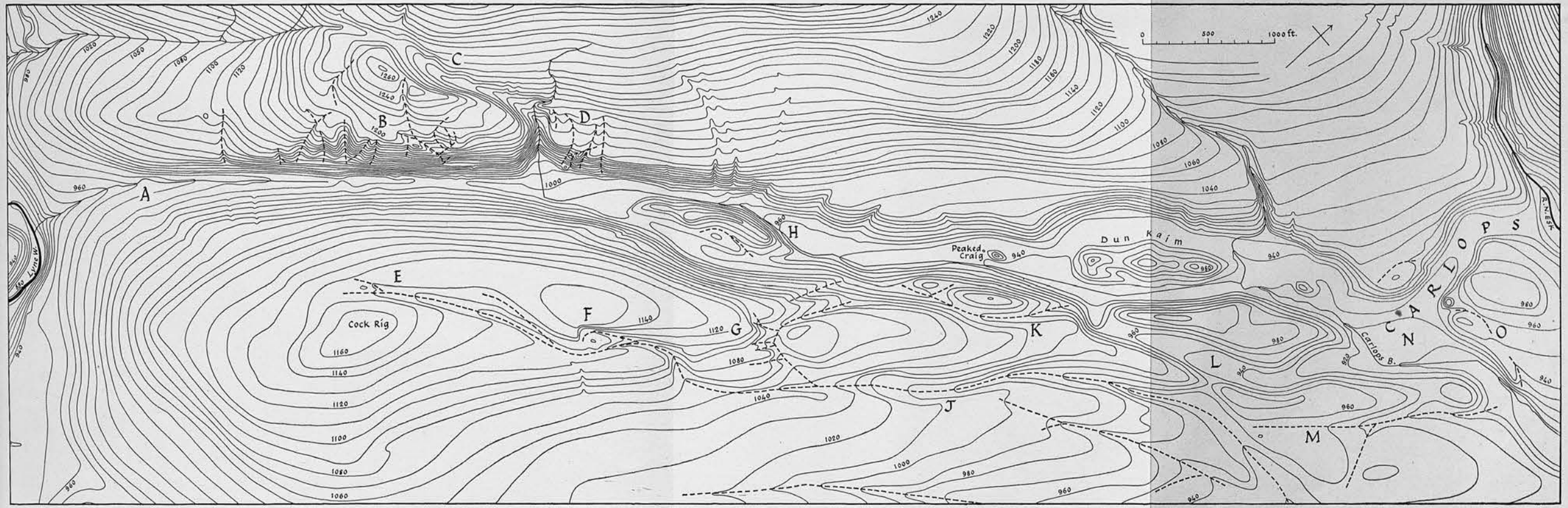


FIGURE 2—The meltwater channels south-west of Carlops.
Smaller channels indicated by broken lines. Contour interval 10 feet.

the 1 : 25,000 Ordnance Survey sheets that were used for the basic field mapping. The more complex parts of the area were also mapped on the six-inch scale. Aneroid measurements were made at a number of significant points and a soil auger was used and sections dug to ascertain the nature of the drift at certain places.

The channel system south-west from Carlops was contoured at 10-foot intervals (Fig. 2) as follows. The six-inch Ordnance Survey map was enlarged to a scale of fifteen inches to the mile and sharp breaks of slope mapped from aerial photographs were added. The altitude of 480 points was determined with an aneroid in traverses between bench marks and spot heights of no more than one and a half hours duration, under stable weather conditions with little or no wind. The aneroid readings were corrected for atmospheric pressure changes during traverses and for temperature, and it is believed that the great majority are no more than three or four feet in error. The reliability of the method was indicated by the close correspondence between the 1000 and 1250-foot contours drawn from the aneroid measurements and these contours as shown on the Ordnance Survey six-inch map.

General Description of the Glacial Drainage System

The principal glacial drainage system runs south-west to north-east across the area shown in Figure 1. It comprises a set of channels, which in many places exceed 60 feet in depth, and includes a belt of fluvioglacial deposits bordering the channel system towards the north-east.

The most complex part of the channel system lies south-west from Carlops. Here most of the channels are aligned south-west to north-east and range from features 100 feet deep to shallow depressions only a foot or two deep. The main channels form a braided system, between the various elements of which are steep-sided ridges. The most remarkable of the 'islands' left between the channels is Peaked Craig (Fig. 2), which forms an isolated sharp-pointed rock knoll about 35 feet high. The larger channels are cut largely or entirely in rock, which comprises Old Red and Carboniferous sediments and lavas, and they often follow faults. The floors of the channels, especially the large ones, are covered with various deposits, including material from their sides, alluvial fans and other fluvial deposits, and peat. The thickness of these deposits is not usually known. Included among the smaller channels of the area is a group trending north-west to south-east that is cut into the north-western side of channel AN in areas B and D, Figure 2.

North-eastwards from Carlops the main channels of Figure 2 are continued by two large channels between 40 and 100 feet deep, the more northerly one being utilized by the North Esk (Fig. 1). A third channel, which continues channel J, Figure 2, is a very faint feature for most of its length but suddenly becomes a steep-sided, V-shaped, rocky gorge in its north-eastern part. All three channels join the 100-foot deep channel YX (Fig. 1) which has a

strongly meandering course. The waters of all these channels continued in a general north-easterly direction along channel UV, which almost everywhere exceeds 100 feet in depth and has a maximum depth of about 175 feet. Although this great trough is occupied by the North Esk, it is clear that this modest-sized stream has not cut it.

North-eastwards from near Carlops a belt of fluvioglacial deposits borders the channels on one or both sides, except that no such deposits are associated with channel YX.

Borehole Evidence

About 800 borehole records were kindly made available to the writer by the Geological Survey and the National Coal Board. Most of these boreholes are located south-east of channel YV where they are often spaced at intervals of approximately 100 feet on a grid pattern and thus provide locally a very detailed picture of the nature of the drift and of the form of the sub-drift surface. Unfortunately, no borehole records at all are available in the vicinity of the channel system south-west of Carlops.

The majority of the boreholes show a drift thickness of less than 10 feet and the maximum thickness recorded is over 68½ feet (base of drift not reached). This bore was put down through the fluvioglacial deposits on the divide between two of the large channels south of point R (Fig. 1). The main feature revealed by the boring operations is that the North Esk channel follows in part the line of a pre-existing depression.

North of the fluvioglacial deposits in bend W (Fig. 1) eight bores indicate till of variable thickness up to a maximum of 31 feet. The bedrock altitude beneath this till is between about 810 and 830 feet O.D., some 40 to 80 feet below the general bedrock level of the adjacent higher ground. Beneath the northern part of the fluvioglacial deposits the bedrock surface rises southwards quite rapidly to about 860 to 870 feet, which is thus the position of the southern side of a buried valley.

In the next bend, U, three bores, situated in the belt of kettles, reveal 45, 50 and 57 feet of drift, comprising fluvioglacial sand and gravel resting on till. The bedrock surface is here about 775 to 790 feet O.D. Immediately south of the lower end of channel T (Fig. 1) the side of the North Esk channel is composed almost entirely of rock and several bores immediately south of channel T show only 5 to 10 feet of drift covering a rock surface at about 850 to 865 feet O.D. The northern side of the buried valley thus lies between points U and T (Fig. 1) and is truncated by the wall of the North Esk channel north-east of point U.

The buried valley continues in bend V, but here the North Esk channel departs from it and is cut largely or entirely in rock. The line of the buried valley is marked at the surface by a broad depression some 30 to 40 feet deep which contains the peat flat in bend V (unshaded area, Fig. 1) and rejoins the North Esk channel near the edge of the map area. Boreholes on the south-

eastern slope of the depression show that the drift increases in thickness north-westwards to a recorded maximum of 30 feet. No borings are available for the axial part of the surface depression. At the south-western end of the depression, where this leaves the North Esk channel, exposures in the side of the latter show that the drift has a maximum thickness of at least 40 feet, but the full thickness is uncertain owing to mass-movement. At the north-eastern end the evidence is clear, for a small tributary of the North Esk has cut down to bedrock and revealed some 30 to 35 feet of till capped by 10 to 15 feet of fluvioglacial material. It thus appears that beneath the surface depression containing the peat flat in bend V is a buried valley with about 30 to 50 feet of drift. This means that the bedrock floor of the valley is at about 750 to 780 feet O.D., giving a depth in bedrock for the feature of roughly 100 feet.

Taken together the evidence from bends W, U and V of the North Esk channel presents a consistent picture. It reveals the existence of a buried valley some 40 to 100 feet deep whose floor level declines in altitude north-eastwards and which was in part utilized by the meltwaters that cut the much deeper North Esk channel. Since the floor of the buried valley is in places covered by a considerable thickness of till its origin antedates the last advance of ice over the area. It may be the old valley of the North Esk or it may be a meltwater channel formed during an earlier glaciation.

In the remainder of the area the borehole evidence is rather scattered and the following points may be briefly noted. In the fluvioglacial belt between the deep channels immediately south of R (Fig. 1) two bores reveal over 60 feet of drift and three others indicate more than 20 feet. It thus appears that the buried valley described above continues south-westwards through this area. In the extreme south of the map area several old boreholes put down through the alluvium that floors the south-western branch of channel YX show that till, as well as sand and gravel, is present. Since this till has a recorded thickness of up to 18 feet it cannot be regarded as ablation moraine let down onto the floor of a subglacial channel when the overlying ice wasted away. Rather does it indicate that the depression was already in existence before small quantities of meltwater utilized it during the last deglaciation. This interpretation is consistent with its subdued form, which contrasts with the sharpness of channel YX north-east of the point at which the side channel from the south-east joins it (Fig. 1). It was through this side channel that much the greater part of the meltwaters that used the deep channel YX flowed during the last deglaciation.

Finally, reference must be made to an old borehole recorded as having been put down on the floor of channel YX immediately west of point X. The channel is here just over 100 feet deep and the bore log records a thickness of 64 feet 8 inches of drift. No other evidence in the area supports this record and it is possible that the site of the borehole has been incorrectly located. The borehole will therefore not be taken into account in the present discussion. If it should be proved correct the sequence of events outlined below would be more complicated.

Origin and Development of the Glacial Drainage System extending North-east from Carlops

The deposits. The first fluvioglacial features to be formed in the area during the last deglaciation were the depositional ones. This is shown by the universal truncation of the deposits by the sides of the deep meltwater channels. If the deposits were restored across the sites of the channels they would form a continuous belt of fluvioglacial material with a general north-eastward decline in surface altitude. Exposures in the sides of the channels and borehole evidence show that the thickness of the deposits varies from about 60 feet down to a thin veneer of a few feet. There are few good sections, but those that do exist show crudely bedded coarse gravel, sometimes with thin layers of sand, in and south-west of bend U. Exposures showing the nature of the material are quite common, however, and indicate that in this area the maximum length of the water-worn stones is usually 6 to 8 inches. North-east of bend U the deposits become distinctly less coarse, and although one section reveals thin beds of gravel with stones up to 8 inches long, the maximum size of the stones is usually between 2 and 4 inches. Furthermore sand is more common and one section (197572) showed that the top 4 feet of one mound is composed entirely of sand.

The surface form of the deposits is usually irregular, comprising mounds and kettles, the largest of the latter (in bend U) containing Marfield Loch, which has a length of 200 yards. The mounds vary in height from about 20 feet down to faint undulations separating kettles 2 or 3 feet deep, as on the inter-channel areas south of R. In a few places the deposits have an approximately flat surface, forming small terraces or flat-topped mounds.

Although at first sight the deposits seem to form a chaotic collection of features, mapping of the individual forms reveals a definite pattern, almost all the mounds trending south-west to north-east. This points to their formation by water flowing along them lengthways, a conclusion consistent with the north-eastward decrease in size of the materials in the north-eastern part of the area. The mounds are thus eskers and the deposits as a whole may be described as an esker system. They were probably formed in part subglacially and in part in open ice-walled channels. The former origin is suggested by the presence in the deposits of angular blocks up to two feet across (for example, at 191569, 179562 and 174560) and the latter by the presence of several small terraces and flat-topped mounds.

The development of the channels. When the erosional phase supervened the various streams that had laid down the esker deposits were replaced, from bend W north-eastwards, by a single major meltwater river that cut the great North Esk channel. The excavation of this channel was in places facilitated in the earlier stages by the presence of the old drift-filled valley, as in bends W and U, but in bend V the eroding meltwaters failed to excavate the buried channel. As already noted, the depression over the site of this buried valley in bend V is

some 30 to 40 feet deep and terminates at both ends in the side of the North Esk channel. It follows, therefore, that at the time the uppermost 30 to 40 feet of the deep rock gorge of bend V were being excavated by meltwaters, the drift-floored depression must have been occupied by a considerable mass of glacier ice.

A further indication of the presence of glacier ice at the side of the North Esk channel during part of its excavation occurs in area STU. Here the Monks Burn flows into the North Esk channel by a V-shaped valley whose lower part is cut in rock. This valley is joined by a streamless flat-floored valley (S) which is cut largely in drift and begins near the sharp bend in the Monks Burn further upstream. Another dry channel (T), cut almost entirely in rock, begins near by and deepens rapidly south-eastwards to join the North Esk channel. That channel T, at least, is not merely an abandoned postglacial course of the Monks Burn is indicated by two aspects of it. First, T is cut about 70 feet into rock at its south-eastern end and thus appears far too large for the small burn to have cut. Secondly, it does not descend to the bottom of the North Esk channel but 'hangs' some 80 feet or so above the channel floor. Channel T was thus cut before the North Esk channel was completely excavated. Furthermore, it is difficult to see why the Monks Burn should have abandoned its fairly straight south-eastward course to the North Esk to turn through a right-angle bend as it now does unless glacier ice were present to block its earlier route. If this reasoning is correct glacier ice must have still bordered the North Esk channel when it had been excavated to about half its present depth.

Along channel YX and at W evidence of the development of the large channel meanders is preserved. The channel wall on the outside of each bend is often steep, but on the inside the slope is normally less. This is most obvious in bend X, where the outer channel wall averages 24 degrees and the inner only 8 degrees. Midway down this latter slope is a marked bench about 50 yards broad. Within the bend to the south a bench also occurs, but at a higher level, and within the bend to the north, but at a lower level, is another bench (too small to be shown in Figure 1). These three benches, being at quite different altitudes, do not relate to a particular stage in the excavation of channel YX, but they do show that the meanders were developed during the excavation of the channel and were not formed by meandering streams superimposed from the ice (as are some channels elsewhere). Farther north, the growth of meander W is recorded by a slip-off slope a quarter of a mile long that declines northwards at an average gradient of approximately one degree. This feature was probably formed quite rapidly as the meltwater river removed the drift fill of the buried valley. The gently inclined slip-off slope is cut off sharply on three sides by the steep wall of the channel proper, indicating that after lateral movement had ceased a period of vertical incision in rock ensued. The slip-off slope has a very faint relief in addition to its general northward gradient, this being indicated by the small features marked on it in Figure 1.

Unlike channel system QV, channel YX is not bordered by fluvioglacial

deposits. This implies that meltwaters did not follow the route of channel YX until after the esker-building phase had terminated. This is in accord with the fact that the channel is aligned at right-angles to the esker belt where it cuts through the latter. For a considerable time channel YX and the three large channels south of point R functioned together, for all are deeply incised, but during this time channel YX carried an ever-increasing proportion of the waters coming into the region until finally it became the only channel that was supplying meltwaters from the West Linton basin (which lies to the south-west).

The evidence in the West Linton basin (such as the extensive thick fluvioglacial deposits around West Linton itself) suggests that channel YV functioned for a considerable time after the other channels in the area covered by Figure 1 had been abandoned and was thus at this time a proglacial channel. It has been shown, however, that, at an earlier stage, the channel was (in part at least) bordered by glacier ice, while still earlier the esker system bordering the North Esk channel was probably partly formed subglacially. It thus appears that one has to visualize a gradual transition from subglacial drainage to open ice-walled drainage and then to proglacial drainage. It seems reasonable to presume that for a considerable time all three types of drainage would have existed. Since the general slope of the ice surface was down towards the north-east (for the ice had come from the south-west), one may therefore visualize the drainage at a given time as being proglacial, open ice-walled and subglacial in successive order from north-east to south-west, the three zones gradually extending south-westwards until the whole system had become proglacial.

The Origin and Development of the Glacial Drainage System South-west of Carlops

The village of Carlops is entirely located in a small part of a complex glacial drainage system. This system is well known to Edinburgh geographers and geologists and has been visited for at least fifty years by excursion groups and individuals interested in the glaciation of the Edinburgh area. No detailed attempt has been made to explain it, however, and only brief references to it occur in the literature. Owing to the complexity of the system a detailed contour map of it was prepared and this is included as Figure 2.

It will be suggested that the channel system was formed subglacially, except that in the final stages parts of it may have been utilized by meltwaters flowing in open trenches in the ice and, subsequently, flowing proglacially, in the manner suggested above for the continuation of the system to the north-east.

The subglacial interpretation. Streams can occupy six possible positions in relation to glacier ice: they can be supraglacial, englacial, subglacial, open ice-walled, marginal or proglacial. In considering the meltwater channels that exist at the present time in formerly glaciated areas one can obviously immediately eliminate as possible agents in their formation supraglacial and englacial

streams since, so long as these streams occupied such positions in the ice, they could not have cut channels in the ground beneath the ice. In the Carlops area a proglacial explanation (except perhaps in the final stages for parts of the system) can be immediately excluded also. The channel system is almost at right-angles to the Lyne and North Esk rivers situated in the Pentlands respectively at its south-western and north-eastern ends. The system as a whole is thus explicable only by glacial diversion; and likewise, at numerous points within it, individual channels or parts of channels can only be accounted for if glacier ice existed in their immediate vicinity when they were being formed.

Such relationships have often been used in the past as evidence of a marginal origin of meltwater channels and in the present area Eckford[6] appears to have envisaged this origin for part of the system under consideration. Many features of the system cannot be explained by this hypothesis, however. For example, the steep-sided rock ridges in the middle of channel AN would require highly improbable ice-margin oscillations to account for them. Several channels, including E, G, K, L and M are situated in part or in whole at higher levels than channels to the north-west of them, so that they cannot be explained in relation to the margin of ice wasting continuously back towards the lower ground. It would be necessary to postulate complex and improbable oscillations of the ice-margin, for which there is no other evidence such as end-moraines. Furthermore, some channels, such as K and the two small parallel channels east of H require the presence of ice on their north-west sides to allow them to be formed. Still other channels, such as those at G and channel E commence on the highest part of a broad divide. For these and other reasons the marginal interpretation proves quite inadequate to explain the channels of the area shown in Figure 2.

It has been mentioned briefly above that in the final stages of formation of some of the channels the meltwaters may have flowed in open channels in the ice (that is, channels cut in rock at the base, with ice walls rising above). This explanation cannot, however, account for the origin and main development of the channel system. Reference to Figure 2 will show that it would necessitate the postulation of narrow elongated wedges of ice balanced precariously on top of the inter-channel ridges in order to keep the waters of the various meltwater streams in their separate paths, a situation that would be especially absurd in relation to the steep-sided, sharp-crested ridges that separate some channels or rise from the floor of channel AN.

Thus, of the six possible types of glacier streams, five have been eliminated as principal agents in the formation of the channel system and only the subglacial remains. It is now necessary to consider the evidence that directly favours a subglacial origin.

(1) Several channels are so positioned that ice must have been situated on both sides of them when they were being formed. For example, channel E starts on the north-west side of a broad divide but crosses over obliquely to the south-east side. The channels at G begin close together on either side of a ridge and then diverge to run down both sides of the ridge. These channels

could have been cut in these locations only if ice was situated on both sides of the hill into which they are cut. Similar reasoning may be applied to channels K, L, M and O (Fig. 2) as well as to those at Q (Fig. 1) but in these cases a slight complication is introduced dependent on the order of formation of the various channels. Channel M may be taken as an example. If M was cut after L then ice must have been situated on the south-eastern side of M to cause the meltwaters to flow along it and ice must also have lain on its north-western side to prevent the waters following the much deeper channel L. Alternatively, if M was cut before L it must have been covered by ice when L was cut (unless one conveniently postulates a readvance of the ice after M was formed). Since this reasoning also applies to L in relation to AN, to K in relation to AN, and to O and the channels at Q (Fig. 1) in relation to the deep channel containing Carlops village and the North Esk, it follows that channels K, L, M, O and Q were cut when ice lay on both sides of each. Since, as pointed out above, the open ice-walled channel concept is unsatisfactory, it follows that these channels were formed subglacially.

(2) Strong evidence for subglacial origin of channels is quite often provided by their up-and-down long profiles. In the present area there are three channels with such profiles: C, K and AN. Unfortunately, in two of these instances the evidence is inconclusive. The highest part of the floor of channel C is an alluvial fan deposited by a small stream that flows into it from the north-west, so that the original profile of the channel is uncertain. Channel AN rises about 45 feet north-eastwards in 900 yards before beginning to descend. Eckford[7] suggested that the south-westward slope was produced by a post-glacial stream. This is not so, however, for the present stream is far too small to have effected so much erosion and its erosional influence is limited to the extreme south-west, where it occupies a small incision at the side of which the original channel floor still survives. There is, however, a strong possibility that the north-eastward rise is due to aggradation. Significantly, the highest part of the channel floor lies near the highest part of the channel wall on the north-western side. Rock debris from this steep slope, which at its highest part rises 200 feet above the channel floor, could well have contributed significantly to the material on the channel floor. The basal part of the rock wall on the north-western side of the channel above the junction with channel C is almost everywhere obscured by a concave debris slope and in places small fans are present also. The south-eastern side of the channel is also unstable, for slump-marks and other minor irregularities indicate that mass-movement is active at the present time. Thus it is likely that a considerable amount of aggradation of the channel floor has occurred so no subglacial significance can be attached to the visible up-and-down profile. The third channel, K, does seem significant, however. No streams follow this channel or flow into it, no peat lies on its floor and its sides are so low that significant infilling by mass-movement seems improbable. Yet the channel floor rises some 20 feet north-eastwards in 150 yards before descending again. It seems that the only satisfactory explanation of the profile of this channel (if, as appears, it is

a truly erosional feature) is erosion by meltwaters flowing subglacially under hydrostatic pressure.

(3) The north-western side of channel AN is cut into by a considerable number of meltwater channels in areas B and D. Since some of these features run down the side of channel AN it is first necessary to show that they are, in fact, channels and not post-glacial gullies. Several such gullies are indicated by the contours in Figure 2, but nowhere is there such a concentration of down-slope depressions as occurs in areas B and D. This concentration is in the least likely place for post-glacial gullying since most of it occurs where the catchment area is very small. Furthermore, the depressions are all cut in rock and some of them are considerably larger than post-glacial gullies elsewhere in the Carlops area. Finally, several of the features follow courses that can be explained only by postulating the presence of glacier ice when they were being cut. The best example occurs opposite the highest point on the floor of channel AN. Here, for a distance of about 100 yards, one dry depression lies nearly parallel with the steep wall of AN and is separated from it only by a very narrow ridge of rock. Unless glacier ice were present this feature could not have been formed, for the stream would have plunged straight down the steep side of AN. The dry depressions in areas B and D are therefore meltwater channels and they were formed subglacially, for unless the slope on which they occur was covered by ice meltwaters could not have cut the channels.

The relief before the main channel system was formed. All the channels in area B terminate part way down the steep north-western side of channel AN some 50 to 80 feet above the floor of the latter. Such sudden endings of down-slope channels often indicate that the waters became englacial thereafter. In the present instance, however, this does not appear to be the explanation. Most of the channels end at the level of a very narrow but well-marked rock bench that extends intermittently along the north-western side of channel AN in this area. This bench is structurally controlled but structure alone does not seem adequate to explain its existence. (Were this the only cause there should be numerous channel-side benches in the Carlops area.) The most satisfactory explanation appears to be that at the time the down-slope channels of area B were cut, channel AN had been excavated only to the level of the bench (about 1050 to 1080 feet), which was then its floor.

A little farther north-east the fairly large channel, C, joins AN. The lowest part of C is much narrower than the remainder and it seems likely that when most of this channel was cut its waters were flowing into AN at a level about 50 to 60 feet above the present floor level of the latter (that is, about 1050 to 1060 feet O.D.). A little farther along channel AN and on its south-eastern side there is a small but clear bench at 1050 to 1070 feet, some 50 to 70 feet above the floor of AN. The sharp-crested rock ridge that rises from the floor of AN at H reaches 1045 feet. Near by, on the north-west side of AN, is a very clear till-covered bench, one-third of a mile long and 100 yards wide, that falls gently

from 1050 feet at its south-western edge to 1020 feet at its north-eastern margin. Further north-east from Dun Kaim to around Carlops the ground between the various channels rises to about 970 to 1005 feet, strongly suggesting that a fairly smooth surface at this level existed here before the channel system was excavated.

Taken together these various remnants point to the presence of a land surface over the site of channel AN that descended from a level of about 1050–1080 feet in the south-west to about 970–1000 feet in the north-east. Between A and H this land surface was the floor of a depression about 100 feet deep. This depression may have been formed by meltwaters, but not during the last deglaciation, since most of its sides and some of the remnants of its floor bear considerable amounts of till. Alternatively, ice erosion may have contributed significantly to its form (or, perhaps, both agents were partly responsible). As already noted, ice moved south-west–north-east in the Carlops area and its most obvious moulding effects are seen along the lower slopes of the Pentland Hills (Fig. 1). The ice moved parallel with the major Pentland Fault system (which coincides with channel AH), so that the chances of a col being formed (or, if it already existed, deepened) by the ice along this line of weakness were high. Bearing in mind the form of the land surface into which most of the channels were cut we may now return to the consideration of the channels themselves.

Contemporaneous formation of sub-parallel channels. Apart from the channels in area B and the greater part of channel C, the evidence suggests that the remainder of the channels in the area south-west of Carlops were cut more or less contemporaneously.

The most remarkable feature of the system is the steep-sided rock knolls and ridges rising from the floor of channel AN. Since it has already been shown that the only hypothesis that accounts for the channel system is the subglacial one, it seems that two explanations of these rock 'islands' are possible: either the meltwaters were diverted from parts of channel AN by the collapse of subglacial tunnels and thus caused to cut a new channel at the side of the old blocked one in certain places; or the channels on each side of the 'islands' were formed contemporaneously. The former explanation seems unsatisfactory for two reasons. First, it is difficult to see why, if a part of channel AN was blocked by collapse of a subglacial tunnel, the meltwaters cut a deep channel in rock instead of eroding the ice block. Secondly, the rock floors of the channels on either side of each rock island are probably of similar depth. This is suggested by the form and size of the channels, although not proved owing to the presence of alluvium and other drift on the channel floors and the absence of boreholes. The second hypothesis, contemporaneous erosion of the double parts of channel AN, is therefore favoured.

This hypothesis can be applied more widely. Although the 'islands' in channel AN are very striking in the field, they are really no different, except in

size, from the rock ridges between other channels. From the 35-foot high steep-sided knoll of aptly named Peaked Craig the rock ridges increase in size through the feature misleadingly named Dun Kaim to the ridges between channels L and AN and between L and M, and finally to the long ridge separating channels AN and EJ. It is not necessary to invoke collapse of ice tunnels to explain the channels between the larger ridges, for these channels are clearly the result of several distinct glacial streams flowing across the area. Furthermore, since contemporaneous subglacial cutting of the double parts of channel AN appears to be the only satisfactory explanation, it is reasonable to presume that the same explanation applies to most of the remainder of the channel system (excluding channel C and the down-slope channels in areas B and D). This interpretation, in fact, provides the simplest explanation of the relationships between certain channels, such as K, which is cut off at both ends by channel AN, and channel L, which begins some 20 feet above the floor of AN. Such rather peculiar relationships would then simply imply that individual elements of a subglacial braided stream pattern had been successively abandoned until only the stream engaged in cutting AN survived.

That such a series of tunnels can exist beneath glacier ice is shown by descriptions in the literature, such as J. C. Stokes's description of a tunnel system he partly explored beneath a glacier in Norway.[8] More widespread evidence is provided by the great esker belts (for example, on the Canadian Shield) that comprise braided systems of steep-sided ridges. Nearer at hand the large Carstairs esker system in Lanarkshire provides similar evidence. The esker system north-east of Carlops described above indicates that a braided glacial drainage system existed in the Carlops area. Since this esker system is dissected by the larger channels it must have been formed contemporaneously with the earlier stages of formation of the channel system south-west of Carlops. Thus, in the Carlops region as a whole, it seems that, before the meltwaters became concentrated in the large channels, a braided subglacial stream system traversed the area, eroding in its south-western part to initiate the channel system and depositing in its north-eastern part to produce the esker system.

The initiation of the channel system. Some subglacial drainage systems show a strong correlation with the relief of the areas in which they occur: for example, both channels and eskers often utilize pre-existing cols in crossing watersheds. In the case of eskers this fact has often been cited as an argument for subglacial rather than englacial origin, although it is not clear how the esker streams are presumed to have located the cols buried beneath the ice. In other instances, however, channels at least show little relation to the main relief features, for they cross spur crests indiscriminately, ignoring cols and even being cut into the tops of hills. These contrasting relationships with relief may reflect different modes of subglacial origin. In the area south-west of Carlops, however, both relationships are present and both can be explained by a single hypothesis. For example, channel AH follows the course of a pre-existing depression, whereas

channel F cuts obliquely across the crest of a hill. The hypothesis that can explain these features is that of superimposition of englacial streams from the ice, a hypothesis recently developed by R. J. Price[9] for a large number of small channels in the valley of the upper Tweed. The alternative would be to assume that the streams that cut the channels were subglacial from the time they first began to flow. The evidence favouring superimposition rather than this alternative interpretation will now be considered.

Before meltwaters began to flow in the Carlops system of channels they followed routes at higher levels in the Pentland Hills, as shown by a considerable number of channels and, in a few instances, by eskers (Fig. 1). On the spur into which the channels south-west of Carlops are cut there is only one major channel that carried meltwaters in a general north-eastward direction, which is the regional trend of the glacial drainage over this and a much wider area. The earlier stages of formation of this channel (P, Fig. 1) presumably correlate with the cutting of some of the other channels on the Pentland slopes to the east, but at a later time the waters from channel P flowed southwards along the valley of the North Esk. This is shown by two remnants of an esker and by the fact that channel P descends to the floor of the North Esk valley. The meltwaters following this valley were probably responsible for the channels at Q (Fig. 1) that 'hang' at both ends well above the floor of the deep channel containing Carlops village.

At the time channel P ceased to function the ice surface in its vicinity must have been at an altitude of at least 1470 feet, which is the approximate height of its floor at its intake. If the meltwaters, after abandoning channel P, flowed marginally to the ice or occupied successively lower subglacial routes there should be some evidence of this on the hill-slopes between channels C and AN, Figure 2, and 1470 feet, for channel P (Fig. 1) is a major channel (up to 100 feet deep in rock) and was clearly cut by large volumes of water. Yet on the Pentland slopes above the Carlops channel system there are only some small vague benches of doubtful origin and two channels (just outside the area covered by Figure 1) whose moderate size and north–south trend suggest that they are not directly related to the major meltwater streams that initiated the Carlops system. It seems, therefore, that the problem of transferring the major meltwater drainage from channel P to the Carlops system can only be resolved if it is assumed that the Carlops system was initiated by streams superimposed directly from the ice.

If one imagines a system of englacial streams being superimposed on to an area of varied relief like that south-west of Carlops, one would expect that where slopes were gentle these streams would incise themselves into the ground beneath. On the other hand, on the steeper slopes one might expect the streams to move both laterally and downwards along the ice-ground contact towards the floors of cols or to areas of gentle slope (that is, towards concave breaks of slope). Furthermore, as suggested by Price,[10] streams superimposed on the higher ground would probably cut down more slowly than adjacent streams still

cutting down through the ice towards lower ground. Hence, one might expect the latter streams to tap the waters of the higher ones (if connections between them existed), thus resulting in the formation of the larger channels in the areas where the pre-existing relief was lowest (for example, in cols). Furthermore, in areas of varied geological structure, one might also expect that streams superimposed on weaker belts, or streams that had moved both laterally and downwards towards these (owing to these weaker belts already being at a lower level as a result of glacial erosion or other processes) would be especially favoured, so that a correlation between channels and lines of structural weakness might be expected. In the area south-west of Carlops such expectations are often fulfilled.

The manner in which channel E crosses the watershed of the elongated hill named Cock Rig is consistent with the idea of superimposition. Here the slopes are very gentle and a superimposed stream would not be induced to move laterally. Towards the north-east, however, below its junction with channel F, channel E suddenly turns through about 60 degrees. This sharp bend corresponds with the convex break of slope that marks the edge of the gently sloping surface of Cock Rig. It thus appears that the waters cutting channel E adjusted themselves to the steeper slope between the 1100 and 1070 foot contours. Channel E then resumes its north-eastward course and for some distance follows a pre-existing relief feature, the concave break of slope that marks the limit of the high ground to the north-west.

Near G several channels start in a small area and for a very short distance follow north-eastward courses, parallel with the regional drainage, before diverging on either side of the ridge in which they are cut. The sudden commencement of these channels in this location can be explained if the streams that cut them were englacial south-west of the points at which the channels begin. This is in accord with the north-eastward trend of the channels in their upper parts. However, the strong relief rapidly affected the streams once they had become subglacial and caused them to move down the ice-rock contact towards more stable positions. Further south-west on the Cock Rig upland channel F starts very suddenly and this suggests that it also was cut by a stream that was englacial south-west of the point at which it begins.

Most of the deeper channels south-west of Carlops are aligned along faults shown on the maps of the Geological Survey. While there is a possibility that the surveyors may have tended to place the faults along morphological features, this seems unlikely to be the full explanation of the correlation, for rock crops out at many points on the sides of the larger channels. Assuming the geological maps to be valid, the following correlations occur. The floor of the south-western part of channel AN lies exactly along the line of the Pentland Fault. Near point H the northern branch of the channel leaves the south-west–north-east line of this fault but resumes the same direction after a short distance along a parallel fault. Where the channel splits on either side of Dun Kaim each branch is along a south-west to north-east fault and the same is true of the

branches on either side of the small rock knoll to the north-east. A north-west to south-east fault lies between this knoll and Dun Kaim, while another fault corresponds with the south–north section of the left branch of the channel round Dun Kaim. The correlations do not apply everywhere, however: for example, for about half its length through the area the main Pentland Fault is not followed by a channel, while a fault follows the crest of the rock ridge H. Nevertheless, the correlations that do exist are so strong as to require explanation. As already indicated, this is provided by the hypothesis of superimposition. The regional trend of the glacial drainage system was parallel with the principal faults. Some elements of this drainage system on cutting through the ice to the ground beneath found themselves in localities where geology or relief were unfavourable and cut only small or moderate-sized channels, such as E, F and G. Other streams, however, were superimposed on the fault lines; or, where previously formed depressions already existed along these faults, the streams may have slipped down into them along the ice-ground contact. Streams favoured by the presence of faults cut down more rapidly than other streams, so that the courses of the latter, such as K and L, were abandoned until the drainage became almost entirely concentrated along the single and double parts of channel AN. This channel therefore became the largest in the area south-west of Carlops.

Conclusion

The Carlops glacial drainage features are part of an extensive system of such features that trends in a general south-west–north-east direction over an area extending at least as far south-west as the Clyde and north-eastwards to the sea around Dunbar. The Carlops channels owe their great size to the concentration of meltwater flow, during the later stages of the decay of the ice-sheet that covered the area, through the gap between the Moorfoot and the Pentland Hills. The evidence indicates that the streams that cut the channels and deposited the related eskers originally flowed englacially and were let down on to the ground beneath, thus becoming subglacial. How deep beneath the ice the streams were flowing when this took place is not known from evidence in the area, but it is considered probable that they were some 300 to 400 feet beneath the ice surface. This figure is suggested since the writer knows of many channels and some eskers elsewhere that prove that meltwaters penetrated 300 feet beneath glacier ice but does not know of any instances where it can be proved that 400 feet was exceeded. The upper 300 to 400 feet of ice is thus visualized as being in a state that permitted the flow of water through it, possibly under high hydrostatic pressure at the base of this zone (allowing up-and-down channels to be formed) and under free flow at higher levels. Since the ice had come from the south-west its surface must have sloped down towards the north-east, so that, unless the slope of the ground in this direction was greater than the slope of the ice surface, the zone in which stream flow was possible would reach ground-level first in the north-east. As the ice surface

continued to waste down, the roofs of the subglacial tunnels became thinner, until they finally collapsed, a process probably hastened as meltwater flow became concentrated in the very large channels. Collapse of the tunnels resulted in the meltwaters flowing in channels whose walls were formed of rock in their lower parts and ice in their upper parts. This process appears to have begun first in the north-east and extended back south-westwards. In the final stages of meltwater flow through the Carlops area the only channel functioning was YV, which at this stage was a proglacial channel, carrying waters from the West Linton basin, where a mass of dead ice was wasting away. Ultimately this channel too was abandoned when meltwaters found their way through the ice to the Tweed valley along the valleys of the Tarth Water and Lyne Water,[11] as had already happened at a slightly earlier time in the near-by Eddleston valley.[12]

In conclusion it may be mentioned that many of the features of the Carlops glacial drainage system are found in glacial drainage systems elsewhere in Britain, and the writer believes that the origins deduced for the Carlops features may have a much wider application.

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THE PERTH RE-ADVANCE IN CENTRAL SCOTLAND

PART I

J. B. SISSONS

As long ago as 1863 A. Geikie[1] showed that a period of valley glaciation followed the decay of the last ice-sheet in Scotland. Thereafter, during the remainder of last century and the first quarter of the present one it was generally held that the wastage of the great *mer de glace*, although it may have been temporarily interrupted from time to time, was, on the whole, a fairly regular process. In 1926, however, Charlesworth[2] introduced a new concept to the Scottish glacial story, for he described a major ice-sheet re-advance. It was not known how much of Scotland had been freed of ice before this re-advance, but during it Highland ice was considered to have occupied the central lowlands and Southern Upland ice to have covered the western and central Southern Uplands. The combined ice-limit was described as extending from St Abb's Head, along the northern slope of the Lammermuirs, across the middle Tweed basin and thence through the Solway lowlands to Stranraer. The features associated with this limit were collectively referred to as the Lammermuir-Stranraer moraine.

In 1933 Simpson[3] demonstrated two distinct re-advances in the Highland-border belt between the Tay and Loch Lomond. The later of these two produced the moraine belt that encircles the southern end of Loch Lomond and the morainic arc that encloses Lake of Menteith and the western part of the Forth carse-lands. Donner[4] has concluded that this re-advance occurred in Zone III of the British pollen sequence, that is, between about 10,800 and 10,300 years ago. Simpson showed that during the earlier re-advance, named by him the Perth Re-advance, ice streams from the Highlands moved down the Earn and Almond valleys to near Perth and along the Forth valley past Stirling as well as into Strath Allan. He also suggested tentatively that the Forth ice extended a considerable distance to the east along the southern side of the Ochils, perhaps as far as Glen Farg due south of Perth.

The purpose of this paper is to attempt to establish the limit of the Perth Re-advance in central Scotland as a whole. The most important area in this context is the Forth valley between Stirling and Falkirk, for not only are certain relationships of ice-limits to outwash fans and raised beaches evident in this area, but a connection between glacial events in the Forth and Clyde valleys can be established. For this reason the middle Forth area will be considered first and in more detail than other areas. However, before discussing the evidence by which the re-advance may be delimited, it is necessary to state in general terms the types of evidence used.

THE GENERAL EVIDENCE FOR THE PERTH RE-ADVANCE IN CENTRAL SCOTLAND

Stratigraphic evidence (a) Proof of an important advance of glacier ice over an area it had previously vacated is provided by the existence over a considerable

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area of a succession of deposits comprising, in order, till, stratified sediments, till. The upper till must be of sufficient thickness to exclude the possibility of its having been deposited from dead ice as ablation moraine on top of stratified sediments accumulated subglacially. Likewise, the stratified sediments must be sufficiently thick and continuous to preclude the possibilities of subglacial or englacial origin and of their having been picked up by the ice from water-laid deposits over which it moved. These requirements are fulfilled over a large area in the middle Clyde basin from the foot of the Campsie Fells[5] southwards to beyond Lanark and again in central Ayrshire.

(b) If the middle member of the three-fold sequence contains organic remains the evidence for a temporary withdrawal of the ice is even more satisfactory. In the two areas referred to such remains have been obtained from several localities. Most of these discoveries were made last century and were interpreted as 'interglacial' (or even 'preglacial'). As pointed out above, however, it was not then appreciated that major re-advances interrupted the decay of the last ice-sheet. In the light of more recent knowledge therefore the organic remains are much more likely to be interstadial. Two points favour this view:—(i) the fossiliferous stratified deposits appear to be covered by only a single bed of till[6] and the two together as known at present are restricted to the western half of the central lowlands (no such sequence has been found as yet in the Forth and Tay basins); (ii) with one exception the organic remains are consistent with or demand, a cool climate.

(c) Another type of stratigraphic evidence strongly suggesting a re-advance of the ice is the extensive development of outwash gravels and sands over thick deposits of clays and silts. Such evidence occurs along the South Esk valley inland from Montrose, where the fine sediments are of marine origin, and in the valleys of the Almond, Earn and Tay near Perth, where the fine sediments are of marine or estuarine origin. The accumulation of the silts and clays strongly suggests a considerable period during which glacier ice was so far away that its outwash streams were unable to supply coarse sediment; the coarse deposits imply close proximity of glacier ice and hence glacier re-advance. Since, at the localities mentioned, till neither overlies nor is interbedded with the stratified deposits, the ice itself did not reach these localities at this time[7].

Geomorphic evidence (a) When the decay of glacier ice terminating in the sea is accompanied by rapid deposition by meltwater streams, extensive sandy beaches are formed beyond the ice-limit, especially in sheltered localities. Within the ice-limit the decaying ice is often buried beneath the stratified deposits and, consequently, the beaches later develop an irregular surface form when the buried ice melts, producing kettle-holes and other dead-ice hollows. In the Forth valley between Stirling and Falkirk both types of beach (now raised) are present: the irregular type occurs near Stirling but is quite absent from near Cowie to Falkirk, in which stretch only smooth raised-beaches exist. The down-valley limit of the irregular type of beach can only be satisfactorily explained as marking the limit of a re-advance. Similar evidence occurs in the Carron valley between Larbert and Bonnybridge.

(b) A similar type of reasoning applies to dead-ice hollows in fluvioglacial deposits in supra-marine localities. As the terraces of the North Esk, West Water, South Esk, Tay, Almond and Earn are followed upstream a point is reached in each valley where the smooth terrace-surfaces are replaced by irregular forms and kettle-holes begin to appear. The point of change is most satisfactorily explained as marking the limit of a re-advance, an interpretation that is supported in several of these valleys (at least) by an increase in the size of the terrace materials up-valley.

(c) The evidence so far mentioned in general terms shows that at various localities from eastern Strathmore to the Clyde basin there is evidence of a re-advance. The distribution of this evidence in a belt across central Scotland and the paucity or absence of such evidence to the east suggests the likelihood of the various phenomena being related to a single major event. This interpretation is supported by the distribution of fluvioglacial features.

On ascending the valleys of the Carron, Forth, Earn, Almond, Tay, South Esk, West Water and North Esk the change in form of the raised beaches or terraces mentioned above is seen to be associated, either on the valley floors or on the adjacent slopes, with the appearance of forms indicative of the former presence there of glacier ice, such as kames, kame terraces and meltwater channels. Often these features appear suddenly in great abundance and clearly indicate the limit of a re-advance. Frequently these fluvioglacial forms ascend the side of a valley and cross the divide into the next valley, thus permitting a correlation to be established between different valleys. In this way, apart from short gaps, it has been possible to trace a continuous limit from western Strathmore to near Lanark.

It may seem that the mapping of fluvioglacial features to establish the limit of a re-advance is a precarious occupation in view of their widespread distribution in lowland Scotland and their complex origins, a contention with which the writer would have agreed until recently. Several factors facilitate this task in the present instance, however, and justify its use. Firstly, the limits of assemblages of fluvioglacial features (individual features are of limited significance) are often very sharply defined. In particular along the southern side of Strath Allan, the eastern side of the middle Clyde basin and the northern slopes of the uplands south of Eaglesham, a line can be readily mapped above which meltwater features are poorly developed or quite absent and below which they abound.

Secondly, west of the line shown in Figure 1 the meltwater channels often form different patterns and frequently occur in much greater numbers per unit area than they do to the east of the line. West of the line (especially on the flanks of Strath Allan, some of the slopes of the Campsies, the northern side of the Vale of Menteith and the southern slopes of the Darvel-Strathaven gap) meltwater channels often form complex, interconnected systems with densities of 20 to 40 per square kilometre. Most of the channels are less than 20 feet and many are less than 10 feet deep. Systems of this type are illustrated in Figure 2 and in a previous article[8]. In contrast, the writer knows no area east of the line

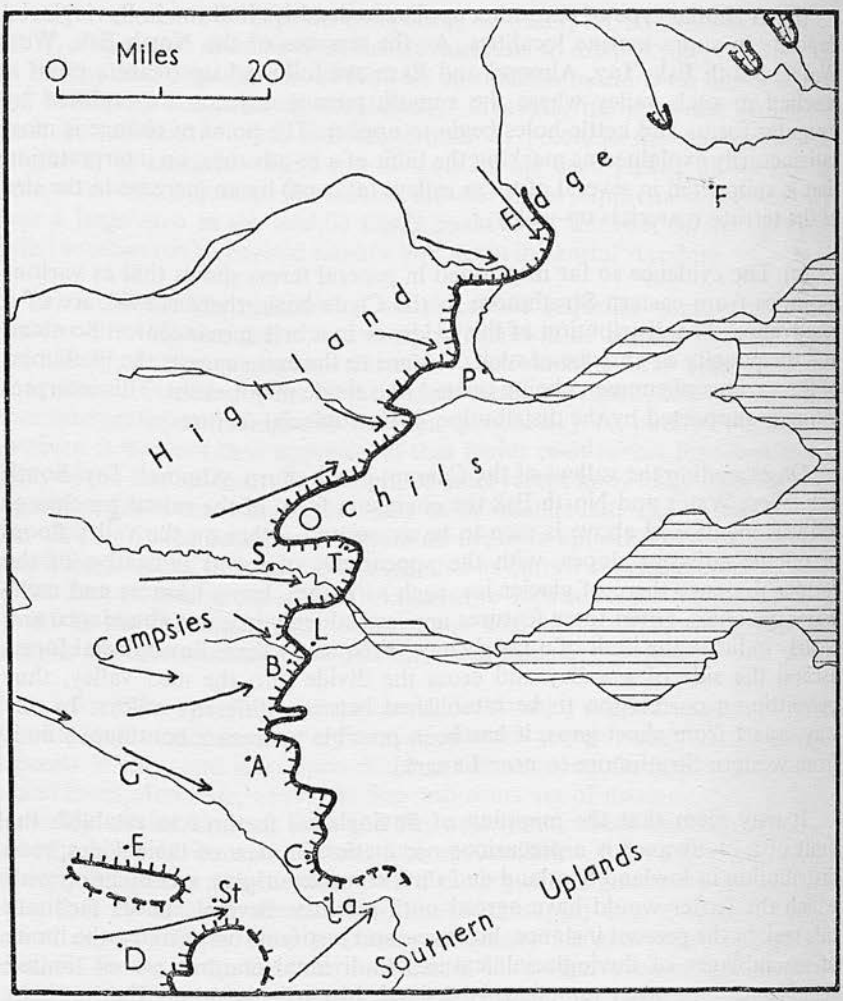


Fig. 1. The limit of the Perth Re-advance in central Scotland. F, Forfar; P, Perth; S, Stirling; T, Tillicoultry; L, Larbert; B, Bonnybridge; A, Airdrie; C, Carluke; La, Lanark; St, Strathaven; D, Darvel; E, Eaglesham; G, Glasgow. Note: nunataks in the Highlands and Campsies are not shown.

where channels form such patterns. The closest approximation is on the western side of the Eddleston valley, Peeblesshire, where occurs the highest density of channels so far mapped in south-east Scotland (two kilometre squares each having about 30 channels); yet even here there are significant differences, the channels being much more variable in size, length and direction than in many areas of similar size and relief west of the line shown in Figure 1[9].

Thirdly, the task of delimiting the re-advance is facilitated by the greater clarity of many erosional and depositional meltwater features west of the line

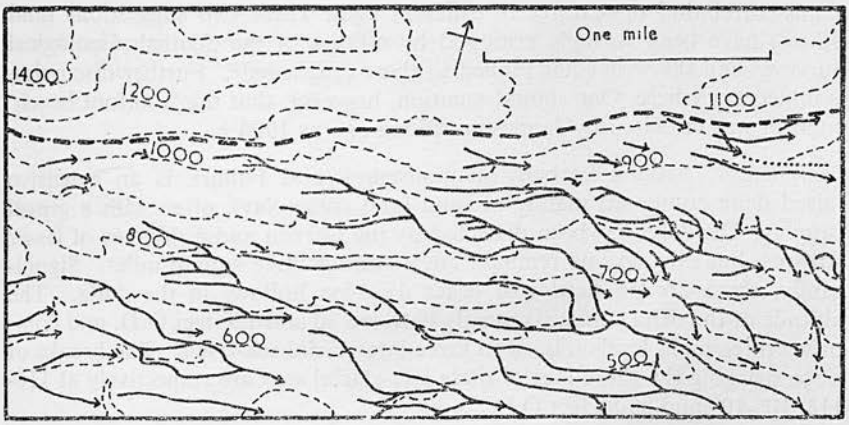


Fig. 2. Meltwater channels on the slopes of the Campsie Fells north and north-east of Kilsyth. Channels are represented by arrows and an esker by a line of dots. The heavy broken line, marking the upper limit of abundant glacial drift and of the meltwater forms, is considered to mark the upper limit of the Perth Re-advance on one of the Campsie nunataks. Immediately east of the eastern margin of the map area this limit begins to decline more markedly.

than east of it. The change on crossing the line is often very marked and is reflected in the ease of mapping: thus, far fewer 'doubtful' features are encountered, meltwater channels only a few feet deep can often be mapped with confidence and many kames are sharper and their slopes steeper. This suggests that the features west of the line are of more recent origin than those east of it and hence are related to a major re-advance of the ice. In accord with this is the greater number of kettles still containing lakes in the western area. Thus in the extensive fluvioglacial deposits between Dunbar and the Pentlands and including the Eddleston valley only two kettle-lakes exist. On the other hand several such lakes exist in Strath Allan and in the Carstairs area and there are about ten in the Blairgowrie-Murthly area.

Together, the facts so far described in general terms indicate a major re-advance of glacier ice to the vicinity of the limits shown in Figure 1. The Scottish glacial literature includes much evidence favouring this view and this will now be drawn upon and integrated with the writer's evidence in a more detailed consideration. In a few instances published interpretations of the evidence differ from those of the writer and will be discussed.

THE FORTH VALLEY

Before discussing the evidence in the Forth area it is necessary to state that the glacial and marine deposits of this area have been the subject of recent controversy. Simpson[10] tentatively suggested that the ice that moved through the Stirling gap at the time of the Perth Re-advance extended eastwards along the south side of the Ochils as far as Glen Farg. The present writer[11] has indicated why this interpretation is unsatisfactory and has suggested that the limit of the re-advance in the Forth valley is in the Stirling area, a view more in accord with evidence described by Dinham[12]. At the same time the writer suggested that the long-accepted '100-foot raised beach' of Scotland is

a mis-correlation of features of different ages. These two suggestions (and others) have been strongly criticised by officers of the Scottish Geological Survey[13] and the writer has replied to these criticisms[14]. Further discussion is unnecessary here. One should mention, however, that the '100-foot beach' concept was criticised by Jamieson as long ago as 1906[15].

Morphology. Around Larbert, Stenhousemuir and Falkirk is an extensive raised delta composed mainly of sand (and some clay), often with a gravel capping. The delta has been dissected by the Carron and a number of lesser streams, but even so one remnant covers about three square miles. Significantly, there are no kettles or other dead-ice hollows in the delta. The altitude of the delta surface is mostly between 80 and 120 feet O.D. and comprises three main levels related to three late-glacial sea-levels. The breaks of slope marking the shorelines of these late-glacial seas are respectively at 116-118, 108-109 and 85-88 feet O.D.[16].

The highest shoreline marks the landward edge of the delta and forms a very clear feature near Larbert. It closely follows the A9 road to Plean, rising to 121-123 feet (due partly to isostatic warping). Near Plean the surface deposits of the beach become noticeably coarser and merge into an outwash plain of coarse gravel that rises regularly to about 150 feet (A, Figure 3). The outwash in turn is replaced by meltwater channels (B, Figure 2) and in the same locality a dead-ice hollow (C) occurs in the outwash. At its north-western margin the outwash is extremely coarse and here a pile of cobbles and boulders, obviously cleared from the adjacent land, covers 40 square yards and includes boulders up to two and a half feet long. Together this evidence indicates the position of the ice-limit when the 116-123-foot shoreline was being formed.

About one and a half miles to the north-east the ice-marginal zone is marked by the conspicuous, coarse, fluvioglacial accumulations known as the Berry Hills (D, Figure 3), a group of massive kames up to 70 feet high. Between and around the kames are kettles (one containing a lochan) and other dead-ice hollows. South of them an outwash plain (E) slopes down to merge with the main outwash plain A. Immediately to the north and east of the Berry Hills the kame deposits pass beneath the upper marine-limit. Here, instead of steep-sided kames, subdued forms occur, producing a gently rolling topography of mounds and dead-ice hollows (including two kettles), interspersed with raised-beach flats whose limits are often difficult to define (F, Figure 3). It is clear that here, at approximately the time the Berry Hills and the features north-west of Plean were being formed, the ice was decaying in the sea and blocks of it were buried beneath fluvioglacial debris, which was itself smoothed over by marine action. But from point G, Figure 3, down the Forth valley at least as far as Falkirk no such features occur, despite the ideal conditions for their development. A significant re-advance of the ice appears to be the only satisfactory explanation of this evidence.

In the southern outskirts of Stirling occur two more outwash plains. Each of these commences at its north-western end in an area of well-defined kames and kettles and slopes down south-eastwards to merge with raised-beach

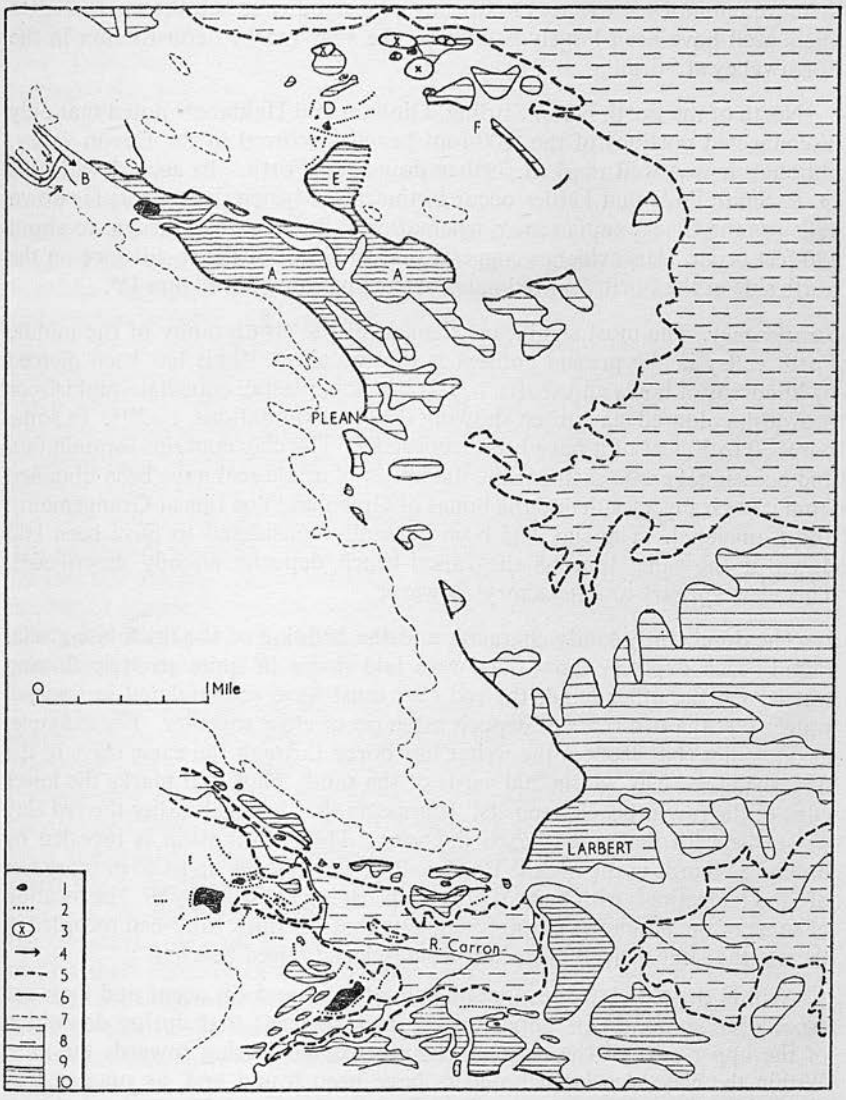


Fig. 3. Fluvioglacial features and raised beaches associated with the Perth Re-advance in the area around Larbert and Plean. 1, Kettles and other dead-ice hollows. 2, Kames. 3, Kames smoothed by marine action. 4, Meltwater channels. 5, Upper limit of post-glacial deposits. 6, Upper limit of *highest* raised beach, outwash plains and kame terraces associated with the maximum of the Perth Re-advance. 7,200-foot contour. 8, Post-glacial deposits: carse 'clay', peat, river terraces and alluvium. 9, Raised-beach flats associated with the Perth Re-advance (these are generalised and only the highest shoreline is shown). 10, Outwash plains; kame terraces and kame plateaux associated with the Perth Re-advance.

deposits. These features show that, following the maximum of the re-advance, there were minor re-advances to Stirling or a significant halt there. The latter might well have been largely caused by the very marked constriction in the Forth valley at Stirling.

North of the Forth below Stirling, Dinham and Haldane[17] noted that only disconnected portions of the '100-foot beach' occurred in the Devon valley, although it was well marked further down the Forth. In accord with this D. E. Smith finds that kettles occur in the raised-beach deposits as far down valley as the Clackmannan area, whereafter well-defined beaches up to about 120 feet occur. His evidence suggests that the limit of the re-advance on the north side of the Forth approximates to the line shown in Figure 1[18].

Stratigraphy. The most significant element of the stratigraphy of the middle Forth valley in the present context is the red clay. "This has been pierced by hundreds of bores and shafts. . . . It is a fine, plastic, chocolate-pinkish- or brownish-coloured clay, often showing delicate laminations. . . ."[19]. In some bores 30 to 50 feet of it have been recorded[20]. The clay contains foraminifera and occasional arctic shells, while the bones of arctic seal have been obtained from it at Grangemouth and the bones of Greenland floe rats at Grangemouth and Camelon[21]. The clay has been generally considered to have been laid down at the same time as the raised-beach deposits already described[22]. This view appears unsatisfactory, however.

The dominantly sandy character and the bedding of the thick late-glacial raised-beach deposits show they were laid down in quite strongly flowing water. On the other hand, the red clay must have accumulated in tranquil water. Yet the two types of deposit often occur close together. For example, at one point (NS 860861) the writer has bored through the carse clays to the underlying red clay within 200 yards of the sandy bluff that marks the lower edge of the raised-beach deposits. It appears that in this locality the red clay passes beneath the raised-beach deposits. This relationship is revealed by borehole records in the middle Forth valley, which show up to 40 or more feet of sand (sometimes with beds of gravel) overlying the red clay[23]. The location of some of the boreholes strongly suggests that the thick sand-bed recorded in them is the same deposit as forms the late-glacial raised beaches.

Within the red clay occasional layers of sand and silt occur and these are especially common in its upper part[24]. This suggests that during deposition of the upper part of the clay, glacier ice was advancing towards the area. Within the clay occasional boulders have been found and, as suggested by Cadell, may have been dropped from melting ice-floes. However, on top of the red clay and intervening between it and the overlying sands there occurs in places a well-marked layer of boulders, some of large size[25]. Melting of ice-floes seems inadequate to explain this sudden onset of boulder deposition: the explanation may be that the advancing ice was sufficiently near to supply abundant bergs that floated down the Forth and dropped their debris as they melted. Perhaps this became possible when the ice had advanced beyond the narrow gap at Stirling and had begun to spread in the middle Forth basin.

In summary, the stratigraphic sequence comprises red clay, implying

distant ice in its lower part and advancing ice in its upper part; a layer of boulders, suggesting that ice had advanced beyond the Stirling gap; and thick deposits of sand supplied by meltwater streams from retreating ice. The stratigraphic evidence thus demonstrates a major re-advance and accords with the morphological evidence already described.

STRATH ALLAN TO PERTH

Morphology. Simpson[26] showed that the ice in Strath Allan and Strath Earn at the time of the Perth Re-advance moved towards the east and north-east, that is, more or less parallel with the trend of the Ochils and the Highland edge. This movement is demonstrated by the distribution of schistose rocks from the Highlands farther west, and of Old Red sandstones in relation to the Ochil volcanic rocks. On the southern slopes of Strath Allan east and north-east of Dunblane the limit of the re-advance is marked for several miles at an altitude of about 800-900 feet by the sharp upward limit of a large system of small interconnected meltwater channels[27]. Many hundreds of similar channels furrow the northern side of Strath Allan and extend along the northern slopes of the Vale of Menteith to near Callander. The upper limit of this vast channel system rises gradually westwards, and is at about 1,000 feet on the northern side of Strath Allan (Simpson[28] specified 950 feet behind Braco). The westward rise of this line, its correspondence with the 'morainic' limit[29], the fact that the line is not defined by a few channels but by the upper parts of numerous channels, and the similarity of the upper limit of channels on both sides of Strath Allan, suggest that this limit was also that of the ice itself at the maximum of the Perth Re-advance.

Simpson[30] pointed out that the ice limit along the slopes of the Ochils is partly defined by the limit of drift composed of sandstones and occasional Highland rocks. This drift extends nearly a mile up Gleneagles. A short distance east of Auchterarder, however, the drift at the foot of the Ochils changes and thereafter contains a significant proportion of the volcanic rocks from those hills. This suggests that eastwards from here the ice failed to reach the foot of the hills. This accords with the evidence of meltwater channels, for the upper limit of these along the basal slopes of the Ochils in the Gleneagles-Auchterarder area declines north-eastwards and channels die away a short distance east of the latter place. On the slopes of the interfluve between the Earn valley and the Crieff-Methven depression, and again on the northern side of the latter depression, other groups of channels, their upper limits declining eastwards, similarly define the ice limit.

On the floor of the Earn valley massive outwash deposits decline gently eastwards and their very abundance makes it difficult to establish the limit precisely. An approximate limit is provided by a line joining the evidence already described on the bounding slopes. This line passes east of the kettle occupied by White Moss Loch, a depression that, to have survived the abundant deposition of outwash, must have been occupied by dead ice at the time.

Although Simpson did not indicate on his maps the limit of the Perth Re-advance, he implied that the ice was rather more extensive in the Earn-Tay area than is shown in Figure 1. This difference of interpretation is a result

of two main factors. Firstly, along much of Strath Earn, and also in the Tay valley for several miles above Perth, the outwash deposits are often dissected by dry channels, some of which are of large dimensions. These channels were cut by outwash rivers (including the proglacial Tay and Earn) as the ice retreated from the re-advance limit. Simpson recognised this, but referred to the dry valleys as 'overflow channels' and appears to have believed their courses were in part ice-controlled. For this there is no evidence. Secondly, since Simpson did not distinguish clearly between fluvioglacial deposits laid down beyond the ice-margin (i.e. outwash) and ice-contact forms (e.g. kames), he was not able to identify the approximate limit of the re-advance in the Earn-Tay area.

Stratigraphy. These criticisms in no way detract from Simpson's important conclusion that a major re-advance is represented in the Earn-Tay area. This is proved by his sections showing, from bottom to top, till, thick varved-clay, and outwash gravels. These occur about two miles north-west and about five miles south-west of Perth. In both sections no layer of till overlies the varved deposits, implying that the ice of the re-advance failed to reach these points.

STRATHMORE

Erratics, ice-moulded forms and fluvioglacial features show that Strathmore was occupied by part of a great ice-stream moving S.W.-N.E.[31]. It is generally agreed that, following the decay of this ice, there was a re-advance: the Perth Re-advance. But different opinions have been expressed regarding the limit of the latter. Watson[32] mentioned that the Tay glacier expanded into Strathmore. Synge[33] said it was "possible" that a moraine six miles west of Forfar marks the limit of the re-advance, but gave no evidence for this suggestion. He also stated: "It is probable that the maximum of the valley glaciers in Glen Clova and Glen Esk occurred at this time", although on his map the Glen Clova re-advance is assigned to a later stage. Rice[34] considered the Perth Re-advance reached the Angus coast around Arbroath and in Lunan Bay. The reasons given for this interpretation are not valid because they are based on a misunderstanding of the nature of the so-called '100-foot raised beach'[35]. Rice acknowledged that his view presented difficulties and in particular noted that till does not overlie the late-glacial marine or estuarine deposits near Perth: thus his interpretation leads to the unacceptable conclusion that ice that failed to reach Perth nevertheless extended eastwards as as far as Arbroath.

Morphology. At the maximum of the re-advance the ice in the Almond valley extended only a short distance on to the low ground around Almondbank. Kettles pit the outwash here but farther down the valley they are absent. Smooth-surfaced outwash terraces, interrupted by the dry, abandoned valleys previously mentioned, extend up the Tay valley as well as up the tributary valley followed by the A9 road. On ascending the latter valley, which was a major meltwater route, the terraces become very fragmentary about one and a half miles south of Bankfoot and kames and dead-ice hollows begin to appear. This southern limit of ice-contact forms can be traced intermittently over the higher ground east of Bankfoot and then continuously on the lower ground

towards the Tay, crossing this river south of its confluence with the Isla.

East and north of the Tay it is difficult to establish the limit precisely owing to the abundance of fluvioglacial features produced during the decay of the ice that occupied Strathmore before the Perth Re-advance. However, in view of the limited distance the Tay ice penetrated southwards towards Perth during the re-advance, it can hardly have extended very far eastwards into Strathmore. The steep-sided kames that extend along the Tay valley from near Caputh to near Bridge of Isla continue only a short distance up the Isla valley and suggest that the re-advance failed to reach the site of Coupar Angus. North of the kame belt within the great bend of the Ericht-Isla is an extensive outwash plain pitted by kettles, some of which still contain lakes. Since this outwash is associated with the Perth Re-advance, the presence of kettles in it shows that this area lies within the limit of the re-advance. Further north, the Highland slopes are furrowed by numerous channels, some of which were mapped by Watson[36]. The distribution of these channels shows that the Ericht glacier and the piedmont lobe of the Tay glacier were confluent along the watershed north-west of Blairgowrie.

It might seem that the re-advance would block the Tay valley below the Tay-Isla confluence to produce a lake in western Strathmore overflowing by the long meltwater channel that leads into eastern Strathmore and the South Esk basin. However, this channel, situated north of Forfar, has an intake at approximately 200 feet O.D., whereas much the greater part of the pitted outwash plain south of Blairgowrie lies below this altitude. Furthermore, the slope of this plain is continued by the outwash terraces that flank the Tay gorge below Bridge of Isla and extend down the valley towards Perth. It thus follows that, while waters from the Tay-Ericht ice-mass may have escaped eastwards into the South Esk drainage for a limited time during the Perth Re-advance, the main escape route for the heavily-laden meltwaters was along the Tay valley. An incidental result of the great accumulation of outwash along and near the Tay was the ponding of the Isla, resulting in the wide, flat floor over which this stream now meanders in Strathmore.

Farther north-eastwards none of the valley glaciers from the Grampians extended into Strathmore at the time of the Perth Re-advance. This is proved by the fact that the south-west to north-east trending kames and meltwater channels of the preceding major glacial phase have not been modified or destroyed by glaciers emanating from the Highland valleys. In accord with this, the red Strathmore drift is not overlain by Highland till. That a re-advance occurred within the Highland valleys, however, is demonstrated by the massive outwash terraces that follow the South Esk valley across Strathmore and the terraced outwash plains of the West Water and North Esk that cover an extensive area around Edzell. These outwash deposits increase in coarseness as they are followed towards the Highland edge: north of Edzell the terraces in some sections are seen to consist largely of cobbles and boulders six inches to a foot across. Almost exactly at the Highland edge in the West Water and North Esk valleys and slightly within it in the South Esk valley the terraces end abruptly in massive kame deposits pitted with kettle holes. Although it cannot be proved by continuous tracing of the ice limit that the

re-advance limits thus clearly demonstrated in these three valleys represent the Perth Re-advance, it is difficult to believe otherwise. Not only are the limits probable when considered in relation to the limits attained farther south-west, but the abundant fluvioglacial phenomena that define them are characteristic of the Perth limit in many other parts of central Scotland[37].

Stratigraphy. Along the northern side of the Firth of Tay around Errol and below Perth is a well-marked raised beach at about 80-85 feet O.D. When followed up the Tay valley past Perth it merges into the outwash deposits and the constituent materials become coarser[38]. Beneath the outwash about two miles north-west of Perth is 40 feet of varved clay[39]. The clay bed continues down the Tay valley for several miles and is represented near Errol by the fossiliferous, arctic, marine-clay described by Jamieson[40]. A re-advance is thus indicated and the sequence of deposits closely parallels that in the Forth valley.

The sequence is repeated in the valley of the South Esk. Here Howden in 1870 described a thick bed of marine-clay with arctic fossils, which he traced seven miles inland from Montrose. The clay is overlain by a thick bed of gravel that becomes more sandy towards the sea. Howden pointed out that the gravels are identical mineralogically with the 'moraines' of Glen Clova (with the addition of some red rocks from Strathmore)[41]. He thus proved the re-advance and recognised that the gravels are the outwash deposits of a glacier that occupied Glen Clova.

Part II of this paper will appear in the April, 1964, issue of this magazine.

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[31] Synge, F. M. *S.G.M.*, 1956, 72: 129-143.
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THE PERTH RE-ADVANCE IN CENTRAL SCOTLAND

Part II

By

J. B. SISSONS

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THE PERTH RE-ADVANCE IN CENTRAL SCOTLAND
PART II

J. B. SISSONS

THE AREA WEST OF THE FORTH

Morphology The smooth-surfaced, thick deposits of sand capped by cross-bedded gravel that comprise the raised delta of the Carron extend up the Carron valley a short distance beyond Larbert (Figure 1).* Between Larbert and Bonnybridge, however, they are replaced by a system of kames, kame terraces and kettles. The floors of the kettles lie as much as 30 feet below the surface of the adjacent delta, thus proving that the Carron valley was occupied by glacier ice when the delta was being formed. The abrupt change from irregular ice-contact deposits to smooth, raised delta-deposits shows that the junction of the two marks the limit of a re-advance: if the deposits had been laid down during a general retreat of the ice from some limit farther east, the great quantities of sand and gravel could not have failed to bury the decaying ice at many points to produce kettles and other ice-contact forms[1].

The highest (116 to 118-foot) shoreline of the Carron delta extends a short distance west of Larbert, but then merges into outwash and kame terraces. This proves that the re-advance in the Carron valley was contemporaneous with that in the Forth valley, for each is associated with the same shoreline, and this can be traced almost continuously between the two re-advance limits. In other words, the limit represented in the Carron valley is that of the Perth Re-advance.

The ice that occupied the lower Carron valley reached it by two main routes: from the north-west across the Campsie Fells and from the west-south-west along the lowland belt that extends from Glasgow. This is shown by striae, erratics, drumlins and meltwater channels. Striae in the Campsie Fells show that the dominant direction of ice-movement was north-west to south-east[2] and this accords with the general direction of slope of the meltwater channels. The slopes of large areas of the Campsie Fells are corrugated by great numbers of small meltwater channels similar to those on the bounding slopes of Strath Allan. Towards and on the valley floors, especially along the Carron and its tributaries and in the Endrick valley, large accumulations of fluvioglacial material form sharp kames and kame terraces pitted by numerous kettles. In places the remains of kame terraces still cling to steep slopes. These fresh fluvioglacial forms extend up to an altitude of about 1,200 feet in the eastern Campsies, showing that only the highest ground stood above the ice surface. These abundant fluvioglacial forms terminate eastwards at the re-advance limit marked in Figure 2. The main route by which meltwaters ultimately escaped from the Campsies was the Carron valley, much of the material they carried being deposited in the Bonnybridge-Larbert area and

*For illustrations see Part I in Vol. 79 (3).

in the Carron delta. During the later stages of ice decay in the area the meltwaters cut the deep gorge in which the Carron flows above Denny.

Contemporaneity in the lower Carron area of the ice from the north-west with that from the west-south-west has been claimed on the basis of striae[3] and is established by the meltwater evidence, which forms a complex but integrated system of erosional and depositional features. The movement from west-south-west is precisely defined by the classic work of A. M. Peach on the Lennoxtown boulder train[4]. These indicator stones occur in a long narrow belt along the southern slopes of the Campsies. Significantly, they are most abundantly recorded on Peach's map within the limits of the Perth Re-advance (as defined in this paper), they extend up to a maximum altitude of about 1,300 feet (thus corresponding closely with the limit of fluvioglacial features), and a slight deflection of the northern limit of the boulder train occurs near Denny (implying contemporaneity of the ice stream that deposited the boulders with that from the north-west across the Campsies.)

The ice that moved along the southern face of the Campsies was part of a much greater ice-movement, as has long been recognised. A large drumlin field splays out from the Glasgow area eastwards and south-eastwards[5]. In the vicinity of the Kilsyth trough the drumlins indicate an eastward movement immediately east of Glasgow and this becomes east-north-east past Kirkintilloch and Kilsyth. Since drumlins can have been finally fashioned only by the last ice to occupy the area in which they occur, it follows that, during the Perth Re-advance, this ice must have come up the Clyde valley over the present site of Glasgow as it moved towards its limit in the Bonnybridge-Larbert area. This movement is also clearly demonstrated by striae[6] and erratics, especially the Garabal Hill granite from the head of Loch Fyne[7].

This evidence immediately raises the question as to how far the ice penetrated into the Clyde basin south-east of Glasgow. Mitchell[8] indicated the limit as lying athwart the Clyde basin in the vicinity of Airdrie and Coatbridge, but gave no reasons for this interpretation. Such a limit seems to be suggested by the fact that the south-eastward-pointing drumlins terminate in this area. Yet it is incompatible with the evidence on the Campsie Fells. On the southern slope of this upland there exists a complex system of interconnected meltwater channels of the type that characterises the Perth Re-advance in Strath Allan and elsewhere (Figure 3). This system of channels carried meltwaters along the hill slopes from behind Kilsyth towards the Bonnybridge area. It attains a maximum altitude of 1,100 feet north of Kilsyth, implying an ice thickness of the order of 1,000 feet. Since the ice stream here moved towards the east-north-east, it is very difficult to visualise its southern lateral margin having been as far north as Airdrie and Coatbridge. One is thus led to suspect that a considerable part of the Clyde coalfield basin was occupied by ice at this time, and this is borne out by the evidence given below.

The well-defined ice limit in the Bonnybridge-Larbert area can be traced southwards along the flanks of the rising ground east of Airdrie and Carluke (Figure 1). West of this line, which is often well defined, kames and dead-ice hollows are often numerous[9], especially on valley floors, while to the east of

it such forms are very infrequent or altogether absent. Occasionally this line is marked by morainic mounds of angular debris (e.g. south-west of Kirk of Shotts, near Muirhouse Farm, 828603). The ice limit rises southwards from about 100 feet O.D. in the Carron valley to a maximum of just over 800 feet south-east of Airdrie. It fingers eastwards into the various minor valleys to define the limits of small tongues of ice that occupied them. Throughout this part of the belt the meltwater channels, which become increasingly numerous towards Bonnybridge, decline in altitude northwards. Further south the direction of meltwater drainage is reversed but, contrary to what one might expect, there is no evidence for a southward decline in the altitude of the ice limit, which remains approximately horizontal at about 800 feet. Furthermore, the fluvioglacial deposits that here clearly define the ice limit merge into the Carstairs esker system. The significance of this is discussed later.

On the opposite side of the Clyde basin, in the Eaglesham-Strathaven area, a similar picture is revealed on the slopes of the high ground that separates the Clyde and Ayrshire basins. South-west of Eaglesham a belt of morainic mounds, described in detail by Richey and others[10], and overlying frost-shattered ground, attains a maximum altitude of 1,000 feet. These morainic mounds are part of the same depositional system as the well-known kames and kame terraces of the Eaglesham area[11]. The upper limit of all these deposits is sharply defined and above it there are no moraines or related fluvioglacial features. The limit falls slightly eastwards and lies at 850-900 feet, on the rising ground south of East Kilbride. The altitude and location of this set of features and the fact that it merges in the Strathaven-Lanark area with those already described implies that they too are part of the Perth Re-advance phenomena.

A further complication is introduced here, however. In the Strathaven area the fluvioglacial deposits defining the re-advance limit merge into those of the Darvel-Strathaven gap. Phemister[12] described the deposits of this gap and showed that, when many of them (including a large flat-topped delta) accumulated in an ice-dammed lake, both ends of the gap must have been blocked by glacier ice. Thus, at the time the Clyde basin was last occupied by glacier ice, a stream of ice moved into the Darvel-Strathaven gap from the west.

On the southern slopes of the gap the maximum extent of this latter ice-stream (attained shortly before the formation of the ice-dammed lake) is clearly marked by the upper edge of a system of meltwater channels and kame terraces, as well as by a chaotic mass of steep-sided kames in the upper Avon valley. The ice-limit defined by this assemblage falls eastwards and north-eastwards from about 1,050 feet (about 5 miles south-east of Darvel) to about 800-850 feet south-east of Strathaven. It thus accords with the Clyde ice-limit in the Eaglesham-Strathaven belt and implies that ice from the west penetrated the Clyde area at the time of the Perth Re-advance, the two ice-masses being confluent.

This in turn implies that ice occupied the Firth of Clyde and much of Ayrshire at this time. This is expectable, for, in order that a great mass of ice

should stream up the Clyde valley past the site of Glasgow, an even greater mass must have been situated in the Firth of Clyde and adjacent territory. The movement from the Firth into central Ayrshire is indicated by the numerous eastward-pointing drumlins lying north-west of Kilmarnock, by the inclination of the meltwater channel systems in the Darvel-Strathaven gap, and by the distribution of drift containing marine shells as revealed by the patient researches of J. Smith[13]. The abundance of shells in this drift shows that there was a period of deglaciation during which the sea occupied the Firth of Clyde, followed by a re-advance during which the ice traversed the sea floor picking up shelly sediments to distribute them widely in central Ayrshire. The limit of the shelly drift coincides very closely with the re-advance limit in this area shown in Figure 1.

The evidence for a re-advance, considered so far, all relates to ice that emanated from the Highlands. One would expect that such an important re-advance would be represented in the Southern Uplands. This expectation is realised and may be demonstrated by three types of evidence.

(i) The ice limit that attains an altitude of over 1,000 feet on the hill slopes south-east of Darvel and is associated with marine shells can be followed southwards towards Muirkirk. However, south of a line passing near New Cumnock and Maybole (and thence a short distance inland past Girvan and Ballantrae) the shelly drift is completely absent and only Southern Upland drift characterised by Loch Doon erratics has been found. Furthermore, "throughout the area of their distribution, the shells are found at all levels up to 1,061 feet", whereas within the area of Southern Upland drift "shells are uniformly absent even in extensive valley bottoms cut below the 600-foot level."[14] This distribution means that, at the time the shelly drift was deposited by the Highland ice, Southern Upland ice occupied the region south of the Maybole to New Cumnock line. This conclusion was expressed in the North Ayrshire memoir of the Geological Survey, where the junction of the two types of drift and the associated ice-movements are clearly illustrated[15].

(ii) As mentioned above, the fluvioglacial features associated with the ice limit along the eastern side of the Clyde basin merge into the Carstairs esker system. This feature has been the subject of much dispute. J. Geikie[16] and Gregory[17] interpreted it as formed along the margin of Southern Upland ice. Charlesworth[18] concluded it was formed along the margin of Highland ice. Macgregor[19], however, recognised that the various ridges comprising the feature have ice-contact slopes on both sides and consequently were formed between ice walls. The feature is, in fact, a complex esker[20] and was formed by a braided stream system flowing east-north-east and north-east along its length. It continues the trend of the abundant fluvioglacial deposits of the Douglas valley (*i.e.*, south-west to north-east) and parallels other fluvioglacial systems in the neighbouring parts of the Southern Uplands. Since large fluvioglacial systems trend parallel with the ice-streams with which they were associated, ice movement towards the north-east and east-north-east is indicated. This accords with striae pointing east-north-east at Carnwath and Lanark[21]. Furthermore, Macgregor[22] stated that, with very few exceptions,

the boulders of the Carstairs esker-system were derived from the Clyde and Douglas valleys, while Ross[23] said that the till at Bonnington (one and a half miles south of Lanark) apparently represents the ground moraine of the Douglas valley-ice from the south-west. An expansion of Southern Upland ice into the Lanark-Carstairs area is thus indicated. It has been stated already that the Carstairs esker-system merges into the deposits associated with the ice that advanced up the Clyde, and that the latter ice was itself confluent with ice that moved through the Darvel-Strathaven gap. Thus three ice streams—from north-west, west, and south-west merged together in the Lanark-Strathaven area.

(iii) Further evidence of a considerable area of ice in the Southern Uplands at the time of the Perth Re-advance is provided by the Clyde area between Carstairs and the Biggar gap. This evidence has been discussed elsewhere[24] and will therefore be stated briefly here. The presence of ice in the Clyde coalfield basin, as described above, clearly means that meltwaters and land drainage from the upper Clyde basin could not have escaped along the lower Clyde valley. The obvious escape-route is the great gap between Clyde and Tweed at Biggar, and Charlesworth[25] postulated an ice-dammed lake overflowing through this gap. However, the floor of the Biggar gap is at 670 feet, whereas there is clear evidence that the ice-dammed lake stood at 700 feet for a considerable time. Ice-contact forms in the gap show that this 30-foot discrepancy is not a result of erosion of the floor of the gap by the escaping lake-waters. The gap must therefore have been blocked by ice when the lake existed. Erratics, striae and fluvioglacial features show that this ice came from the Southern Uplands[26].

Stratigraphy The glacial sequence of lowland Ayrshire and especially that of the Clyde coalfield basin provide excellent evidence of a re-advance of the ice. In an area extending from the foot of the Campsie Hills southwards to beyond Lanark and including Airdrie, Wishaw and part of Glasgow, thick stratified deposits beneath till are of widespread occurrence[27]. The stratified deposits in places exceed 100 feet in thickness and, although gravels are present in places, consist largely of sand with beds of laminated clay. They therefore prove the existence of a lake which, as Bell[28] and J. Geikie[29] pointed out, was inevitably produced by the ice advancing up the Clyde valley past Glasgow. As the ice advanced into the lake it often severely contorted the upper layers of the lacustrine deposits and sometimes incorporated masses of them in the till it laid down. In places, however, the lake deposits were frozen solid before the ice reached them, this being shown by well-preserved fossil frost-wedges at a few points[30]. The lake was extinguished when the ice advancing into the Clyde basin became confluent with the Southern Upland ice and with the ice that moved through the Darvel-Strathaven gap[31]. Significantly, the stratigraphic evidence proving the re-advance appears not to have been recorded east of the limit, based on morphological evidence, that is shown on the eastern side of the Clyde basin in Figure 1.

At several points the lacustrine deposits associated with the re-advance have yielded organic remains and occasionally such remains have been found

in the overlying till. At Chapelhall, near Airdrie, peat, twigs and branches were found in sands and laminated clays beneath 114 feet of till[32]. In the same locality bones of a mammoth were obtained from the sand beneath the till[33]. Near Carluke four feet of moss was found beneath forty feet of "surface clay"[34], and in the till near Carluke a reindeer antler was discovered[35]. At Burnhead Quarry, two miles east of Airdrie, peat in till was almost entirely devoid of tree remains and contained *Betula nana*, now restricted to the higher ground of the Highlands[36]. At Faskine, near Airdrie, another piece of transported peat in till contained *Salix herbacea*, which also is now restricted to high mountains in Britain[37]. Near Glasgow a mammoth tooth[38] and reindeer remains[39] have been obtained from stratified deposits beneath till. This evidence, apart from providing further proof of a re-advance, implies that the re-advance was preceded by a cool interstadial, not by a warm interglacial. A recent examination of the beetles in a sample of peat from the Burnhead site[40] supports this interpretation[41].

In Ayrshire the principal stratigraphic evidence of a re-advance (apart from the important shelly drift) is the stratified deposits overlain by as much as 76 feet of till in the Kilmaurs area near Kilmarnock. These stratified deposits have provided mammoth tusks and reindeer antlers, as well as marine fossils indicative of arctic or sub-arctic conditions[42]. In this area also therefore, a re-advance following a cool interstadial is indicated.

THE SIGNIFICANCE OF THE PERTH RE-ADVANCE

The limit of the Perth Re-advance as now defined permits the resolution of certain aspects of the glaciation of central Scotland that have formerly seemed anomalous. It also permits the integration of certain pieces of evidence that have formerly seemed unrelated.

(i) It has been known for almost a century that the greater part of the late-glacial marine clays of the Forth and Tay areas are far more arctic than those of the Clyde and this difference has given rise to considerable discussion[43]. While the difference may be partly due to differences in sea temperature between the two areas such as exist at present owing to the general oceanic circulation, this explanation alone is recognised as inadequate. The more arctic character of the greater part of the Forth and Tay clays is now seen to be the result of their having accumulated before and during the Perth Re-advance, whereas the Clyde marine clays could not begin to accumulate until after the Perth Re-advance when the climate had ameliorated considerably.

(ii) The arctic marine or estuarine clays of the Forth, Earn, Tay and South Esk are approximately of equivalent age, for they are all overlain by the coarse deposits associated with the Perth Re-advance.

(iii) The highest raised-beaches in both the Forth and Clyde areas have often been referred to in the literature as "the 100-foot beach"[44]. South of the Forth between Falkirk and Stirling extensive areas of late-glacial raised beach lie *above* 100 feet, attaining a maximum altitude of between 120 and 125 feet. In the Glasgow area the highest late-glacial beaches are everywhere

below 100 feet[45]: shorelines at 73·5 feet O.D. near Paisley[46] and 82 feet O.D. at Garscadden[47] have been mentioned as maxima at specific points. There is thus a discrepancy of some 45 feet in about twenty miles. This discrepancy cannot be explained by isostatic warping, for the Forth raised-shorelines rise westwards along the east-west part of the Forth (Dunbar-Larbert), and the fact that the highest post-glacial shoreline continues to rise to the head of the Forth carse-lands near Gartmore (due north of Garscadden) suggests that the isostatic rise up the east-west part of the Forth is unlikely to be reversed before the Glasgow area. This, in fact, suggests that the figure of 45 feet tends to minimise the significance of the discrepancy between the highest raised shorelines in the Forth and Clyde areas. The discrepancy is clearly a result of the ice of the Perth Re-advance having occupied the Clyde area at the time the highest raised-beaches were being formed in the middle Forth valley. It follows that the so-called "100-foot beach" of the literature includes shorelines of markedly different ages. The term "100-foot beach" is thus misleading and should be abandoned.

(iv) Simpson[48] concluded from a section in the Earn valley near Crieff that the ice had retreated at least as far as Crieff before it advanced again, and from a section near Perth estimated the time involved as at least 640 years. The evidence in the Clyde area allows a considerably greater retreat to be deduced.

Beneath the drifts of the Clyde coalfield basin are numerous buried channels. Many of these are deep, steep-sided valleys with recorded depths of 300 feet in the Clyde channel[49] and at least 170 feet in the Avon channel[50]. The channels are clearly river-cut and some have been traced for several miles. The infilling of many of the channels is similar: the greater part is occupied by the deposits of the lake held up by the advancing ice and on top of this is the till of the re-advance. The absence of a basal till from these channels shows that they were cut (or, at least, cleared out) by streams in the interstadial before the Perth Re-advance[51]. For such a drainage system to have existed the Clyde estuary below Glasgow must have been freed of glacier ice.

A considerable deglaciation is also indicated by the marine fossils recorded in the till. Shells have been obtained from till near Airdrie[52] and ten samples of till from different parts of the Glasgow area all yielded foraminifera[53]. In central Ayrshire the widespread shelly drift has been referred to already, and J. Smith[54] reported that a sample of it yielded 2,300 specimens of one species of foraminifera. These records imply that in the interstadial the sea occupied the Firth of Clyde and the Clyde estuary[55]. Therefore during the Perth Re-advance the ice moved forward at least fifty miles in the Clyde area.

(v) The Perth Re-advance appears to mark the last phase of really severe cold in Scotland. With the melting of the ice of this stage the seas became less cold and the highly arctic fauna disappeared. On land the locations of the few finds of mammoth and woolly rhinoceros suggest that these animals became extinct with the climatic amelioration that followed the Perth phase. The clarity of the fluvioglacial features also implies that they have not been modified by severe solifluction.

(vi) Charlesworth omitted Simpson's Perth Re-advance from his map of re-advance limits in the British Isles[56]. In southern Scotland he marked the "Lammermuir-Stranraer re-advance" limit described in the introduction to Part I of this paper. The present writer[57] has shown that the deposits of much of the eastern and central parts of the "Lammermuir-Stranraer moraine" do not mark the limit of a re-advance. Thus the western part of Charlesworth's re-advance limit is left without a continuation. As shown in Figure 1, Simpson's Perth Re-advance has been extended to the borders of the Southern Uplands and, likewise, is left without a continuation. It is not improbable that the two represent the same re-advance.

[1] It may seem that this interpretation ignores the 'Polmont kame', which extends from Falkirk past Polmont towards Linlithgow, and has been interpreted as an important marginal feature by Gregory (*Trans. geol. Soc. Glasg.*, 1913, 14: 199-218) and Charlesworth (*Trans. roy. Soc. Edinb.*, 1926a, 55: 25-50). However, Cowie (*Trans. geol. Soc. Glasg.*, 1913, 14: 360) suggested the feature was formed by a subglacial river flowing west to east. This latter interpretation is fully justified, for the feature is steep on both sides, sinuous, has an undulating crest-line, and in places divides into more than one ridge: it is thus a typical esker.

[2] Peach, A. M. *Geol. Mag.*, 1909, 46: 26-31. Clough, C. T., *et al. The geology of the Glasgow district*, Mem. geol. Surv., 1925.

[3] *Summ. Progr. geol. Surv.*, for 1906 (1907).

[4] Peach, A. M. *Op. cit.*

[5] Upham, W. *Amer. Geol.*, 1898, 21: 235-243; Clough, C. T., *et al. Op. cit.*; Elder, S., *et al. Trans. geol. Soc. Glasg.*, 1935, 19: 285-287.

[6] Young, J. *Trans. geol. Soc. Glasg.*, 1871, 3: 298-310; Bell, D., *Ibid.*, 1874a, 4: 300-310; Elder, S., *et. al. Op. cit.*

[7] Hopkins, W. *Quart. J. geol. Soc. Lond.*, 1852, 8: 20-36; Bell, D. *Op. cit.*; *Idem. Trans. geol. Soc. Glasg.*, 1896, 10:16-24; Clough, C. T., *et. al. Op. cit.*

[8] Mitchell, G. F. *Advanc. Sci.*, 1960, 17: 313-325.

[9] Few of these abundant sand and gravel deposits are shown on Geological Survey One-Inch sheets 23 and 31. They were observed by Charlesworth (*op. cit.*), however.

[10] Richey, J. E., *et al. The geology of North Ayrshire.* Mem. geol. Surv., 1930: 336-338.

[11] *Ibid.*, pp. 332-336.

[12] *Summ. Progr. geol. Surv.*, for 1925 (1926).

[13] Smith, J. *Trans. geol. Soc. Glasg.*, Supp., 1898, 11:134 pp.

[14] Richey, J. E., *et al. Op. cit.*, p. 321.

[15] *Ibid.*, Fig. 32, p. 321. In the Survey memoir these ice-movements are considered to relate, not to the last occupation of the area by ice, but to an earlier period. This conclusion is not in accord with that of Charlesworth (*op. cit.*). It is also opposed to Smith's many descriptions of the shelly drift as the upper drift: significantly, this important point is not mentioned in the Memoir. To a large extent the view put forward in the Memoir depends on the interpretation of crossing "drumlins" in the Kilmarnock area, it being concluded that a west-east set of drumlins was formed before a north-north-east to south-south-west set. As pointed out in the Memoir (p. 323) the latter are mostly only 100 to 200 yards long. These dimensions are exceptionally small for drumlins, and mapping and examination of them leads one to suspect they are not, in fact, drumlins. Their true origin is uncertain, but they may well represent subglacial stream courses into which till was squeezed by ice pressure following their abandonment by meltwaters, thus resembling features described in Canada by Gravenor and Kupsch (*J. of Geol.*, 1959, 67: 48-64) and Stalker (*Geol. Surv. Canada, Bull.*, 57, 1960).

[16] Geikie, J. *The Great Ice Age*, 3rd ed., 1894: 187.

[17] Gregory, J. W. *Op. cit., Idem. S.G.M.*, 1915, 31: 465-476.

[18] Charlesworth, J. K. *Op. cit.*

[19] Macgregor, M. *Proc. Geol. Assoc.*, 1927, 38: 495-499.

[20] Sissons, J. B. *Geol. Mag.*, 1961, 98: 380-392.

[21] Smith, J. *The drift or glacial formation of the Clyde drainage area*, handbook for Brit. Assoc., Glasg., 1901: 520-527.

[22] Macgregor, M. *Op. cit.*

[23] *Summ. Progr. geol. Surv.*, for 1926 (1927).

[24] Sissons, J. B. *Trans. Edinb. geol. Soc.*, 1961, 18: 175-193.

[25] Charlesworth, J. K. *Op. cit.*

[26] The lake overflowed near Garvald House, east of Dunsyre.

[27] Bell, D. *Trans. geol. Soc. Glasg.*, 1874b, 4: 63-69; Bennie, J. *Ibid.*, 1871, 3: 133-148; Dron, R. W. *Ibid.*, 1916, 15: 1-3; Dick, R. *Trans. Edinb. geol. Soc.*, 1870, 1: 345-352; Geikie, J. *Op. cit. Summ. Progr. geol. Surv.*, for 1906 (1907); *Ibid.*, for 1907 (1908); *Ibid.*, for 1911 (1912); *Ibid.*, for 1926 (1927); Clough, C. T., *et al. Op. cit.*; *Idem.*, *The economic geology of the central coalfield of Scotland, Area VIII*, 1920.
[28] Bell, D. *Op. cit.* 1874b.
[29] Geikie, J. *Op. cit.*
[30] Galloway, R. W. *Biul. Peryglac.*, 1961, 10: 169-193.
[31] With the subsequent decay of the ice in the Clyde area lakes re-formed in places and their deposits overlie the till laid down during the re-advance.
[32] Geikie, A. *Trans. geol. Soc. Glasgow.*, 1863, 1: 1-190.
[33] *Ibid.* Smith, J. *Op. cit.*, 1901.
[34] Hunter, J. R. S. *Trans. geol. Soc. Glasg.*, 1882, 6: 84-94.
[35] Young, J. and Craig, R. *Ibid.*, 1871, 3: 310-321.
[36] Dunlop, R. *Ibid.*, 1888, 8: 312-315.
[37] Bennie, J. *Ibid.*, 1896, 10: 148-152.
[38] *Trans. geol. Soc. Glasg.*, 1882, 6: 291.
[39] Macgregor, M. and Ritchie, J. *Proc. roy. Soc. Edinb.*, 1940, 60: 322-332.
[40] Coope, G. R. *Trans. geol. Soc. Glasg.*, 1962, 24: 279-286.
[41] The only site that appears not to accord with this interpretation is at Cowdon Glen, ten miles south-west of Glasgow. Here plant and animal remains of temperate species occur in stratified beds. According to one view these beds are covered by till, but according to another the overlying "till" is material that has slipped from the steep sides of the valley in which the deposits occur. The latter interpretation seems the more likely and was given in the original detailed description of the site by R. Craig (*Ibid.*, 1874, 4: 17-32).
[42] Bryce, J. *Quart. J. geol. Soc., Lond.*, 1865, 21: 213-218; Young, J. and Craig, R. *Op. cit*; Craig R. *Trans. geol. Soc. Glasg.*, 1888, 8: 213-223.
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[44] *e.g.*, Clough, C. T., *et al. Op. cit.*, 1925; Dinham, C. H. and Haldane, D., *The economic geology of the Stirling and Clackmannan coalfield*, Mem. geol. Surv., 1932.
[45] Clough, C. T., *et al. Op. cit.*, 1925.
[46] Blair, M. *Trans. geol. Soc. Glasg.*, 1893, 9: 139-150.
[47] Mitchell, G. F. *New Phytol.*, 1952, 50: 277-286.
[48] Simpson, J. B. *Trans. roy. Soc. Edinb.*, 1933, 57: 633-645.
[49] Stark, J. *Trans. geol. Soc. Glasg.*, 1902, 12: 52-57.
[50] Clough, C. T., *et al. Op. cit.*, 1920.
[51] It is not intended to suggest that every buried channel in the Clyde basin is of this age and origin. In particular, the deep buried troughs extending below sea-level probably owe much to erosion by the ice itself.
[52] Geikie, A. *Op. cit.*, p. 58.
[53] Wright, J. *Trans. geol. Soc. Glasg.*, 1896, 10: 263-272.
[54] Smith, J. *Op. cit.*, 1898.
[55] The marine deposits at Kilmaurs show that the interstadial sea here attained at least 100 feet above present sea-level.
[56] Charlesworth, J. K. *Trans. roy. Soc. Edinb.*, 1955, 62: 769-928.
[57] Sissons, J. B. *Op. cit.*

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Peat bogs in a Post-glacial sea and a buried raised beach in the western part of the Carse of Stirling

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Synopsis

Borehole evidence shows that in the western part of the Carse of Stirling there occurs a buried raised beach whose surface is between about 30 and 36 ft O.D. The beach appears to have been formed after about 10 300 B.P. and before about 9500 B.P., more probably around the latter date. The buried raised beach is normally overlain by peat and this in turn by carse clay. Over part of the area the carse clay is covered by surface peat. In two areas the carse clay is absent and thick peat occurs instead. Part of this thick peat accumulated as the carse clay was being deposited around it.

Introduction

The Carse of Stirling extends westwards from Stirling for fifteen miles, with a width of up to four miles. For the most part it is an almost flat plain below 50 ft O.D. composed of estuarine muds—the ' carse clay '—laid down in a post-glacial sea and subsequently elevated and slightly tilted eastwards owing to isostatic recovery from glacial loading. The western part of the plain is interrupted by the Menteith moraine, which forms a conspicuous ridge system rising to over 100 ft O.D. The moraine is breached by several gaps cut by meltwaters, the largest of which is utilised by the Forth.

A considerable part of the Carse of Stirling was once covered by peat. Much of this has been cleared by man over the centuries, but in the western part of the region large areas of peat still remain (Fig. 1). The principal peat mosses are known as East Flanders Moss, West Flanders Moss and Gartrenich Moss. Along with smaller mosses they form almost featureless wastes amidst the farmlands on the carse clay, although part of Gartrenich Moss was recently afforested while West Flanders Moss is now being prepared for afforestation.

It has long been known that the mosses of the Carse of Stirling are normally underlain by carse clay. This has been demonstrated for large parts of the three principal bogs that now remain by the Scottish Peat Survey (1962), which has put down a large number of bores through the mosses. The Peat Survey bores, however, also revealed the interesting fact that in certain areas the carse clay is quite absent and is replaced for its full thickness by peat. These anomalous areas are clearly revealed by the unpublished Peat Survey maps showing contours of the base of the peat at vertical intervals of one metre (West Flanders Moss) or half a metre (East Flanders Moss). The contours represent the peat as occupying steep-sided closed depressions in the carse clay, the East Flanders depression having dimensions of about a mile by three-quarters of a mile, and that of West Flanders about one mile by one and a quarter miles. While the drawing of these

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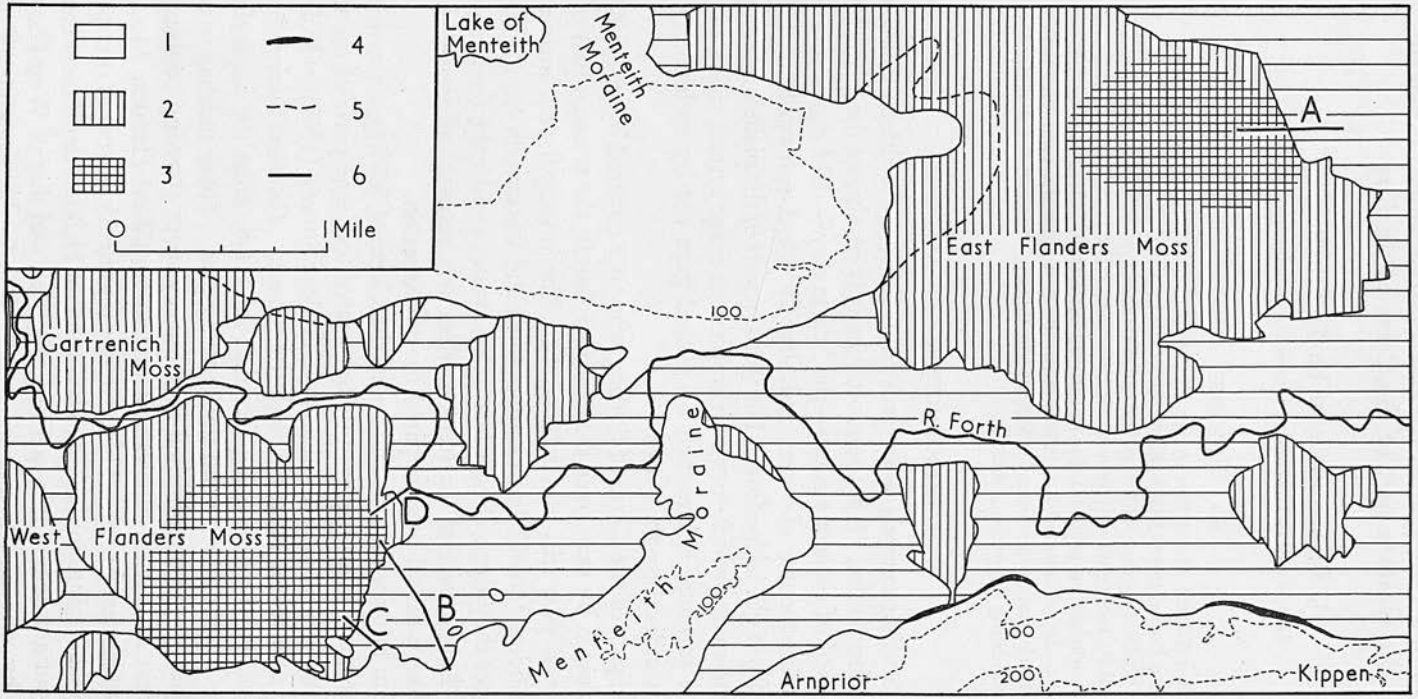


FIG. 1. The Carse of Stirling in the vicinity of the Menteith moraine. 1, Carse clay. 2, Main areas of surface peat, mostly underlain by carse clay. 3, Areas where carse clay is absent and thick peat occurs instead (based mainly on boreholes made by the Peat Survey). 4, Narrow strips of surface peat along the margin of the carse plain. 5, Margin of concealed carse where significantly different from edge of peat. 6, Lines of section, lettered as in Fig. 2.

depressions on the Peat Survey maps represents an accurate and valid interpretation of the Peat Survey borehole evidence, it raises geomorphological difficulties. Since the carse clay is essentially uplifted mud-flats the development of large closed depressions does not seem possible. Hence the writers put down lines of closely spaced boreholes across the margins of the depressions to examine the relationships between carse and peat in the hope of resolving this anomaly. A Hiller-type peat sampler was used and the boreholes were accurately levelled.

Main Elements of the Stratigraphy

The locations of the lines of borings are indicated in Fig. 1 and sections along these lines are plotted in Fig. 2. The main features of the stratigraphy are similar in all the sections.

The lowest deposits usually encountered were a grey silty fine sand or grey clay; where both these deposits occurred the latter overlay the former. With increasing depth the clay begins to alternate with the silty sand, the bands of the latter becoming thicker until it comprises the whole deposit. The surface of these grey deposits is almost flat and it was on this surface that peat began to accumulate.

Where overlain by carse clay the peat is normally 2 to 3 ft thick, but as the carse clay thins out at the margins of the deep mosses the peat increases in thickness to 3 to 5 ft. The peat merges upwards into the overlying carse clay, the transition zone usually varying between 6 in and a foot in thickness.

The carse clay is usually between 8 and 11 ft thick in the areas studied, but as the deep mosses are approached it rapidly tapers out owing to the altitude of its base rising and the altitude of its surface falling. Where the carse clay begins to thin out rapidly reeds become common and the clay merges into a deposit that is almost entirely reeds with minor amounts of clay.

The present extent of the peat on the ground surface is artificial. The surface of the higher parts of the mosses shown in the sections (above about 50 ft O.D.) represents the natural surface of the peat. Elsewhere the peat has been either removed completely to expose the carse clay or partly removed.

The Buried Raised Beach

The top of the grey clay and silty sand underlying the peat is flat or very gently inclined. In section A (Fig. 2) it ranges between 31·1 and 32·5 ft O.D. in four boreholes spaced over a distance of nearly half a mile. In section B nine boreholes give its height as between 32·5 and 34·3 ft, in section C seven boreholes show a height of between 32·8 and 36·5 ft, and in section D the height varies between 29·1 and 32·1 ft in six boreholes. The Peat Survey bores show that a surface between about 30 and 36 ft O.D. prevails beneath almost the whole area of the two deep mosses (Fig. 1). Such an extensive flat or gently inclined feature composed of fine-grained water-laid sediments can only be interpreted in this area as a raised

beach. Jamieson (1865, Fig. 9) appears to have envisaged the existence of a buried beach, of late-glacial age, and Durno (1958) suggested that the deep peat of East Flanders Moss rests on late-glacial marine deposits.

A late-glacial age is suggested by the character of the deposit, which in part resembles glacial rock flour, and by the apparent absence of visible plant remains (except for the uppermost part in association with the overlying peat). The deposit contrasts greatly with the carse clay, which contains abundant vegetable matter. Durno's (1956) pollen analyses of the deep peat overlying the buried beach in East Flanders Moss show that the lowest metre of the peat in the column he studied belongs to Zone V of the British pollen sequence. This zone has been radiocarbon-dated by Godwin (1961) as extending from about 9550 to about 9000 B.P. In view of the considerable thickness of Zone V peat on the buried beach it seems probable that the beach is at least as old as the early part of Zone V (i.e. probably around 9500 B.P.). On the other hand, the buried beach is clearly younger than the Menteith moraine for it occurs both inside and outside this moraine. Donner has dated the moraine by pollen analysis as of Zone III age (*c.* 10 800 to 10 300 B.P.). Although Donner's dating is based in part on negative evidence, there is at present no reason to doubt his conclusion. Thus it appears that the buried raised beach was formed not earlier than about 10 300 and not later than around 9500 B.P. Since the buried beach appears to be composed of glacio-marine deposits it could have been formed during the deglaciation that followed the formation of the Menteith moraine or in association with a later glacial readvance.

Borehole lines B and C were in part located to determine the position and altitude of the buried shoreline. In section B, borehole 13 shows the buried peat resting on fluvioglacial sand and gravel of the Menteith moraine complex. In borehole 14, however, the grey silty sand, 2·4 ft thick, intervenes between the peat and the fluvioglacial deposits. The buried shoreline thus lies between these two bores and its altitude is probably slightly above 34·3 ft O.D. (this being the altitude of the top of the buried beach deposits in borehole 14). In section C, borehole 24 was put down about 25 yd from the edge of the carse plain. Beneath the peat the grey clay and silty sand of the buried beach were only 1·0 ft thick and rested on sand and gravel of the moraine complex. The surface of the buried beach deposits in this borehole was at 36·6 ft O.D. It thus appears that in the area around sections B and C the buried shoreline lies quite close to the carse shoreline and some 10 to 13 ft lower.

Peat Bogs in the Carse Sea

The peat on the buried raised beach implies that, after the beach was formed, there was a relative fall of sea-level that exposed the beach as a land surface. How far the sea-level fell is at present unknown. At a later date sea-level rose

and the peat was covered by carse clay, except in the area of the two deep mosses.

All four sections show that, away from the margins of the deep mosses, the buried peat layer is of fairly constant thickness, increasing only slightly, if at all, towards the deep mosses. This suggests that the transgression of the post-glacial sea over this part of the peat was fairly rapid. Around the margins of the deep mosses, over a distance of about 150 to 200 yd the peat increases considerably in thickness, the base of the carse clay rising, until this clay tapers out. The point at which the carse clay ends is between about 37 and 40 ft O.D. in all four sections. It thus appears that the rate of encroachment of the sea on the accumulating peat slowed down until, by the time the carse clay had built up to an altitude of between 37 and 40 ft the sea was unable to encroach further. Thereafter, the shoreline of the carse sea retreated as the area of peat growth extended at the expense of the still-accumulating carse clay. When the sea finally withdrew completely peat growth probably occurred over the whole area. A sample from 4 in above the base of this peat, overlying carse clay in East Flanders Moss, has been radiocarbon-dated as 5492±130 B.P. (Godwin and Willis 1962).

Durno (1958) inferred from the absence of carse clay in the deep part of East Flanders Moss that the transgression that laid down the carse clay " did not reach as far west . . . as the present-day Flanders Moss ". This suggestion is clearly at variance with the extent of the carse clay as shown in Fig. 1 and with the borehole evidence of the Peat Survey, which shows that carse clay, at its usual altitude (for this area) of between 44 and 48 ft O.D., underlies large areas of both East and West Flanders Mosses. It is clear from Fig. 1 that the deep peat of East Flanders Moss is completely surrounded by the carse clay, while the deep peat of West Flanders Moss is encompassed by the carse clay except where it adjoins the slopes of the Menteith moraine on its southern side.

It is not known what factors determined the precise limits of the deep peat bogs. On the other hand, reasons can be suggested for the restriction of these deep mosses to the western part of the Carse of Stirling (the mosses further east in the Forth valley always overlie the carse clay). Probably of considerable importance was the tidal situation in the area when the carse clay was being deposited. Measurements by the writers of the modern mud flats near Grangemouth and elsewhere in the Forth area show that these have normally accumulated to between 3 and 4 ft below high water mark of ordinary spring tides at their landward edges. One may infer that the carselands were similarly related to H.W.M.O.S.T. This would mean that for the greater part of each semi-diurnal tidal cycle the carse surface (at any point in time) would not be covered by the sea. This effect would be emphasised at the western end of the Carse of Stirling, for the Carse of Stirling was a long, almost land-locked estuary at the time it was occupied by the post-glacial sea. Conditions were doubtless comparable to those of some British estuaries today, towards the head of which mud flats are covered

by the sea for only a short time at high water springs and sometimes not at all at high water neaps. Such conditions may well have favoured peat accumulation in the western part of the Carse of Stirling.

The rate of isostatic recovery may also have been a factor in assisting the peat to maintain itself against the encroaching sea. Isostatic uplift has been greatest in the Forth region at the western end of the Carse of Stirling (Sissons 1963): hence the retardation of the rising sea-level by isostatic recovery would be more significant in this area than elsewhere.

As noted above, Durno has shown that the lowest metre or so of the peat of East Flanders Moss accumulated in Zone V, that is, before about 9000 B.P. On the other hand radiocarbon assay of the top of a thin layer of sub-carse peat at about 21 ft O.D. at Airth Colliery site a few miles south-east of Stirling gave a date of 8421 ± 157 B.P. (Godwin and Willis 1961). Very similar radiocarbon dates were obtained for two samples of sub-carse peat from the Earn valley. Since these radiocarbon dates are believed to establish approximately the date of the transgression of the carse sea over the peat at the sites they represent, it follows that the transgression in the East Flanders Moss area (further inland, nearer the centre of isostatic recovery, and at a higher altitude then and now) could not have occurred before about 8400 B.P. Hence a considerable accumulation of peat must have taken place in the East Flanders Moss area before the transgression occurred there. It may well be that this peat accumulation was favoured by the almost flat surface of the buried raised beach on which it rests. If this peat growth commenced as the sea abandoned the buried raised beach immediately after the beach was formed, as seems likely, it would seem that the date of this buried feature is around 9500 B.P.

Durno's pollen diagram for East Flanders Moss (1956, Fig. 3c) shows a marked diminution in heaths and a sudden increase in bog moss about half way up Zone VI (i.e. roughly between about 8500 and 8000 B.P. if the rate of peat accumulation was fairly constant). It seems very likely that this sharp change in the type of peat resulted from the marine transgression that encompassed the bog. At Durno's pollen site this occurred when the peat had accumulated to a depth of about 5 ft (ignoring compaction).

It is apparent from Fig. 1 that in certain areas the clearance of surface peat by man has been halted by the deep mosses. Clearance has progressed furthest in this respect along the margin of West Flanders Moss in the vicinity of section B. Here the surface of the carse can be seen to descend sharply towards the unreclaimed moss, the margin of which is an artificial depression. As can be seen in Fig. 2 (section B) the peat has been almost completely stripped from the sloping upper surface of the rapidly thinning carse wedge, clearance having stopped within 5 yd of the point where the carse clay ends. The line of contact between carse and peat is here accurately marked by the sudden change from cultivated ground to water-logged peat. In the vicinity of sections A and

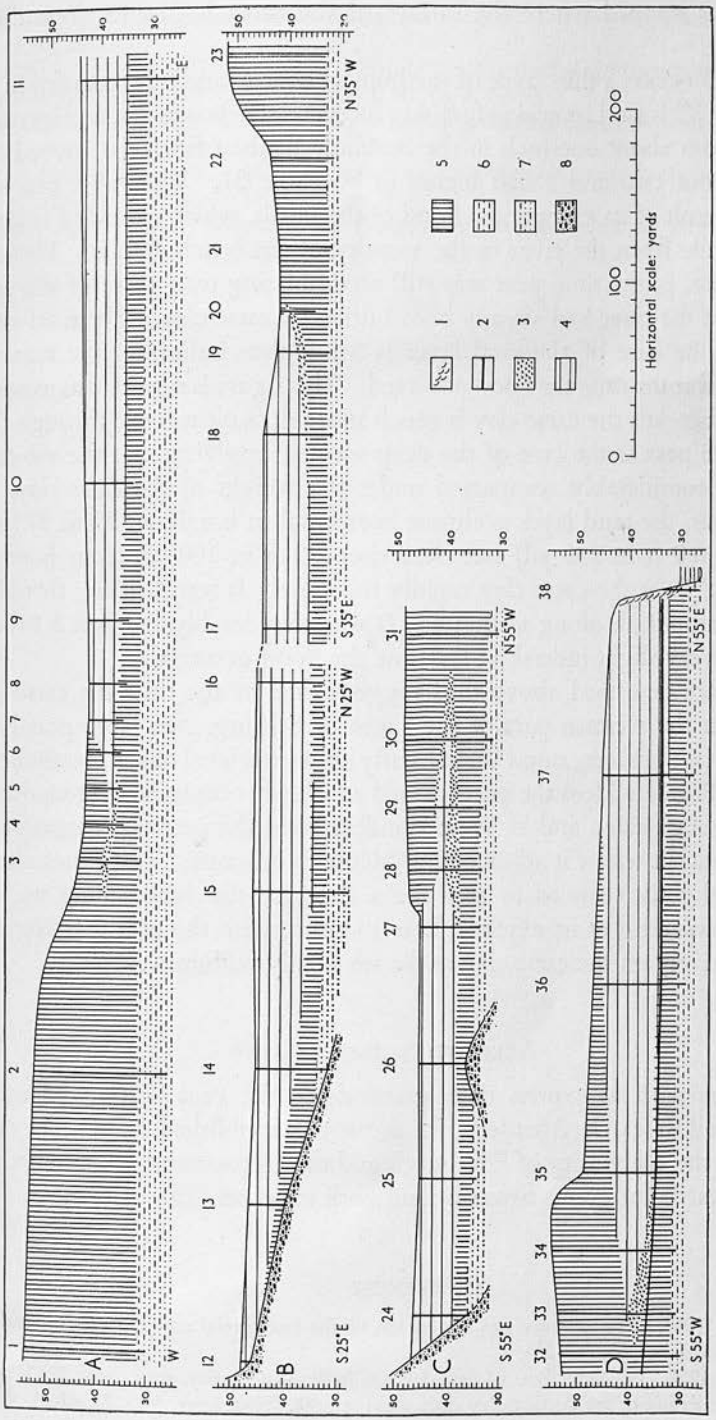


FIG. 2. Sections along the lines A to D in Fig. 1. 1, Slumped material in bank of River Forth. 2, Thin band of medium and coarse sand in section D. 3, Reeds with a little carse clay. 4, Carse clay. 5, Peat. 6, Grey clay. 7, Grey silty fine sand. 8, Fluvioglacial sand and gravel. The break in section B represents a distance of 430 yd.

C clearance has stopped where the surface of the carse begins to slope down sharply.

In section D occurs a thin layer of medium to coarse sand not recorded in the other sections. This sand layer was found in six successive boreholes, and increased in thickness from about one inch in the borehole furthest from the River Forth (no. 32) to about two and a half inches in borehole 37. This layer can only represent the result of an exceptional flood of the Forth, which extended at least a quarter of a mile from the river in the vicinity of the borehole line. The sand layer shows that, at the time peat was still accumulating over most of this area, the peat nearest the river had already been buried by carse clay. The total range of altitude of the base of the sand layer is 3·1 ft thus indicating the maximal relief of the area at the time the flood occurred. This figure is regarded as maximal since the peat beneath the carse clay is much more difficult to bore through than the comparable peat at the base of the deep mosses, implying that the sub-carse peat has been considerably compacted under the weight of the carse clay. In accord with this, the sand layer is almost horizontal in boreholes 35 to 37 (total altitude range 0·9 ft in 230 yd) but then rises 2·3 ft in 100 yd from borehole 35 to borehole 32, as the carse clay rapidly thins out. It seems likely, therefore, that the relief amplitude along section line D was considerably less than 3·1 ft and was probably very slight indeed, at the time the flood occurred.

The evidence described above thus suggests that, at the time the carse clay accumulated in the western part of the Carse of Stirling, two large peat bogs, fringed with a belt of reeds, stood very slightly above the level of the accumulating clay. In East Flanders Moss the peat formed an island completely surrounded by the sea during high tides, and in West Flanders Moss the peat was encompassed by carse clay except where it adjoined the Menteith moraine. The accumulating mosses were slightly reduced in area for a time by the encroaching sea, but subsequently expanded at its expense, later merging with the peat that began to develop extensively on the carse clay as the sea finally withdrew.

Acknowledgements

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SOME PRE-CARSE VALLEYS IN THE FORTH AND TAY BASINS

J. B. SISSONS, R. A. CULLINGFORD and D. E. SMITH

In a recent issue of this magazine J. S. Smith[1] drew attention to certain gullies cut into late-glacial raised beach deposits situated by the Cromarty Firth. He described the gullies in some detail and stated that they are graded to the "25-foot" raised beach. He noted that, despite the considerable size of the gullies, there are no debris cones where they debouch on to the "25-foot" beach and, therefore, justifiably deduced that the gullying pre-dated or occurred contemporaneously with the formation of this beach. However, he subsequently neglected the former possibility and concluded that the gullies and the "25-foot" raised beach were formed at the same time. He also attributed the gullying to "the pluvial conditions of the early Atlantic climatic phase" about 5,000 years ago.

In the Forth and Tay basins gullies similar to those described by Smith, as well as important valley systems, are cut into late-glacial deposits in many localities. In these two areas, however, it can be shown conclusively that the gullies and valleys are older than the "25-foot" raised beach and were, in fact, in existence at least 8,000 to 8,500 years ago.

THE GLENCARSE AREA

On the northern side of the Firth of Tay a well-defined area of raised beach deposits extends just over one mile west and two miles east of Glencarse village. These deposits consist largely of fine sand and form a predominantly flat surface that is dissected by numerous gullies and valleys. The shoreline has been measured at nine points between B and C, Fig. 1, the heights varying between 53·0 and 55·1 feet O.D. That these deposits are of late-glacial age is indicated by the existence, at a lower level, of the laminated brick-clay at Inchcoonans, near Errol, approximately three miles east-north-east of Glencarse. This deposit was investigated by Brown[2] and Davidson[3] and was found to contain a rich shell fauna of a highly arctic character. The original ground surface at the most recently worked pit has been levelled at 39 feet O.D., and the arctic clay is not covered at this height by any post-glacial marine sediments, the upper limit of which is well-marked by the margin of the carse-lands, some 6 or 7 feet lower.

Most of the gullies and valleys around Glencarse are dry, and many of them extend back to the sharp break of slope between the beach deposits and the steep rocky Braes of the Carse. The bottoms of the larger valleys are flat and, when followed downstream, are seen to merge into the extensive flat surface of the carse-lands, the shoreline of which, levelled at 18 points in the map area, is at 32-33 feet O.D. It might be concluded that the cutting of the valleys and gullies was contemporaneous with the deposition of the carse, but this interpretation is shown to be incorrect by borehole evidence.

By Glencarse village is a large flat-floored valley with steep sides 15 to 20 feet high. Part of this valley is utilised by the A85 Perth-Dundee road and by the railway. The valley is joined by numerous gullies but is drained only

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by a small artificial ditch known as the Pow of Glencarse. Nine boreholes have been put down in the floor of the valley, of which two are included in the line marked A on Fig. 1 and are discussed later. Of the other seven, three (nos. 5 to 7) were sunk in connection with a new road bridge and four by the writers.

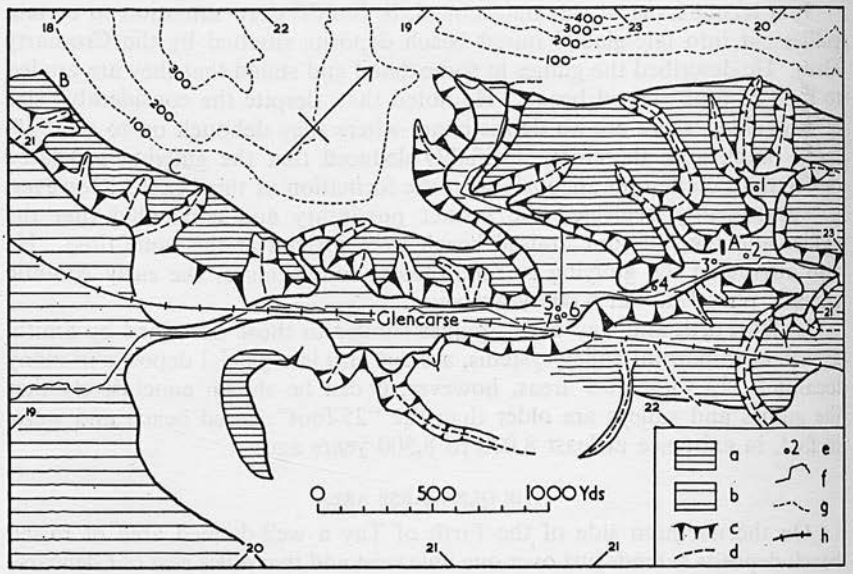


Fig. 1. Valleys and gullies cut in late-glacial raised beach deposits around the village of Glencarse. a. late-glacial raised beach deposits; b. flat-surfaced carse deposits; c. steep sides of valleys and gullies; d. axes of dry valleys; e. position and number of boreholes; f. ditches and streams; g. contours at interval of 100 feet; h. railway. A, B and C referred to in text.

No. of Borehole	1	2	3	4	5	6	7
Nat. Grid Ref.	No. 2041 2290	No. 2057 2268	No. 2051 2257	No. 2042 2234	No. 2023 2201	No. 2020 2202	No. 2022 2199
Ground level (Feet O.D.)	37·6	35·2	33·5	33·8	31·8[1]	27·9[1]	30·9[1]
Surface Peat	5·1	—	—	—	—	—	—
Carse	2·8	12·7	9·6	7·0	14·5	10·8	10·8
Peat	4·7	0·1[2]	4·0	0·6	1·5	1·5	3·3
Fine sand	0·4[3]	0·2[3]	0·6[3]	1·6[3]	21·0	23·8[4]	17·9

Table 1.—Thicknesses of strata (in feet) in boreholes 1 to 7.

Notes: 1. Tops of bores on banks of ditch.
2. Not true peat, but abundant plant remains in silty sand.
3. Base of deposit not reached.
4. Includes 1·8 feet "soft silty clay" immediately below peat.

The stratigraphy of the boreholes was very similar (Table 1), the carse deposits, comprising grey silty clay with plant remains, overlying compact

peat. The only exception was borehole 2, in which the peat was represented by a thin layer of abundant plant remains in grey silty sand. Beneath the peat in all the boreholes came grey silty sand.

In addition to the seven boreholes recorded in Table 1, two boreholes and eleven spiral-auger bores were spaced at close intervals along a line (A, Fig. 1) approximately at right-angles to the axis of the valley. The purpose was to establish the relationship of the deposits found in the valley bottom to the late-glacial beach deposits into which the valleys are cut. The tops of these bores (nos. 8 to 20) were levelled, and a cross-section constructed (Fig. 2). Borehole no. 3 is also included in this cross-section, which extends from the approximate centre of the valley to a point just over half way up the valley side. The cross-section shows clearly that the peat layer, of variable thickness, feathers out against the fine sand, and that the carse clay in turn wedges out against the peat. At the base of the valley-side slope both the peat and the carse are covered by fine sand that has moved downslope. The deposit below the peat is a pale grey to yellowish-grey fine sand, similar in all visible respects to the fine sand seen at the surface all over the area of late-glacial beach deposits. Bore 19 passed through 11 feet of yellowish-grey fine sand, indistinguishable from that beneath the peat except for a brighter yellow colour at the surface. As no deposit intervenes between the fine sand and the peat, it is concluded that the latter rests directly on the floor of a valley cut into the late-glacial beach deposits (i.e. that this valley was not transgressed by the sea until carse times). The three commercial boreholes show that the fine sand is of great thickness, extending down to below Ordnance Datum.

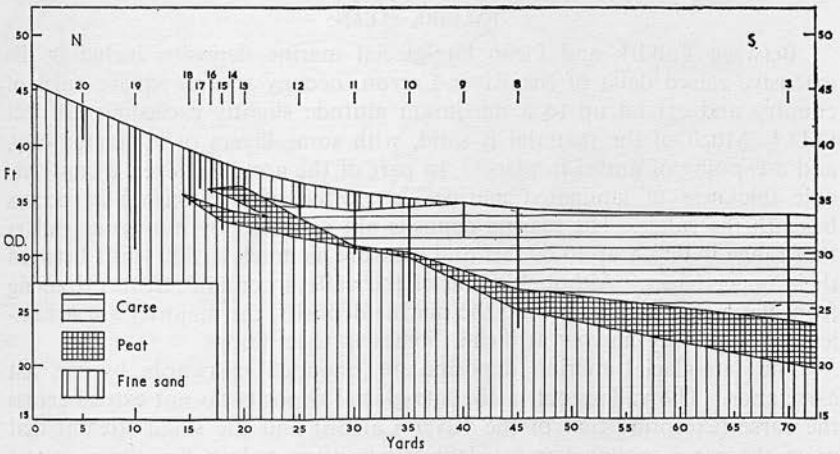


Fig. 2. Cross-section along line A, Fig. 1 (site of boreholes 8 to 20) and through borehole 3.

These deposits, together with the morphology, indicate the following sequence of events:—(i) A raised beach at about 55 ft. O.D. was formed in late-glacial times; (ii) A negative movement of base-level occurred and the beach was dissected by numerous streams; (iii) Peat accumulated on the floors of the larger valleys; (iv) A rise of sea-level converted the larger valleys

into long narrow estuaries in which carse muds accumulated and buried the peat.

POWIS MAINS, NEAR AIRTHREY

On the northern side of the Forth valley around Airthrey, near Stirling, is a large deposit of kettled fluvioglacial sand and gravel whose higher parts exceed 100 feet O.D. Three marked dry valleys are incised into this deposit on its eastern side near the farm of Powis Mains. The largest of these valleys attains a depth of about 30 feet and leads into the carse, whose margin here lies between 39 and 41 feet O.D. The carse extends a short distance into the mouth of this valley, rising slightly as it does so.

Two borings were made in the mouth of the valley. The more southerly bore (Nat. Grid Ref. NS/8156.9584) passed through 6·3 feet of carse clay and then through 2·0 feet of peat, beneath which was red till. The other bore (NS/8156.9591) met with 7·5 feet of carse clay, under which was 1·0 feet of peat, followed by gravel. In the latter bore the carse clay contained a few stones in its upper part. These are probably present owing to soil creep assisted by cultivation of the adjacent valley slopes. Otherwise in both bores the carse deposits were characteristic of the Forth valley generally, comprising a fairly tough surface layer of silty clay overlying a very soft layer of silty clay often referred to as sleech. These fine deposits, comparable in origin to the present mud-flats of the Forth, clearly could not have accumulated when the deep valley was being cut in the sands and gravels. As in the example cited from the Tay area, it appears that the carse and the peat that underlies it accumulated after the valley was excavated.

FALKIRK-PLEAN

Between Falkirk and Plean late-glacial marine deposits, including the extensive raised delta of the River Carron, occupy several square miles of country and extend up to a maximum altitude slightly exceeding 120 feet O.D.[4] Much of the material is sand, with some layers of laminated clay, and a capping of gravel in places. In part of the area, however, a considerable thickness of laminated marine clay, often of a pinkish hue, occurs beneath the sands. The marine deposits are dissected by numerous valleys that range in depth up to 60 feet and measure as much as 300 yards between their upper edges. Although some of the valleys contain streams draining from the higher ground behind the marine deposits, the majority are streamless or have only narrow artificial ditches in their floors.

The late-glacial marine deposits are bounded eastwards by the flat carselands. The valleys cut in the late-glacial deposits do not extend across the carse (excepting that of the River Carron) and the small streams that cross the carse occupy narrow depressions often only a few times greater than their own limited width. It is clear from this, and from the absence of alluvial fans at the mouths of the valleys cut in the late-glacial deposits, that these valleys have not been excavated since the carse was deposited. Furthermore, the carse extends several hundred yards up some of the larger valleys, giving rise to poorly-drained flat floors. A bore put down in one of these flat valley floors near the point at which it opens out into the carse-lands (NS/8565.8490) showed, from the surface downwards, 8·8 feet of carse

deposits, 0·4 feet of peat, 2·1 feet of silt and silty fine sand with numerous plant remains, 2·1 feet of pink laminated clay, and 0·2 feet of coarse sand. The peat layer was also found in two bores further up the valley.

The valley of the River Carron exhibits a similar pattern. Between Falkirk and Larbert the Carron flows in a valley about 70 feet deep and three-quarters of a mile wide cut into the deposits of its late-glacial delta (Fig. 3). On the sides and floor of the valley are several terraces, of which by far the most conspicuous is an almost horizontal feature whose inner margin lies between about 36 and 39 feet O.D. This terrace is the local equivalent of the main level of the carse and is mostly of similar composition, but with a veneer of fluvial sand in places. The terrace fingers into the mouths of several of the valleys tributary to the Carron (e.g. valleys A and C, Fig. 3).

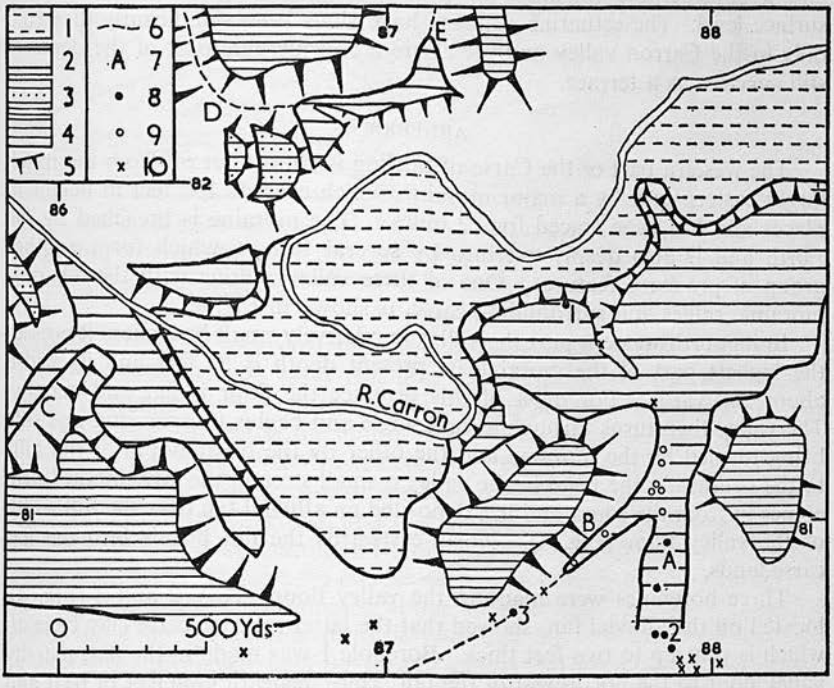


Fig. 3. Landforms of part of the Carron valley near Falkirk. 1. floodplain and lowest post-glacial terraces; 2. main post-glacial terrace; 3. intermediate terraces; 4. remnants of the late-glacial delta of the Carron; 5. bluffs due to dissection of features 2 to 4; 6. axes of some pre-carse valleys; 7. principal pre-carse valleys, lettered A to E; 8. boreholes and trial pits (numbered 1 to 3) in which peat found beneath other deposits; 9. boreholes with peat at the surface; 10. boreholes in which peat not reported. The larger blank areas on the map represent areas where there has been considerable human modification of the landforms, e.g. by the built-up area of part of Camelon and Falkirk in the south and south-east and by the tip-heaps of the Carron ironworks in the north-east.

The relationship of the terrace to the valleys is demonstrated by a series of boreholes and trial pits made in connection with the new Falkirk sewage scheme. Trial pit 1, situated by the outer edge of a large remnant of the terrace, showed 9·5 feet of clay overlying 2 feet of dry peat. In valley A four boreholes and a trial pit showed peat up to 4 feet thick covered by several

feet of sandy clay. In valley B and the adjacent part of valley A eight boreholes showed that peat extends from the surface down to depths as great as 10 feet. Further up valley B a trial pit (no. 3) revealed 5·5 feet of sand and sandy clay overlying a bed of peat one foot thick.

The above evidence demonstrates the following sequence of events. The late-glacial marine deposits of the Falkirk-Plean area were dissected by numerous streams, including the River Carron. Subsequently peat accumulated on the floors of these valleys. A marine transgression then occurred and converted the lower parts of some of the valleys into estuaries. This caused the peat to be buried beneath estuarine muds and beneath fluvial deposits. In some other valleys, such as part of B, Fig. 3, the growth of peat was able to keep pace with the aggradation of estuarine deposits and a considerable thickness of peat accumulated up to present ground surface level. The estuarine deposits have since been significantly dissected only in the Carron valley and even here a considerable part of the deposits still survives as a terrace.

ARNPRIOR

The western part of the Carse of Stirling is almost cut off from the much larger eastern part by a major moraine, which exceeds 100 feet in height in places and has been traced for 12 miles.[5] The moraine is breached by the Forth and is also deeply cut into by several valleys, which form another group of pre-carse features. One of these valleys, along with the adjacent morainic ridges and surrounding carse, is shown in Fig. 4.

In its north-western part the valley is relatively small but where it crosses the highest part of the moraine its present depth is 50 feet and its width about 200 yards. For most of this distance the floor is flat and marshy. The valley then turns through a right-angle and begins to open out, one side being formed by the moraine and the other by the basal slopes of the hills to the south. At the bend in the valley a small stream, the Arngibbon Burn, comes in from the south and has deposited an alluvial fan over the full width of the valley floor (Fig. 4). North-eastwards the fan blends into the flat carse-lands.

Three boreholes were made in the valley floor. Nos. 2 and 3 (Fig. 4), located on the alluvial fan, showed that the latter rests on carse clay beneath which is peat up to two feet thick. Borehole 1 was made in the flat, marshy valley floor to the north-west of the fan. Here, beneath nine feet of peat and lacustrine clay, carse clay five feet thick was passed through, and this was followed by two and a half feet of peat and clay. These boreholes thus show that, after the valley had been excavated, peat accumulated on its floor and was covered by the carse clay during a marine transgression. Subsequently the Arngibbon Burn built a fan on top of the carse and caused a lake, now drained, to be formed in the western part of the valley.

AGE OF THE VALLEYS

The carse deposits that occupy the lower reaches of the valleys described in the Forth and Tay areas are related to a major marine transgression. The shoreline (so-called 25-foot shoreline) associated with this transgression, which has been accurately levelled by the writers at several hundred points,

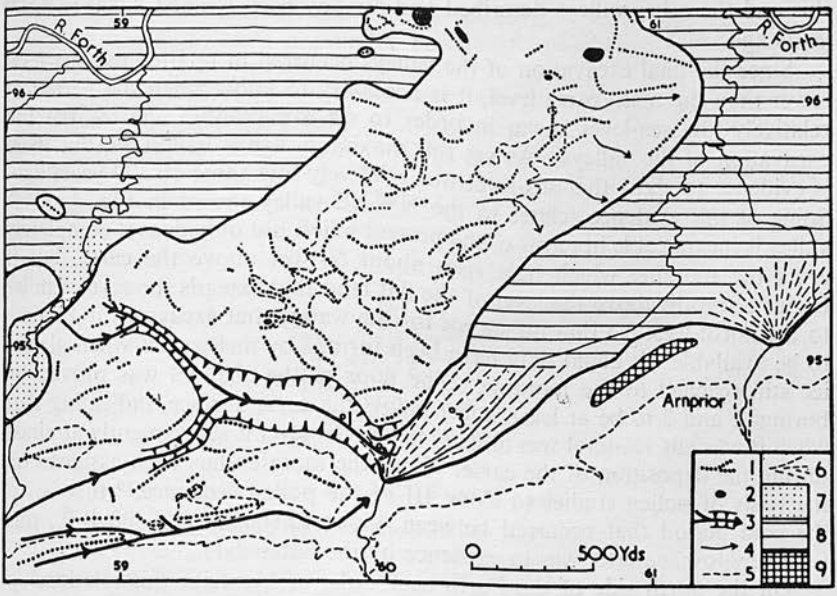


Fig. 4. Part of the major moraine that occurs near Arnprior. 1. Individual ridges comprising the moraine; 2. kettles; 3. major meltwater channels; 4. minor meltwater channels; 5. 100-ft. contour; 6. alluvial fans; 7. fluvial sand not disposed as fans; 8. flat carselands; 9. outwash terrace.

slopes in a general eastward direction down the Forth and Tay valleys.

For example, it is at 48-49 feet near Gartmore (at the extreme western end of the Forth carse), 42 feet at Stirling, about 30 feet in Edinburgh, 25 feet at Aberlady and 20 feet at the mouth of the Tyne (East Lothian); 33-34 feet at Bridge of Earn and 27-28 feet at Leuchars. The shoreline resulted from the major post-glacial eustatic rise of sea-level produced especially by the final melting of the North American and European ice-sheets, as a consequence of which sea-level rise in central Scotland was more rapid than the isostatic uplift of the land. It has been concluded from evidence elsewhere[6] that this major rise of sea-level culminated about 5,000 to 5,500 years ago. In the Forth valley an age of this order is indicated by radiocarbon assay of a sample of peat from very near the base of a moss that overlies the carse, which gave a date of 5,482 ± 130 Before Present.[7] If the main level of the carse is of approximately the same age throughout the Forth and Tay areas, the final infilling by the carse of the lower parts of the valleys that dissect the late-glacial deposits occurred some 5,500 years ago.

The peat met with in the boreholes referred to above is part of a widespread layer that underlies the main carse deposits and whose existence has long been known. For example, this peat layer enabled Jamieson in 1865[8] to demonstrate the major post-glacial transgression. The peat has been radiocarbon-dated at two points in the Tay basin and one in the Forth basin, the dates obtained being 8,411 ± 157, 8344 ± 143, and 8411 ± 157.[9] The date in the Forth valley was obtained at the Airth colliery site and is situated less than a mile from one of the valleys considered above. It is thus apparent that

this and the other valleys described had already been formed 8,000 to 8,500 years ago.

Since the final excavation of the valleys occurred in relation to a sea-level lower than the main carse level, it is necessary to know when the period of relatively low sea-level began in order to set a maximum age for the full excavation of the valleys. As yet this maximum age is uncertain, but there is evidence to show that sea-level was relatively low some 10,300 years ago. Some of this evidence relates to the NW-SE valley shown in Fig. 4. This valley is inexplicable in terms of the present relief: not only does it completely cross the moraine, which here rises about 60 feet above the carse, but it begins abruptly above the level of the flat land that extends for several miles to the north-west, so that no source for the waters that excavated it appears to be available. It could only have been formed by meltwaters when glacier ice still reached to the moraine. The floor of the channel was proved by borings 2 and 3 to be at least 9 feet below the carse surface, indicating that when it was cut sea-level was already below the level it subsequently attained during the deposition of the carse. Since the moraine has been assigned on the basis of pollen studies to Zone III of the pollen sequence,[10] that is, to the cold period that occurred between about 10,800 and 10,300 B.P., this relatively low sea-level was in existence by the latter date.

On the north side of the Forth near Lake of Menteith this relationship is demonstrated by fluvioglacial deposits. This kettle lake lies immediately to the west of the moraine and is drained by the Goodie Water, a fairly small stream that utilises a gap cut through the moraine by meltwaters (Fig. 5). The gap contains coarse outwash gravel whose surface slopes down from 77 feet O.D. to below 50 feet O.D., whereupon it passes beneath the carse. The outwash can then be followed in the banks of the Goodie Water, which here occupies a straight artificial cut, and its surface can be seen to continue to descend to stream level. On top of the outwash in the stream-bank section is the sub-carse peat and on top of this the carse itself, which thickens east-north-eastwards as the outwash diminishes in altitude. Boreholes were put down beyond the point where the outwash disappeared below stream level and in three of these the outwash was encountered (Fig. 6). The outwash could not have been laid down once Lake of Menteith came into existence and must have been deposited when the lake site was still occupied by glacier ice. Since the lowest point at which the surface of the outwash was met with (borehole 2, Fig. 6) was 37·8 feet O.D., sea-level could not have been above this altitude, some ten feet below the adjacent carse shoreline, when glacier ice still existed immediately west of the moraine.

Thus, the valleys described in this paper as cut in late-glacial deposits must have been formed not later than about 8,000 to 8,500 B.P., could have been formed as early as 10,300 B.P., and may have been produced at an even earlier time.

CAUSE OF THE VALLEYS

With the exception of the meltwater channel shown in Fig. 4, the cause of the valleys is not known. It is possible that they were initiated in late-glacial times when sea-level was high and ground water consequently higher than now, and that they continued to be excavated in response to a falling

water-table controlled by the relative fall of sea-level. Alternatively, the valleys may have been cut under periglacial conditions when the ground

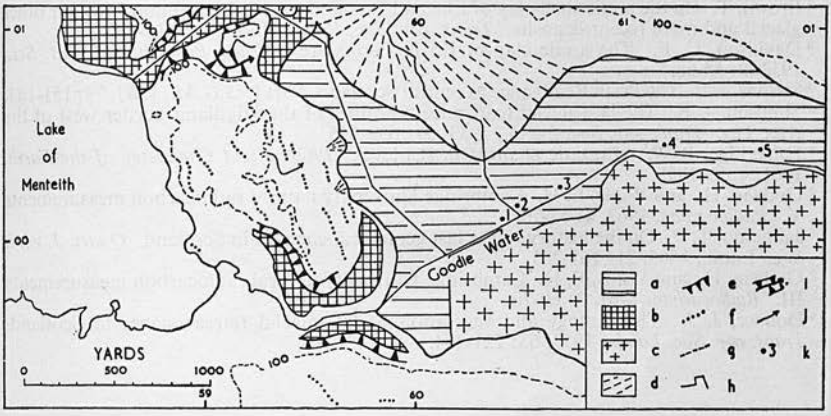


Fig. 5. Landforms east of Lake of Menteith. a. carse-land with occasional patches of surface peat; b. outwash terraces; c. peat; d. alluvial fans; e. steep slopes due to meltwater erosion; f. individual ridges of the major moraine that borders the east side of Lake of Menteith; g. 100-ft. contour; h. streams, mostly in artificial courses; i. large meltwater channel; j. small meltwater channel; k. boreholes (see text and Fig. 6).

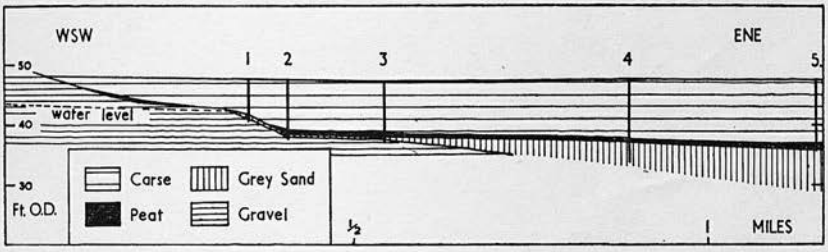


Fig. 6. Section through the deposits along the course of the Goodie Water. The left part of the diagram is based on levelled sections exposed in the sides of the stream channel at low water (marked on diagram) and the remainder on the boreholes whose location is indicated in Fig. 5.

was frozen and precipitation could not be absorbed by the now porous sands and gravels. With reference to this hypothesis it may be significant that proof of permafrost having existed since some of the late-glacial raised beaches of the Forth valley were laid down is provided by fossil frost wedges. For example, such frost wedges were recently visible in the raised delta deposits of the River Carron in a sand pit near Stenhousemuir at about 80 feet O.D. and in a gravel pit near Larbert at the same altitude. It may also be noted that if permafrost conditions operated during the deglaciation that followed the Perth Readvance then the concept of a falling water-table controlling valley excavation would not appear feasible. In this event stream erosion when frozen ground conditions existed would appear to account for the valleys.

The writers wish to express their gratitude for borehole data to the Burgh Engineer for Falkirk and to the County Surveyor for Perth and Kinross.

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VIII.—Raised Shorelines Associated with the Perth Readvance in the Forth Valley and their Relation to Glacial Isostasy.*

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Synopsis

Detailed mapping and accurate levelling of fluvioglacial and raised marine features in the Forth Valley between Stirling and Burntisland show that the most conspicuous late-glacial raised beaches of this area are associated with the Perth Readvance and possess a marked gradient down the Forth Valley. The relation of the raised beaches to fluvioglacial deposits is described and discussed and this leads to the conclusion that the Perth Readvance resulted in a renewed downwarping of the land, the response to glacial loading and unloading being immediate and rapid.

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General Geomorphology of the Area

East of the vicinity of Bo'ness on the south side and of the neighbourhood of Kincardine on the north side, the Forth is for the most part flanked by steep bluffs 50–100 ft. high. Away from the Forth, behind the bluffs, the ground is strongly ice-moulded in a direction that is never far from west-east. The depressions between the ice-moulded ridges often contain the flattish-surfaced deposits of former lakes or harbour fluvioglacial forms.

North-west of Kincardine the Forth narrows rapidly and upstream of Alloa is a strongly-meandering river, although it is tidal to a point 2 miles above Stirling. West of Bo'ness and of a point midway between Culross and Kincardine the river is bordered by a plain formed of the post-glacial carse muds. This carse plain is usually about 2 miles wide on the south side of the river and about 1 mile wide on the north side. It fingers into many valleys, especially those of the Devon and Carron. The plain is particularly extensive from Stirling eastwards to the junction of the Devon with the Forth, its total width here approaching 4 miles. The plain is usually backed by a sharp bluff and this feature is continuous with the bluffs referred to above as bordering the Forth itself further east. The late-glacial raised beaches considered in this paper are very often situated at the top of this bluff and from them one can look out across the carse-lands or down to the Forth.

The Stirling-Kincardine-Bo'ness lowland is surrounded on three sides by much higher

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ground through which are two major gaps of especial importance in relation to glacier movements. To the north the fault-line scarp of the Ochils rises abruptly 1500–2000 ft. above the carse plain. To the west the Kilsyth-Gargunnock hills rise less impressively to altitudes of about 1500 ft. Between these two uplands lies the deep gap utilized by the Forth and commanded by Stirling. South of the Kilsyth-Gargunnock hills, and between them and the moorlands that reach nearly 1000 ft. around Kirk O' Shotts, lies the Kilsyth-Larbert gap. This gap, which is joined by the River Carron near Larbert, leads directly from the Clyde lowlands to those of the Forth.

Previous Work

(a) *Glaciation*

At a time prior to the formation of the raised beaches, discussed below, the whole of the area considered in this paper was covered by glacier ice. Subsequently, there was extensive deglaciation, followed by a major readvance. During this readvance, which may be referred to as the Perth Readvance (Simpson 1933), the western part of the Central Lowlands of Scotland again became ice-covered (Sissons 1963*b*, 1964). In the Forth area the main ice stream from the Highlands passed through the Stirling gap and the ice then spread out eastwards and south-eastwards as a lobe that extended to the vicinities of the sites of Kincardine and Plean. At the same time an extensive mass of Highland ice occupied the Clyde coalfield basin and covered much of the Kilsyth-Gargunnock hills. A tongue of this ice, utilizing the Kilsyth-Larbert gap, reached almost to the site of Larbert in the Carron valley. The ice masses in both the Forth and Carron valleys terminated in part in the sea, then relatively higher than now, and the meltwater streams poured into the sea great volumes of debris that contributed significantly to the formation of beaches.

(b) *Raised Beaches*

It is necessary to consider here only the most recent literature on the raised beaches of the Forth valley since the earlier literature has been discussed at some length in a recent paper (Sissons 1962). In that paper, which was written before the fieldwork described below was commenced, it was suggested on theoretical grounds that some late-glacial shorelines in Scotland were likely to have considerably steeper gradients, due to isostatic warping, than had hitherto been recognized. Thus it was tentatively suggested that a shoreline at 140 ft. O.D. at Stirling might slope eastwards down the Forth Valley with a gradient of about 2·25 or 2·33 ft. a mile so that at Edinburgh it might be at 65–70 ft. and at Dunbar at 10–15 ft. O.D.

These suggestions were strongly criticized by Earp, Francis and Read (1962). These authors included in their paper a diagram showing a horizontal raised beach at between 120 and 135 ft. O.D. extending for about 10 miles from west of Stirling to east of Kincardine. In the text they emphatically stated that differential warping had not occurred. On the other hand, Donner (1963), in the second part of a paper dealing with raised beaches in Scotland as a whole, concluded that the so-called "100-foot raised beach" is tilted towards a direction south of east in the Forth valley. Donner drew isobases for this beach, representing it as descending from about 115 ft. O.D. near the Forth Bridges to about 75 ft. O.D. in the vicinity of Dunbar.

The evidence and interpretations given in the present paper differ from those of both Donner and of Earp *et al.* There are two main reasons for these differences.

(i) The height data used by Earp *et al.* were not obtained by instrumental measurement of raised shorelines in the field but were taken from the information published on Ordnance Survey maps. Although Donner measured features instrumentally he used an aneroid barometer and himself stated that measurements at the same point differed by up to two metres. His datum plane was the barnacle line, which varies in altitude, and which he admitted was not always present. Furthermore, Donner's isobases for the "100-foot raised beach" in the Forth area are based on only four measured points. In the present paper the writers have attempted to eliminate so far as possible the uncertainties about height data by accurately levelling raised shorelines and associated fluvioglacial deposits at several hundred points.

(ii) The heights used by both Donner and Earp *et al.* as applying to raised beaches relate in some instances to features that the present writers have identified as glacial outwash plains or kame terraces. This difference of interpretation has arisen because in their surface expression the raised beaches and the fluvioglacial spreads often appear to be identical, for both usually form apparently-flat features backed by rising ground. Accurate closely-spaced height measurements show, however, that whereas the raised beaches described below have average gradients of between about 2 and 3 ft. per mile, the fluvioglacial spreads have typical gradients of between 15 and 40 ft. per mile (Sissons 1963*a*) Furthermore, the fluvioglacial spreads are composed predominantly of gravel, while the raised beaches consist almost entirely of sand, silt and clay.

Few faunal remains have been discovered in the raised beaches, this being usually attributed to the cold climate that prevailed when they were being formed. Reddoch (1825) gave notice of the discovery of "the Bones of a Quadruped, found in a bed of clay, and of razor-shells found in a bed of sand under the clay, near Camelon, 90 ft. above the present level of the Forth". Knox (1825) identified the bones as those of a seal. The writers were recently informed by a workman that he had found a bed of "sea-shells" in stratified sand at a depth of 8–10 ft. below ground level near the centre of Falkirk. (Nat. Grid Ref. NS/888801). The altitude of this bed was between 80 and 90 ft. O.D. Remains of a seal have also been found about 3 miles north-east of Burntisland in the Tyrie clay-pit at about 30 ft. above high-water mark (Allman 1858).

J. Geikie (1894, pp. 271–273), referring to some of the raised beaches considered in this paper, stated that they consist chiefly of tough, finely-laminated plastic brick-clays, these deposits being overlain by alternations of fine sand and sandy clay. He remarked that the brick-clays contained scattered striated stones, presumably dropped from floating ice. Geikie also stated that in places the beds are "much crumpled and confused" and attributed this to the action of sea-ice. Such contortions were observed by the writers in a sand pit near Camelon. Normally, however, sections show more or less horizontal layers of sand, silt and clay.

Fieldwork

The fieldwork was carried out independently by the authors, one (D.E.S.) working on the north side of the Forth, and the other on the south side. In both areas raised beaches and glacial forms were mapped on a scale of 1 : 10,560. The shorelines of the raised beaches and certain glacial features were then accurately levelled from Ordnance Survey bench-marks, using the latest available heights of the latter (based on the second geodetic levelling) as

indicated on bench-mark lists or on the most recently-published 1 : 2,500 Ordnance Survey sheets.

The raised beaches are usually well-marked features, typically between 50 and 200 yds. wide but, in some areas, considerably wider. They usually slope gently towards the Forth and where they abut against higher ground a break of slope occurs. This break of slope appears to be entirely a depositional feature produced by the accumulation of the marine deposits against a pre-existing slope of rock or glacial drift. No evidence of marine erosion was observed. Neither were raised spits or bars seen in association with the beaches. It thus seems likely that the area was not being subjected to strong wave action when the beaches were being deposited: in part this reflects the distance of the area from the open sea, while it is probable that wave action in winter would be precluded by the presence of sea ice.

All measurements on raised beaches were made as near as possible to the break of slope that marks the former shoreline. The shoreline itself can rarely be measured since it is normally obscured owing to mass-movement from the rising ground behind it. Consequently the altitudes obtained for the raised shorelines are very slightly (probably less than a foot) below their true altitudes. Measurements were made at intervals of about 60–80 yds. where practicable.

Since their formation the raised beaches have been modified in various ways, especially in recent times by human influences. Numerous minor precautions were therefore taken to minimize errors that might be caused by such modifications. Some of these precautions were as follows:

(i) Gullies, alluvial fans and areas of considerable mass-movement were avoided wherever they could be detected.

(ii) Where field boundaries follow the shoreline the artificial break of slope caused by material accumulating against the wall or hedgerow was avoided.

(iii) It was found that where field boundaries run at an angle to the shoreline the ground in their immediate vicinity was sometimes up to a foot or so above the surrounding ground height even on an essentially-flat raised beach. It is presumed that this has been mainly caused by the blowing of soil from the fields. Shoreline measurements were therefore not made within about 10 yds. of present or detectable former field boundaries.

(iv) Where the ground was characterized by minor irregularities (*e.g.* potato fields) the base of the graduated staff was placed at points representative of the average height of the ground in the immediate vicinity.

(v) Part of the area considered is or has been a coal-mining area and it is possible that a few measured points have been slightly affected by mining subsidence. As a precaution against the serious error of subsidence of the bench-mark on which a traverse was based (which would have produced high readings for all the measured points on that traverse), either fundamental bench-marks were used, or traverses were commenced at one bench-mark and terminated at another on the assumption that, unless subsidence was exactly the same at the two bench-marks, it would be revealed by the closing error.

The most serious human modification of the raised beaches is associated with the construction of buildings, roads, railways, etc. As a result it was not possible to measure the features in certain areas, most notably Rosyth-Inverkeithing, Bo'ness, Falkirk-Camelon and St Ninians-Stirling. It may be added that some of the points whose altitude was measured have since been destroyed by construction works.

In addition to levelling the shorelines, related fluvioglacial features were also levelled. On outwash plains and kame terraces measurements were usually made near the break of slope where the features adjoin higher ground in the same way as on raised beaches. The highest points of kames were levelled as also were the lowest points on the rims of kettle holes.

Raised Beaches and Associated Fluvioglacial Forms

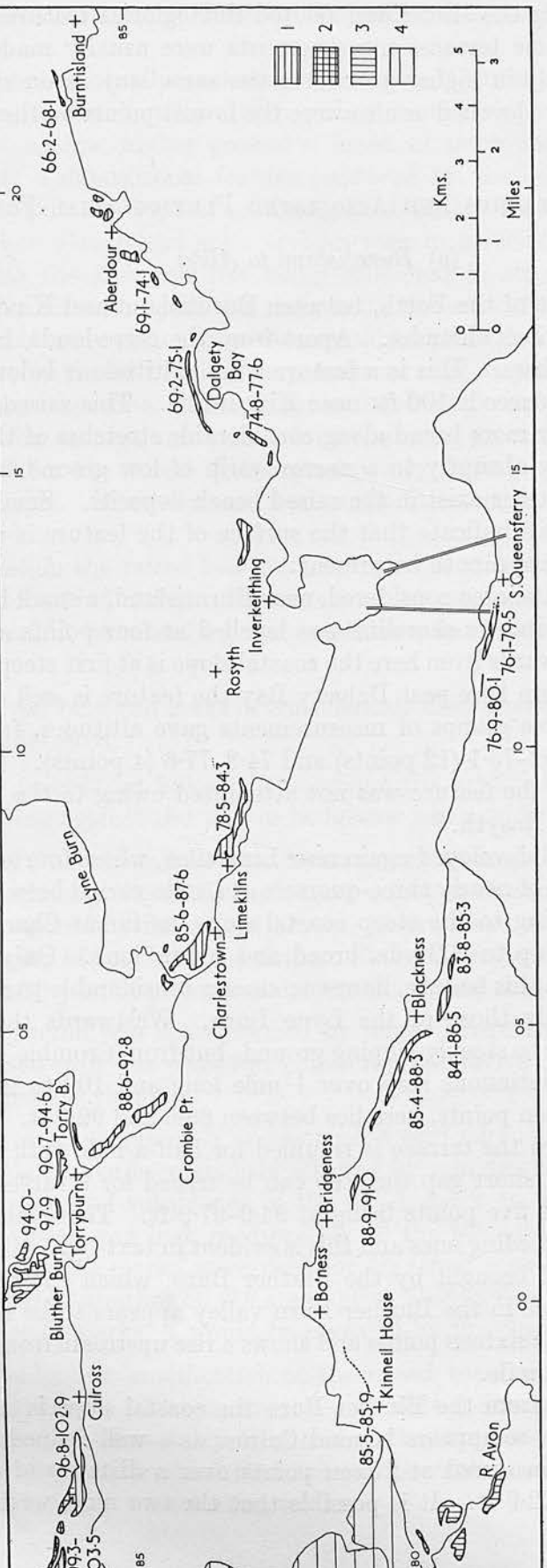
(a) *Burntisland to Alloa*

Along the northern side of the Forth, between Burntisland and Kincardine, remnants of raised beaches occur at various altitudes. Apart from the carse-lands, however, one feature stands out above all the others. This is a feature whose altitude is below 70 ft. near Burntisland but approaches and exceeds 100 ft. near Kincardine. This raised beach forms a conspicuous step 50–200 yds. or more broad along considerable stretches of this coastline. Below it the land often falls away abruptly to a narrow strip of low ground bordering the Forth. Unfortunately, no large sections exist in the raised beach deposits. Small sections, however, and the character of the soils indicate that the surface of the feature is usually composed of sand although locally clay appears to be present.

In the extreme east of the area considered, near Burntisland, a small but marked remnant of the raised beach occurs and its shoreline was levelled at four points as between 66·2 and 68·1 ft. (text-fig. 1). Westwards from here the coastal slope is at first steep, but near Aberdour the slope moderates and from here past Dalgety Bay the feature is well developed in several localities. In this area three groups of measurements gave altitudes, from east to west, of 69·1–74·1 ft. (10 points), 69·2–75·1 (12 points) and 74·8–77·6 (4 points). For the next 4 miles westwards measurement of the feature was not attempted owing to the buildings and dockyards of Inverkeithing and Rosyth.

The raised beach is well developed again near Limekilns, where fourteen measurements of its shoreline over a distance of nearly three-quarters of a mile varied between 78·5 and 84·3 ft. Thereafter it is absent, owing to the steep coastal slope, as far as Charlestown, where it is resumed as a marked step up to 300 yds. broad and 1 mile long. Only five measurements (85·6–87·6 ft.) were made on this feature, however, since a considerable part of it is modified by old river deposits, especially those of the Lyne Burn. Westwards the continuity of the feature is interrupted owing to steeply-sloping ground, but from Crombie Point to near Torryburn it forms an almost continuous step over 1 mile long and 100 to 200 yds. broad. Its shoreline, as levelled at fifteen points, here lies between 88·6 and 92·8 ft. West of the breach produced by the Torry Burn the terrace is resumed for half-a-mile with a shoreline at 90·7–94·6 ft. (4 points). After a short gap the step can be traced for a further quarter mile, the break of slope as levelled at five points being at 94·0–97·9 ft. These last measurements are rather high in relation to preceding ones and this is evident in text-fig. 5 (at D). This anomaly is probably due to material brought by the Bluther Burn, which enters the Forth nearby. The most conspicuous terrace in the Bluther Burn valley appears to be related to the raised beach; it has been levelled at sixteen points and shows a rise upstream from 93·5 ft. to 104·0 ft. in about three-quarters of a mile.

For 2 miles westwards from the Bluther Burn the coastal slope is steep and the raised beach absent, but the latter re-appears beyond Culross as a well-defined feature. Here the height of the shoreline as measured at fifteen points over a distance of three-quarters of a mile is between 96·8 and 102·9 ft. It is possible that the two most westerly measurements



TEXT-FIG. 1.—Raised beaches associated with the Perth Readvance in the Burntisland-Bo'ness area. (1) Raised beaches: the ranges of the levelled heights, in ft., obtained on the various parts of the shorelines are indicated. (2) Fluvio-glacial terraces. (3) River terraces. (4) Carse and reclaimed land.

(C, text-fig. 5) are on glacial outwash, in which case the height range of the shoreline would be 96·8–100·7 ft. After a short break the feature resumes as a broad terrace 1 mile long and up to 300 yds. wide, the shoreline altitude varying between 99·3 and 103·5 ft. (13 points). Following another short break the beach reappears near Kincardine for a further 700 yds., five measurements on it ranging between 103·7 and 106·2 ft. Here the margin of the high ground changes direction to trend south-north and about 1½ miles to the north a probable remnant of the feature forms a terrace with a break of slope measured at six points as lying between 105·9 and 108·5 ft.

The above evidence shows a consistent pattern. Over an east-west distance of 18 miles a total of 112 measurements has been made on thirteen separate features and the averages of the heights for the successive features increase continuously from east to west, as follows: 67·2, 71·2, 72·7, 76·2, 80·7, 86·5, 90·5, 93·4, 95·7, 99·0, 101·0, 105·3 and 107·0 ft. Furthermore, although the indented coastline between Rosyth and Aberdour is more favourable for the formation of the beach and its preservation as a raised beach than the straighter, often steeply sloping coastline to the west of Rosyth, the beach is better developed in the western area, attaining its optimum development towards Kincardine.

In total contrast to the above picture is the complete absence of the beach west and north-west of Kincardine. This is especially significant since here considerable areas at the appropriate altitudes are ideal for the formation and preservation of the feature. Thus, around Clackmannan and Alloa, over a distance of 4 miles, there occurs between the ice-moulded ridges a series of broad depressions that descend well below the 100-ft. contour. Although these depressions are extensively floored with marine depoists, the latter rarely exceed 75 ft. (text-fig. 2). Under these circumstances the absence of the beach from this area can be adequately explained only if the area were occupied by glacier ice, and it has been suggested elsewhere (Sissons 1963*b*) that this ice belonged to the Perth Readvance.

The limit of the readvance appears to have been attained along the slopes of the rising ground north of Kincardine where kames and irregular mounds of till occur. In this area a small gravel outwash terrace slopes down from 123 ft. to the level of the westernmost raised beach remnant (A, text-fig. 5). A much longer outwash terrace, fed by a meltwater channel, begins near Kincardine and over a distance of nearly 1¼ miles descends from 130 ft. to the level of the raised beach, here at about 100 ft. (B, text-fig. 5). It thus appears that outwash from the ice was being poured into the sea at the time the beach was being formed, thus helping to account for the considerable extent of the beach in the Kincardine-Culross area. It is also evident that the beach corresponds with the maximum of the readvance or very slightly post-dates it. Furthermore, the absence of the beach in the area west and north-west of Kincardine suggests that while glacier ice still remained in this area there was a relative uplift of the land.

(b) *South Queensferry to Bo'ness*

On the south side of the Forth between South Queensferry and Bo'ness by far the most conspicuous raised beach is a feature whose shoreline rises westwards from about 77–78 ft. near South Queensferry to about 90 ft. at Blackness. These heights are very similar to those obtained immediately opposite on the north side of the Forth.

Between the Edinburgh City boundary (Cramond) and South Queensferry the feature appears to be absent, probably mainly because along much of this coastline the ground rises



TEXT-FIG. 2.—Raised beaches and fluvioglacial features associated with the Perth Readvance in the Grangemouth-Bannockburn area. (1) Raised beaches related to the Perth Readvance: the ranges of the levelled heights, in ft., obtained on the various parts of the shorelines are indicated. (2) Outwash plains, kame terraces and flattish-topped kames. (3) Raised beach deposits below 75 ft. O.D. in the Alloa-Clackmannan area. (4) Carse and related post-glacial deposits, including reclaimed land. (5) Kames. (6) Kettles and other dead-ice hollows. (7) Meltwater channels. (8) Provisional line suggesting limit of Perth Readvance.

very steeply from the present shore. In the 7½-mile stretch from South Queensferry to Bridgeness, however, the raised beach has been mapped over a distance of 3½ miles. Occasionally it is a narrow step only 20 yds. or so wide but for the most part it forms a prominent terrace 100 yds. or so wide with a maximum width of about 200 yds. The back-slope is usually steep and at first sight suggests an abandoned marine cliff. That it is not of this origin is indicated by the fact that in places (*e.g.* south of Blackness and between Blackness and Bridgeness) the raised shoreline gradually trends away from the steep backslope yet the latter retains its steepness and clarity. The backslope represents the flanks of ice-moulded ridges against which the sea laid down its deposits. No good sections occur in the raised beach but its surface is often sandy. A brick and tile works used to exist near the northern margin of the beach immediately south of Blackness (at NT/051797) and Milne-Home (1871) stated that 15 or 20 ft. of finely-laminated clay was worked.

The altitude of the raised shoreline has been determined at sixty points between South Queensferry and Bridgeness. These measurements were made on six separate features and the averages of the heights for the features show an increase from east to west, as follows: 77·5, 78·0, 84·7, 85·3, 87·0 and 90·1 ft.

(c) *Bo'ness to Falkirk*

Between Bo'ness and Falkirk the pattern is less simple, partly because a raised beach at a slightly lower level than the main one is present in places, and partly because the built-up areas of Bo'ness, Laurieston and Falkirk obscure considerable areas of the raised beaches.

Immediately west of Bo'ness in the grounds of Kinneil House occurs a small raised beach remnant about 100 yds. long. The feature may continue eastwards into Bo'ness for about half-a-mile but is here so modified by roads, buildings, and rubbish dumps that it can be mapped only tentatively and cannot be measured. The shoreline of the clearly-preserved part near Kinneil House was measured at two points, however, as 85·5 and 85·9 ft. These heights are clearly below those of the western part of the shoreline described above as occurring between Bridgeness and South Queensferry. That shoreline, rising westwards, has already attained about 90 ft. about 2½ miles to the east of Kinneil House, and if projected to the latter locality at the same slope would be at 94–95 ft. It thus appears that the feature near Kinneil House represents a raised shoreline that is slightly lower than the one described in the previous section.

Westwards from Kinneil House to the vicinity of the river Avon the carse plain is backed by a very steep bluff that rises to about 100–150 ft. O.D., and neither of the raised shorelines occurs here. Between the Avon and Falkirk, however, both shorelines are present, but at no point was the one definitely identified above the other. Immediately west of the river Avon a clear but narrow raised beach extends for 700 yds. and measurements at eight points gave heights between 98·8 and 104·3 ft. (text-fig. 2). It is probable that the higher measurements obtained on this feature are too high, for they were made at its narrowest part where the surface material is gravel, probably due to downwash from the steep back-slope. The measurements of 98·8, 101·2 and 101·3 are probably the most representative of the true altitude of this shoreline remnant.

After an interval of a quarter of a mile the feature is resumed north of Polmont for a distance of 700 yds. Here it is much broader, with a maximum width of over 200 yds. The surface material is mostly sand which was visible to a depth of several feet in places, and this appears to be underlain by marine clay. Gullies interrupt the feature, but five measurements of the shoreline made away from the gullies gave heights between 100·8 and 102·9 ft.

Farther west, at Laurieston, is a raised beach about two-thirds of a mile long and 50–100 yds. broad. The feature is partly modified by roads and buildings and only three measurements on unmodified ground were possible. These gave heights of 95·5, 98·6 and 99·9 ft. After a gap of a few hundred yds. the feature is resumed and seven levelled points lay between 97·7 and 102·4 ft. Thereafter the feature is lost in the built-up area of Falkirk.

When mapping the four shoreline remnants between the Avon and Falkirk the impression was gained that they were all part of one shoreline. It may seem that this interpretation is confirmed by the levelling for the twenty-three measured points all lie between 95·5 and 104·3 ft., and if the five less satisfactory measurements on the feature near the Avon are omitted, the range is 95·5–102·9 ft. Closer examination of the data, however, indicates that two shorelines are represented. The average heights of the measurements on each of the features, from east to west, is 102·3 ft. (or 100·4 ft. for the three more satisfactory measurements), 102·0 ft., 98·0 ft. and 99·9 ft.: if interpreted as one shoreline these figures indicate a gradient down towards the west. Yet the sixty points between South Queensferry and Bridgeness show a gradient down towards the east, and to the north-west of Falkirk other measurements (discussed below) also show a definite slope down the Forth valley. On the other hand, if the slope of the South Queensferry to Bridgeness shoreline (*i.e.* the upper shoreline) as calculated by linear regression analysis is projected into the areas occupied by the four features, the following heights are obtained from east to west: 99–100, 100–101, 103 and 104 ft. Thus the two more easterly features are close to this line, and the two more westerly are several feet below it, suggesting that the two former are remnants of the upper shoreline and the two latter of the lower.

Near the river Avon, which occupies a rock gorge, well-marked terrace features exist between altitudes of 104 and 120 ft. The features tend to decline in altitude downstream and they may have been laid down when the higher raised beach was being formed.

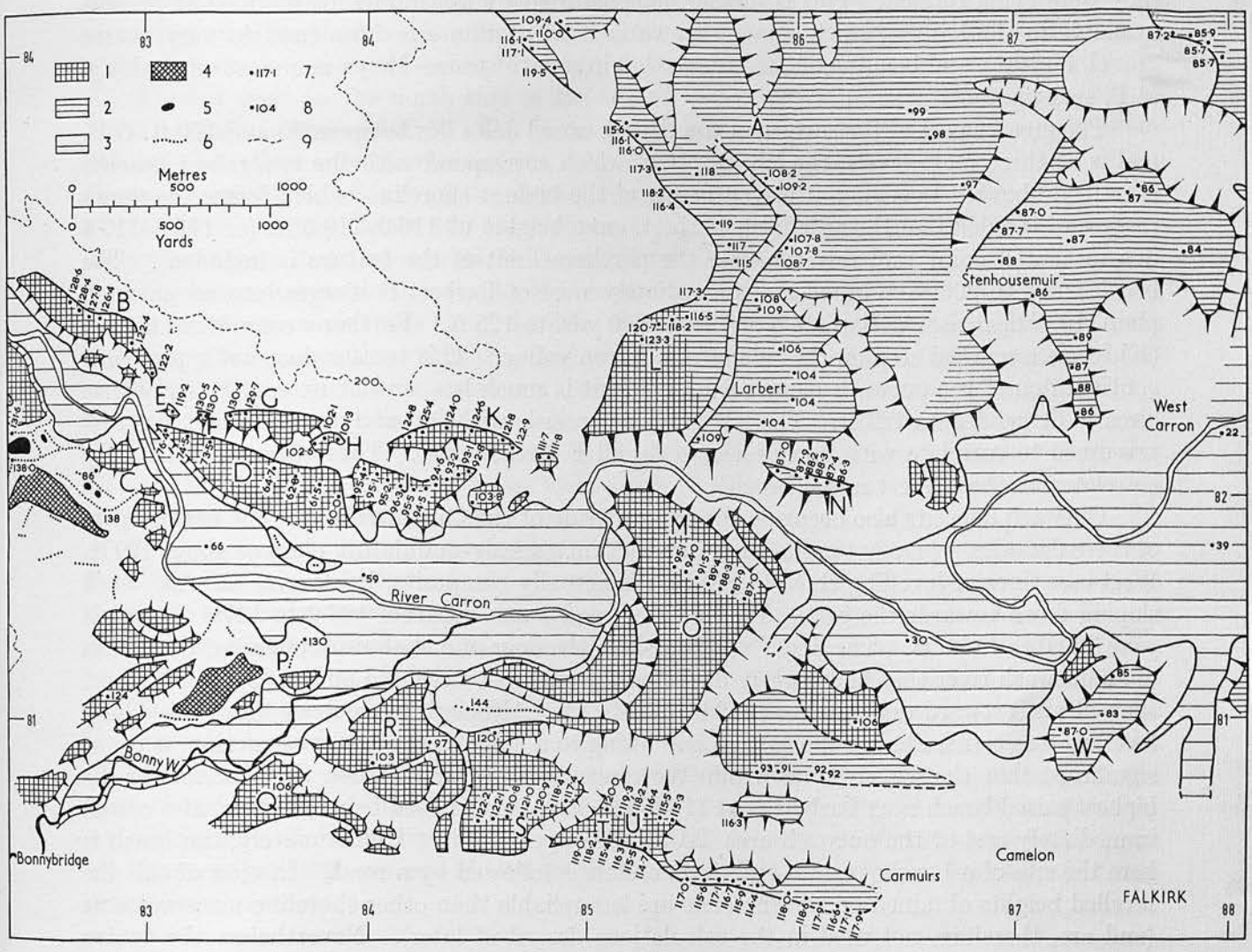
(d) *Falkirk to Larbert and Bonnybridge*

In this area the principal late-glacial marine feature is the extensive raised delta * of the river Carron. If the parts of the delta subsequently destroyed by the Carron and other streams were reconstructed its total area would approach six square miles. Westwards from Larbert and Camelon the delta merges into outwash deposits and these in turn are soon replaced farther westwards, in the Carron valley, by ice-contact forms comprising kames, kame terraces and kettles (text-fig. 3).

The surface of the raised delta is almost everywhere composed of sand, sometimes mixed with a little gravel. Near the Carron at least 40 ft. of sand is exposed in pits, with occasional thin layers of laminated clay. Away from the river the sand becomes thinner and the laminated clay more important. The main mass of the clay underlies the sand and boreholes show a gradual transition from one to the other. For example, north of Larbert (at A, text-fig. 3) a group of boreholes 35–45 ft. deep showed stiff clay, overlain by soft clay with layers of sand, followed by clayey sand, with pure sand a few feet thick at the surface. Westwards from Camelon and Larbert the sands are capped in many places by a sheet of coarse gravel comprising steeply-dipping fore-set with overlying gently-dipping top-set beds. The surface of

* The term "delta" is used here for convenience. It is not implied that the feature was built above contemporary sea-level. Rather was it comparable with the "delta" now being formed at the mouth of the river Esk, east of Edinburgh, which feature is largely exposed at low tide but covered at high tide.

these deposits normally slopes eastwards and this is also the prevailing direction of their dip. The whole sequence of deposits is now exposed in a large recently-opened pit immediately south of the Carron (M, text-fig. 3).



TEXT-FIG. 3.—Raised beaches and fluvioglacial features associated with the Perth Readvance in the Carron valley. (1) Outwash plains, kame terraces and flattish-topped kames. (2) Raised beaches related to the Perth Readvance. (3) Flood-plain and post-glacial terraces of the River Carron and Bonny Water. (4) Dead-ice hollows other than kettles. (5) Kettles. (6) Kames. (7) Points levelled by authors, to nearest tenth of a ft. (8) Points levelled by Ordnance Survey, to nearest ft. (9) Contour at 200 ft. Letters explained in text.

The section here is as follows:

Top-set gravels and coarse sand	6– 8 ft. thick
Fore-set gravels and coarse sand	8–10 ft. ,,
Sand, with layers of clay becoming more numerous towards the base	10–12 ft. ,,
Stiff grey laminated clay	7– 8 ft. ,,
Till	7– 8 ft. ,,
Bedrock	—

A significant point about this pit is that there is no evidence of disturbance by glacier ice of any of the deposits overlying the till, and this is equally true of all the extensive sections seen in these deposits farther east. Yet a few hundred yards to the west a kame terrace

flanks the Carron valley, three-quarters of a mile to the south-west a kame occurs, while 1 mile to the west begins a kame and kettle complex which occupies much of the floor of the Carron valley. These features testify to the presence of glacier ice in their immediate vicinity when they were being formed. This is further demonstrated by the numerous sections at present available in the kames, which show great variations in grain size from sand to very coarse gravel, faulting, and bending of the beds so that in some instances they are almost vertical (*e.g.* at P, text-fig. 3).

The greater part of the surface of the Carron raised delta lies between 80 and 120 ft. O.D. and is at three main levels, the two upper of which correspond with the two raised beaches described above. Levelling at eleven points of the highest shoreline, which forms the upper margin of the delta northwards from Larbert, gave heights of 116·0–119·5 ft. (or 115·6–119·5 if a measured point probably too near the northern limit of the feature is included). The beach is up to 300 yds. broad and immediately west of Larbert it merges into an outwash plain (L), which rises westwards 6 or 7 ft. in 250 yds. to 123 ft. Farther west a kame terrace (K) forms a marked step on the side of the Carron valley. This terrace does not represent a continuation of the outwash plain, for its gradient is much less, and at its eastern end it has already descended to 122 ft. The low gradient accords with its sandy composition and it is presumed to correlate with kame terraces G and F farther west. These three terraces may be related to the highest raised beach.

Outwash deposits also occur on the opposite side of the Carron valley. The eastern part of these deposits, where not gullied, appears to form a gently-undulating plain at about 120 ft. West-east lines of levelling show the feature is actually composite, its several elements each sloping down towards the east. The line S showed a descent from 122·2 to 117·4 ft. and U gave 120·0 to 115·3 ft. The line T was run along the floor of a shallow depression, the course of an outwash river that had slightly dissected the outwash measured on line U: this channel showed an eastward descent from 119·0 to 114·7 ft. Although on each of these three lines levelling was terminated at the eastern end owing to a gully or artificial obstruction, it seems significant that the features end within the range 114·7–117·4 ft. when the shoreline of the highest raised beach near Larbert is at 116·0–119·5 ft. This raised beach, in fact, also occurs immediately east of the outwash area T-U, and merges into it. Unfortunately, the beach is here the site of a housing estate and the shoreline is followed by a road. In view of this the levelled heights obtained on the shoreline are less reliable than other shoreline measurements (and are, therefore, not used in the calculations described later). Nevertheless the twelve measurements all lie between 114·4 and 118·6 ft. over a distance of half-a-mile (115·6–118·6 ft. if a measurement near a gully is excluded). It thus appears that on both sides of the Carron valley outwash merges into the highest raised beach.

The second step on the Carron delta has a maximum width of over half-a-mile and a back-slope 6–10 ft. high. Part of the shoreline is built over, but five measurements on open ground varied between 107·8 and 109·2 ft. A correlation between this feature and the fluvioglacial deposits has not been established, although kame terraces B and E in text-fig. 3 may be related to it.

The third level on the Carron delta, best developed around Stenhousemuir, is up to three-quarters of a mile wide. Numerous Ordnance Survey spot heights on this feature lie between 84 and 88 ft. The shoreline, as levelled at thirteen points, lies between 84·3 and 87·7 ft. Near Larbert a small remnant of this feature occurs (N, text-fig. 3) and here the break of slope as levelled at six points lies between 86·3 and 88·2 ft. South of the River Carron, at W, another remnant of the feature is present, its flattish surface lying between 83 and 87 ft.

Farther west is another remnant of the same level (V). Much of this is obscured by buildings but Ordnance Survey spot heights give its altitude as about 90 ft. The western part of this feature merges into outwash and, after a break caused by a gully, the outwash covers a large area (O), the greater part of which slopes down eastwards. A line of levels showed an eastward decline of the surface from 95 to 87 ft. Contemporaneity of formation of this outwash plain with the lowest raised beach thus appears to be established.

Farther west on the north side of the Carron valley terrace H and the higher parts of terrace J slightly exceed 100 ft. The line of levels shown on terrace J was run along the floor of a broad shallow outwash channel cut into the terrace, the height of the eastern end of the channel being 93 ft. South of the Bonny Water terraces Q and R slope down-valley from about 110 ft. to just below 100 ft. (Ordnance Survey heights). The altitude of all these features suggests that they are related to the lowest raised beach.

The kame and kettle topography that begins in the angle between the Bonny Water and the river Carron and extends along the floor of the Carron valley shows no indication in form or in the numerous sections of marine action. Yet much of this area is below the two higher shorelines. The lowest points on the rims of two of the kettles are 103 and 93 ft. respectively. Most of the kames are sharp-crested although one (C) has a broad flat top 250 yds. across. The highest parts of most of the kames lie between 127 and 138 ft. and are thus of similar altitude to the kame terraces on the valley side to the north, implying approximate contemporaneity of formation.

The sequence of events revealed by the Carron evidence can now be summarized. Ice from the west advanced towards the lower Carron valley, its approach being recorded by the change in marine sedimentation from clays, to clays with layers of sand, to sand. The ice attained its maximum extent in the vicinity of the line shown in text-fig. 2. Kame terraces and outwash formed at this time relate to the highest raised shoreline of this area at about 115–118 ft. As the ice decayed and became stagnant following the maximum, lower kame terraces were formed and a kame and kettle topography developed in the central part of the valley. The form, composition, location and altitude of this topography, along with the location and altitude of the lower kame terraces show that glacier ice still existed to, or within a short distance of, its maximum extent when the two lower shorelines were formed. There was thus a negative movement of base-level of at least 30 ft. while the ice was still at or near its maximum extent.

(e) *Larbert-Stirling*

The highest raised shoreline forms a well-marked feature for much of the distance between Larbert and Plean. The main gap in it occurs about 1 mile south of Plean where rock crops out at numerous points about the altitude of the shoreline. Levelling of the shoreline shows an increase in height towards Plean, the averages of the measurements on the several sections of the shoreline, commencing with those at Larbert and terminating with those at Plean, being 116·8, 117·6, 119·0, 120·8 and 121·4 ft. No good sections exist, but small exposures and augering show that the surface of the raised beach is composed of sand, while underlying laminated clay forms the frontal slope that passes down beneath the carse-lands. Towards Plean the sand becomes very abundant and, if its surface were restored across the numerous gullies and valleys that dissect it, would form a plain up to 1 mile wide. A short distance north-west of Plean the raised beach merges into a gravel outwash plain up to half-a-mile broad and over 1 mile long, and this in turn merges into a kame terrace that continues for a further half mile (text-fig. 2). Eighteen measurements on this combined fluvioglacial feature

show a continuous increase in height west-north-westwards from 127 ft. to 163 ft., the slope gradually increasing in this direction also (text-fig. 6). This is consistent with the composition of the feature, the gravel becoming increasingly coarser towards the west-north-west, to become very coarse with boulders near its upper limit. Here a dead ice hollow occurs and beyond this a group of meltwater channels. Two boreholes about half-way along the outwash plain have drift records as follows:

	Ft.	Ft.
Gravel	20	20
Sand	36	56
Clay	29	85
Bedrock	—	—
Gravel	13	13
Running sand	33	46
Firm brown clay	15	61
Brown till and stones	5	66
Bedrock	—	—

The drift sequence here thus closely parallels that described in the Carron area, the clay suggesting quiet marine sedimentation, the very thick deposit of sand, the approach of glacier ice, and the gravel the very close proximity of glacier ice in front of which the outwash plain accumulated. The ice limit when the plain ceased to be formed is approximately indicated by the dead ice hollow. The merging of the outwash plain with the highest raised beach indicates approximate contemporaneity of this beach with the readvance maximum, while the continuation of this beach to the Carron area, where it again merges with outwash, establishes the contemporaneity of the readvance to near Plean with that already described in the Carron valley.

The second raised beach of the Carron delta is poorly developed between the delta and Plean. The best remnant of it occurs a short distance north of the delta, where it forms a distinct step below the highest raised beach and where its shoreline was levelled at two points as 109·4 and 110·0 ft. However, it becomes a well-marked feature east and north-east of Plean, where it occurs intermittently around the tongue of high ground that protrudes south-eastwards from Cowie. Here the shoreline as measured at seventeen points lies between 113·4 and 116·6 ft. The feature reaches its optimum development in the Bannockburn-St Ninians area. Part of Bannockburn is built on it and Ordnance Survey heights here show it to be at about 120–123 ft. The shoreline was levelled at seven points in open ground on either side of Bannockburn, the heights ranging between 120·5 and 122·9 ft.

Near Cowie there is a marked change in the landforms around the level of the second shoreline. Instead of simple smooth sandy terrace features there occurs over an east-west distance of 1 mile a belt of mounds, depressions and discontinuous terraces that are often composed of coarse gravel. Some of the mounds are kames and the depressions between are kettle holes or other dead-ice hollows. The rims of two kettles were measured as 103·2 and 107·6 ft., respectively, and one of the other dead-ice hollows lies below the 100-ft. contour. These features are thus well below the altitude of the second shoreline and imply the presence of dead ice in this area when that shoreline was being formed.

In the Bannockburn-St Ninians area the frontal slope of the second raised beach descends steeply to the carse plain and is composed of till in its lower part and of raised beach sands in its upper part, some 30 ft. of the latter having been seen at one point. The beach is interrupted by the valley of the Bannock Burn and by several smaller valleys, while the rather large depression in it at A, text-fig. 2, may possibly be a dead-ice hollow on the sides of which gullies have developed. Between these interruptions the raised beach forms a very conspicuous

step up to a quarter-of-a-mile broad for a distance of nearly 2 miles. In St Ninians the character of the feature alters. Numerous shallow sections exposed during recent building operations show that the surface deposits change from sand to gravel which tends to become coarser north-westwards. The ground also begins to rise regularly from about 125 ft. in the centre of St Ninians to about 140 ft. near Torbrex (Ordnance Survey heights), forming a gently-sloping plain over half-a-mile in length. It is thus clear, despite the complications introduced by roads and buildings, that the second raised beach merges into an outwash plain in St Ninians. The north-western margin of this plain is very irregular in plan and overlooks a well-developed kame and kettle topography, much of which is formed of very coarse gravel (text-fig. 4). Immediately to the north occurs a second outwash plain, which is interrupted by kettle holes in its western part. This plain is mostly covered with buildings and levelling was not attempted, but Ordnance Survey heights (especially those on the 1865 six-inch map when the area had few buildings) indicate a descent from about 125 ft. in the west to about 115 ft. in the east. This outwash thus relates to a lower sea-level than the adjacent, more southerly, outwash plain.

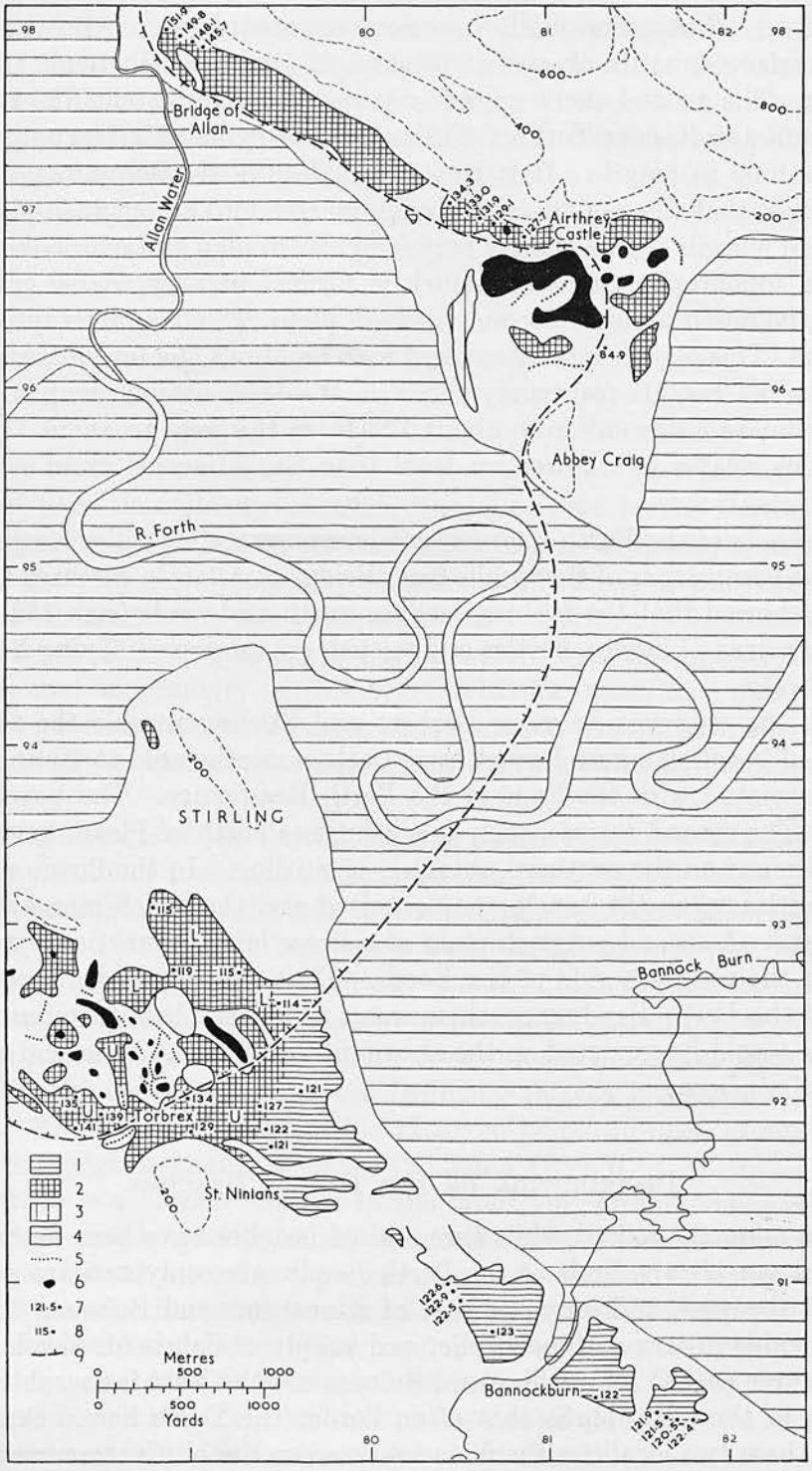
On the north side of the Forth a prominent kame terrace, 1½ miles long and 300–400 yds. broad, follows the basal slopes of the Ochils from Bridge of Allan to Airthrey Castle. Levelling at fifteen points showed that the feature declines south-eastwards from 152 to 127 ft. Here it terminates in an area of kames, kettles and flattish gravel deposits lying between the Ochils and the Abbey Craig.

In summary, the evidence between Larbert and Stirling reveals the following pattern. The highest raised beach can be followed from Larbert northwards to Plean, where it merges into outwash associated with the limit of the Perth Readvance. The beach does not occur north of Plean. The second raised beach does continue north of Plean, however, and can be followed to St Ninians on the southern outskirts of Stirling. In the Cowie area there is clear evidence for dead ice when the beach was deposited and the beach merges into an outwash plain in St Ninians. A second outwash plain at a lower level occurs nearby and a long kame terrace exists on the northern side of the Forth. These features thus imply that, following the maximum of the Perth Readvance, there was a significant halt or minor readvance in the Stirling area. As would be expected on the above interpretation, the second raised beach does not occur west of Stirling.

Distribution of the Raised Beaches

It may seem anomalous that, while three raised beaches have been described in one part of the Forth area outside the limit of the Perth Readvance, only two are present in part of the remainder of the area, and only one east of Kincardine and Bo'ness. This distribution, however, is explicable in terms of local relief and supply of debris for beach formation.

Much of the area east of Kincardine and Bo'ness was the least favourable for raised beach formation owing to the steep bluffs that often border the Forth here. On the other hand, westwards from these two localities the distance between the bluffs increases rapidly and here conditions were more favourable. They were especially favourable on the south side of the Forth between Stirling and Falkirk for here debris was provided by glacial rivers draining from the high ground to the west as well as from the Clyde basin. The optimum conditions occurred in the lower Carron valley, for along this valley were discharged vast quantities of meltwater following the readvance maximum. The slopes of the upper Carron valley and the southern slopes of the Kilsyth hills are trenched by hundreds of meltwater channels for more



TEXT-FIG. 4.—Fluvioglacial forms and raised beaches around Stirling. (1) Raised beach related to upper outwash plain south of Stirling. (2) Outwash plains, kame terraces and flattish-topped kames (U = upper outwash plain south of Stirling; L = lower outwash). (3) Raised beach with shoreline at 73–76 ft. O.D. (4) Carse and associated post-glacial deposits. (5) Kames. (6) Kettles. (7) Points levelled by authors, to nearest tenth of a ft. (8) Points levelled by Ordnance Survey, to nearest ft. (9) Suggested approximate position of ice-margin when kame terrace extending past Bridge of Allan, upper outwash plain south of Stirling, and St Ninians-Bannockburn raised beach were being formed.

than 7 miles west of Larbert: all these channels lead eastwards, the only escape route for their waters being the lower Carron valley. Another important meltwater system, about 10 miles long, fed into the lower Carron area from the south and south-west from the vicinity of the sites of Airdrie and Coatbridge. Thus the extensive raised beach deposits of the Larbert-Falkirk area are readily explicable.

The extent and distribution of each of the three beaches also reflects to a considerable degree the availability of debris from glacial rivers. The lowest beach has only been identified in the Carron delta. The second beach is very well developed here also, but when traced down the Forth becomes feebler until it is lost altogether eastwards from Bo'ness. Northwards from Larbert this beach becomes faint, but then gradually increases in clarity to form a very marked feature from Bannockburn to St Ninians, where abundant meltwater debris from the ice-mass, terminating in the Stirling area, was available when it was being formed. The highest raised beach is a broad feature around Plean, for here it merges into outwash. It remains a well-marked but narrow feature when traced southwards to Larbert, where it becomes broader again in response to the fluvioglacial material from the Carron valley. On the north side of the Forth this beach is best developed towards Kincardine, where it merges with outwash.

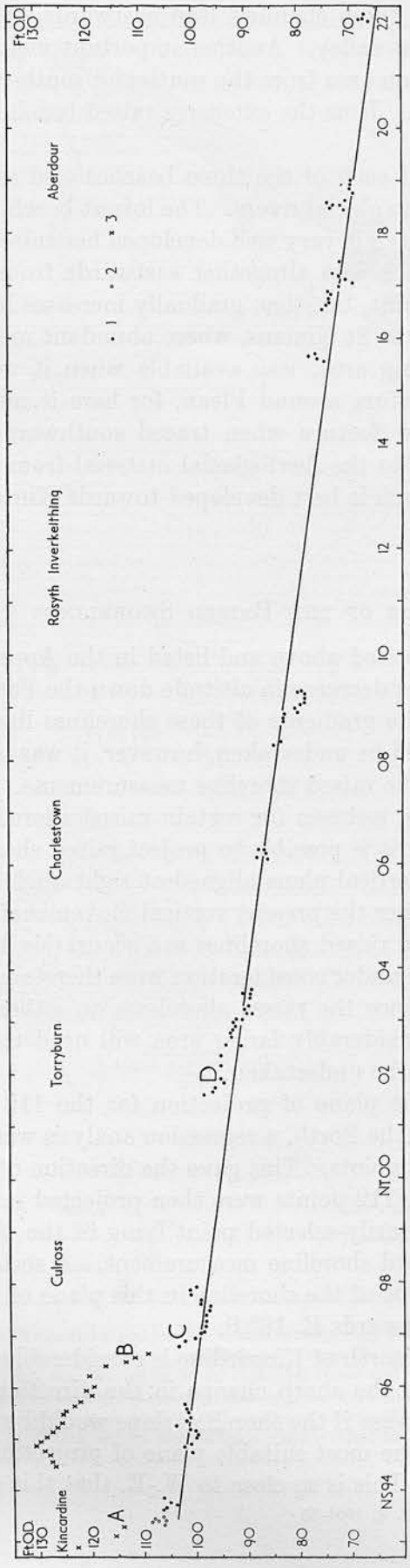
Gradients of the Raised Shorelines

The height measurements discussed above and listed in the Appendix show clearly that the first and second raised shorelines decrease in altitude down the Forth Valley. In order to obtain the best representation of the gradients of these shorelines linear regression analyses were carried out. Before these could be undertaken, however, it was necessary to decide on a plane (or planes) of projection for the raised shoreline measurements.

In parts of Scandinavia, where isobases for certain raised shorelines or for current isostatic uplift have been constructed, it is possible to project raised shoreline measurements in a limited width of country into a vertical plane aligned at right angles to isobases. In Scotland this cannot be done, since neither the present vertical movements of the country nor the former movements as deduced from raised shorelines are accurately known. Planes of projection suitable for parts of the area under consideration were therefore adopted. No attempt has been made to draw isobases, since the raised shorelines on either side of the Forth are only a short distance apart. A considerably larger area will need to be studied before the construction of reliable isobases can be undertaken.

To determine the most suitable plane of projection for the 112 measurements on the raised shoreline on the north side of the Forth, a regression analysis was carried out, using the grid eastings and northings of these points. This gave the direction of the best plane of projection as W. 16° N.-E. 16° S. The 112 points were then projected into this plane and their distances determined from an arbitrarily-selected point lying in the plane of projection and to the west of the westernmost raised shoreline measurement. A second regression analysis was then made to determine the slope of the shoreline in this plane of projection. This gave a gradient of 2·22 ft. a mile down towards E. 16° S.

The group of six measurements north of Kincardine is considerably out of alignment with the remaining 106 heights owing to the sharp change in the direction of the old coastline. Calculations were, therefore, made to see if the shoreline slope would be materially affected by omitting them. The alignment of the most suitable plane of projection for these 106 points was found to be W. 6° N.-E. 6° S. This is so close to W.-E. that this direction was used for



TEXT-FIG. 5.—The raised shoreline heights on the north side of the Forth projected into a vertical plane aligned west-east. (1) Shoreline heights used in the regression analysis. (2) Shoreline heights about 3 km. north of Kincardine not used in the regression analysis. (3) Heights on fluvioglacial terraces. Letters A to D explained in text. Distances in km. given by National Grid eastings. The regression equation is $Y = 77{\cdot}011 - 0{\cdot}734X$ (originating at NS 93 km.) and the standard error of the regression coefficient is 0·012.

determining the shoreline slope, since grid eastings of the measured points could be taken directly from the Ordnance Survey six-inch sheets. The slope obtained was 2·17 ft. a mile (text-fig. 5). A further calculation omitting also the four isolated heights near Burntisland gave a slope of 2·22 ft. a mile.

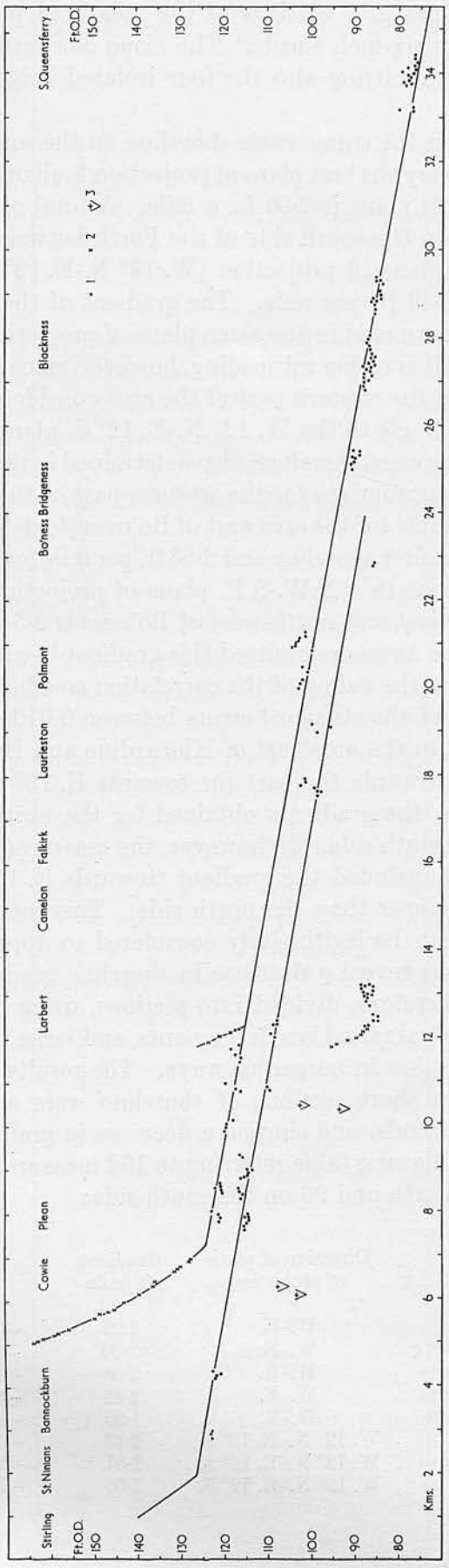
For the sixty measurements on the comparable shoreline on the south side of the Forth between Bo'ness and South Queensferry the best plane of projection is aligned W. 13° N.-E. 13° S. The gradient of this shoreline in this plane is 2·06 ft. a mile. A total of ninety-six measurements was made on this shoreline on the south side of the Forth between South Queensferry and Plean. In terms of the best plane of projection (W. 12° N.-E. 12° S.) the gradient for these ninety-six measurements is 2·43 ft. per mile. The gradient of the 45 points measured on the second shoreline was also determined in the same plane of projection: this gave a slope of 3·32 ft. per mile. This latter result is rather misleading, however, since most of the measurements on the second shoreline are in the western part of the area considered where the general trend of the shoreline is at a sharp angle to the W. 12° N.-E. 12° S. plane of projection. The slopes of the first and second shoreline were, therefore, also determined in relation to a N.W.-S.E. plane of projection. This plane is appropriate for the western part of the area (and hence for the second shoreline), but is less suitable for the area east of Bo'ness (text-fig. 6). The gradients obtained were 2·88 ft. per mile for the first shoreline and 2·83 ft. per mile for the second shoreline.

The gradient of the first shoreline in a N.W.-S.E. plane of projection as calculated from the thirty-six measurements lying west and north-west of Bo'ness is 2·54 ft. per mile. If the five doubtful measurements near the Avon are omitted this gradient becomes 2·67 ft. per mile.

In the above regression analyses the values of the correlation co-efficients ranged between −0·98 and −0·996 and the values of the standard errors between 0·019 and 0·006.

The above analyses show that in the area east of Kincardine and Bo'ness the calculated slopes of the first shoreline down towards the east (or towards E. 13° S. or E. 16° S.) vary between 2·06 and 2·22 ft. per mile, the gradients obtained for the north side being slightly steeper than those obtained for the south side. If, however, the measurements west of Bo'ness on the south side of the Forth are included the gradient (towards E. 12° S.) is increased to 2·43 ft. per mile, thus becoming steeper than the north side. This seemed to suggest that, although the shoreline gradients can be legitimately considered to approximate to uniform slopes, more detailed analyses might reveal a decrease in shoreline gradient down the Forth Valley. The first shoreline was, therefore, divided into sections, using the natural divisions provided by the extent of the individual raised beach remnants, and series of regression analyses were carried out combining the sections in numerous ways. The results of all these analyses were similar, excepting only when short sections of shoreline were analyzed (when local anomalies outweighed the general trends) and showed a decrease in gradient down the Forth valley. This is illustrated by the following table referring to 102 measurements on the highest shoreline on the north side of the Forth and 96 on the south side:

Location and number of Measurements	Direction of plane of projection	Gradient ft./mile	r	S.E.
(1) North side: 102 measurements	W.–E.	2·22	−0·99	0·013
(2) Western 42 measurements of (1) above	W.–E.	2·30	−0·88	0·061
(3) Eastern 60 measurements of (1) above	W.–E.	2·08	−0·97	0·025
(4) Western 57 measurements of (1) above	W.–E.	2·43	−0·94	0·033
(5) Eastern 45 measurements of (1) above	W.–E.	1·95	−0·95	0·043
(6) South side: 96 measurements	W. 12° N.–E. 12° S.	2·43	−0·996	0·006
(7) Western 36 measurements of (6) above	W. 12° N.–E. 12° S.	2·61	−0·99	0·017
(8) Eastern 60 measurements of (6) above	W. 12° N.–E. 12° S.	2·06	−0·98	0·019



TEXT-FIG. 6.—The raised shoreline heights on the south side of the Forth projected into a vertical plane aligned N.W.–S.E. (1) Shoreline heights. (2) Heights on outwash. (3) Heights of kettle rims. Slope of outwash near Stirling based on Ordnance Survey heights. The upper outwash slope near Larbert appears part way along the first raised shoreline owing to the plane of projection of the diagram. For the upper shoreline the regression equation is $Y = 76{\cdot}629 - 0{\cdot}558X$ and the standard error of the regression coefficient is 0·006; for the lower line these are respectively, $Y = 73{\cdot}872 - 0{\cdot}569X$ and 0·017.

Causes of the Shoreline Gradients

It is clear from the broad patterns of the distribution of raised beach deposits in Scotland that by far the most important cause of shoreline tilting is isostatic recovery of the land following its depression by the weight of glacier ice. An additional minor cause of tilting in the Forth area might be subsidence of the North Sea region, which is considered to be taking place in and near the southern part of that sea (Godwin *et al.* 1958, Jelgersma 1961).

Differential isostatic recovery also provides the simplest explanation of the increase in gradient of the highest shoreline up the Forth Valley. Such an explanation has been adopted elsewhere, as, for example, by Goldthwait (1908) for raised shorelines bordering Lake Michigan. In the Forth area it is possible, however, that the explanation of the changing gradient is not quite so simple. Three complicating factors in particular seem available. These are as follows:—

(i) *Availability of Debris for Beach Formation.*—It has been shown that the distribution and clarity of the raised beaches in the area reflects the availability of debris from contemporary glacier ice. It is, therefore, possible that the increase in coarseness and abundance of debris towards the ice-limits resulted in beaches being built up higher in relation to contemporary sea-levels as the ice-margins were approached. Against this being the principal explanation of the changes in gradient is the fact that the measurements on the first and second shorelines show little or no evidence of such an effect where they cross the Carron area, despite the abundance of debris that was being poured into that area when the shorelines were being formed.

(ii) *Tidal Influences.*—Although tidal range normally increases up estuaries this is considered to have had a very minor influence on the shoreline gradients. At the present day mean high water springs is 8·2 ft. O.D. at Dunbar, 9·8 ft. at Grangemouth, 10·5 ft. at Alloa and 10·6 ft. at Stirling (Admiralty Tide Tables 1964), a total increase of only 2·4 ft. from Dunbar to Stirling. At the time the raised beaches under consideration were formed the western part of the estuary, where the present height of high water mark increases most rapidly westwards, was far wider than at present, and it would, therefore, seem likely that the increase in height of high-water mark up the Forth valley was even less then than it is now.

(iii) *Diachroneity of Shorelines.*—It is generally considered that the more marked raised shorelines in an area affected by glacial isostasy were formed when land and sea-level changes were in the same sense and approximately equal for a considerable period of time (*e.g.* Wright 1937, p. 415). Since isostatic recovery of the land is differential it follows that, in theory, any raised shoreline in such an area will be slightly diachronous. It is not known to what extent this applies to the shorelines described in the Forth area, but the possibility exists that they are slightly older in the west than in the east. This could account (or partly account) for the slight increase in gradient westwards.

The Raised Shorelines in Relation to the Perth Readvance and Glacial Isostasy

It has been shown that the highest shoreline merges into outwash north of Kincardine, near Plean and around Larbert and Camelon, and that it does not exist farther west in the

Carron Valley or farther up the Forth Valley despite the presence of suitable sites for its deposition and preservation. This means that while the ice was still at or near its maximum extent there was an upward movement of the land in relation to sea level. It has been deduced that such a movement occurred in the Kincardine area and that it was at least 30 ft. in the Carron area. In the latter area there is additional evidence that permits a greater relative uplift to be deduced. In addition to the Carron terraces already described, there are lower terraces, composed of coarse gravel, with gradients (as shown by levelling) comparable to those of the steeper outwash plains and kame terraces. One of these terraces (D, text-fig. 3) descends to the level of the present flood-plain of the river and appears to pass below it at an altitude of below 60 ft. O.D. These terraces contrast strongly with the typically sandy nature and relatively gentle gradients of the present flood-plain deposits of the river. This strongly suggests that the terraces are formed of outwash laid down during the period that this river carried glacial meltwaters as the ice wasted back from the Perth Readvance limit. This would mean that relative uplift of the land in the Larbert area was at least 60 ft., and probably considerably more, during the time that glacial meltwaters were discharged through the Carron valley.

The evidence in the Stirling area also indicates considerable relative uplift of the land while glacier ice was still present in the area. As already described, the second raised beach, at about 125 ft. merges into an outwash plain in St Ninians. The adjacent, lower outwash plain, however, starts at about this level and slopes down to at least 115 ft. The kame and kettle topography associated with these outwash deposits descends below the 100-ft. contour in places. On the north side of the Forth also, around Airthrey Castle, a kame and kettle topography extends below the 100-ft. contour and the rim of the largest and lowest kettle here is at only 85 ft. Considerable relative uplift is also implied by the difference in the altitudes of the highest raised shorelines east and west of Stirling. For 15 miles west of Stirling the post-glacial carse plain forms an extensive lowland below 50 ft. O.D. Although streams likely to have carried considerable volumes of material during deglaciation (such as the Allan Water, River Teith and River Forth) flow into this sheltered plain, and although the margins of the plain are often favourable for the deposition and preservation of raised beach deposits, the highest identifiable raised shoreline in this area (all have been levelled) lies at 73–76 ft. O.D., this being a small feature situated near Bridge of Allan, close to Stirling. This altitude contrasts strikingly with the broad raised beach a short distance east of Stirling with a shoreline at about 125 ft. O.D. The only adequate explanation of this evidence appears to be that while the ice-margin stood at or very near the Stirling gap there was an upward movement of the land in relation to sea-level of at least 50 ft. If the highest shoreline, which terminates at Plean, is projected to Stirling it lies at 135–140 ft. O.D., implying that between the time of the maximum of the Perth Readvance and the time the ice finally vacated the Stirling area there was a relative land uplift of at least 60–65 ft. To appreciate the significance of this relative uplift it is necessary to consider certain broader aspects of land-sea-level relationships.

The concept of glacial isostasy was first introduced by Jamieson, in 1865, in his classic study of Scottish raised beaches. Apart from Jamieson, only Wright has studied the subject in any detail in relation to Scotland and his views have not been seriously questioned. Wright's views were first enunciated in 1914, elaborated in 1925, and summarized in 1937. He regarded isostatic recovery of the land from glacial loading as a continuous process. This view appears to have been subsequently accepted by all who have studied raised beaches in Scotland and was accepted by the present writers until recently. This view cannot be reconciled, however, with the evidence presented in this paper.

The late-glacial and post-glacial raised beaches of Scotland, as of other important glaciated areas, have resulted primarily from the complex interactions of two types of change: eustatic changes of sea-level reflecting essentially the variations in glacier volume in the world as a whole, and isostatic depression and recovery of the land related to the waxing and waning of local glacier ice. It has long been known that climatic deterioration producing glacier advance (or readvance) results in a world-wide lowering of the oceans as water is transferred to the land masses to build up the glaciers. In recent years, with the advent of radio-carbon dating, it has become possible to correlate past sea-level changes in many parts of the world and thus to construct increasingly elaborate graphs of eustatic changes of sea-level, including their relations to glacier oscillations (*e.g.* Curray 1960, Fairbridge 1961). Implicit in such correlations is the synchroneity of climatic changes as reflected in glacial advances and retreats. Such synchroneity has long been assumed and recent work associated with radio-carbon dating is tending to confirm this assumption. For example, climatic fluctuations in parts of Europe and America dating back to the last interglacial have been compared and found to give very good agreement (Flint and Brandtner 1961).

This pattern may now be considered in relation to the Perth Readvance in the Forth valley. As yet this readvance is not dated by radio-carbon assay and one can only state that it is not younger than 12,000 B.P.: therefore the correlation with glacial events in North America and continental Europe is not known. The Perth Readvance, however, is a major event in Scottish late-glacial history, involving a forward movement of the ice of at least 50 miles in the Clyde area (Sissons 1963*b*, 1964, 1965). Hence its approximate contemporaneity with important readvances elsewhere can hardly be doubted, especially as the later and much less extensive post-Alleröd readvance in Britain is known to correlate with readvances in Scandinavia and North America. Thus one can assume that, as the ice margin moved down the Forth valley and into the Clyde area during the Perth Readvance, world sea-level was falling, to reach a minimal level about the time of the readvance maximum. If, throughout this time, there was continuous isostatic recovery of the land from its depression by an earlier more extensive ice-sheet, then, in the Forth valley the level of the sea in relation to the land should have fallen continuously until the ice was at its maximum extent. Yet the reverse of this occurred for, as has been shown above, the maximum of the readvance corresponded with a high sea-level now represented by the highest raised beach. The only satisfactory way of resolving this anomaly is to postulate considerable downwarping of the land by the ice of the Perth Readvance. Thus Wright's theory of continuous isostatic recovery must be modified. It should be pointed out, however, in fairness to Wright, that, when he first put forward his theory in 1914, the concept of major ice-sheet readvances in Scotland had not been introduced, and it was not until 1926 that an ice-sheet readvance in Scotland was first identified by Charlesworth.

It also appears that the initial response of the earth's crust to glacial loading and unloading was more or less immediate and rapid, contrary to Wright's premise that "the rise of the land when first freed from the ice is presumed to be slow" (1937, p. 415). This is suggested by the correlation of the highest shoreline with the readvance limit, especially when it is borne in mind that, for this to occur (*a*) the depression of the crust had to exceed the amount of contemporary eustatic sea-level fall, and (*b*) isostatic recovery from the more extensive ice-mass that occupied the Central Lowlands of Scotland prior to the Perth Readvance had to be reversed. With reference to this latter point, it may be noted that the down-warping of the land during the Perth Readvance implies that considerable isostatic recovery from depression under the earlier, much more extensive ice-mass had already occurred, for otherwise the

additional load imposed by the Perth Readvance would not have caused depression. As the ice of the Perth Readvance decayed following its maximum, immediate crustal response to glacial unloading is indicated by the absence of the highest shoreline within the readvance limit, as well as by the evidence in the Carron valley and around Stirling, which indicates relative uplift of the land of at least 60–65 ft. while glacier ice was still near these locations. If contemporaneity of glacier oscillations is accepted the *absolute* uplift of the land must have been greater than these figures owing to contemporaneous eustatic sea-level rise.

In relation to the preceding deductions it might be legitimately argued that proof of precise synchroneity of glacier oscillations in various parts of the world is not established. Such a contention is justified, for the methods of dating at present available do not allow absolutely precise chronologies to be built up. Nevertheless, even if major glacial readvances and retreats in various parts of the world were not exactly synchronous, the essential deductions made above would not be invalidated.

Sequence of Events

At some time prior to 12,000 B.P. ice from the South-West Grampians expanded to form an ice-sheet that occupied the Clyde basin and the western part of the Forth region, depressing the land as it did so. The ice reached its maximum extent near the site of Larbert in the Carron valley and near the sites of Plean and Kincardine in the Forth valley. Along part of its limit the ice terminated in the sea and coastal outwash plains merged into beach deposits. These deposits have now been elevated and tilted to form a raised beach whose shoreline slopes down the Forth valley with slightly-diminishing gradient from about 105 ft. near Kincardine to about 65 ft. near Burntisland and from slightly below 125 ft. near Plean to between 75 and 80 ft. by South Queensferry. Following the maximum of the readvance the ice became stagnant, at least in some areas such as the Carron valley, near Cowie and around Alloa and Clackmannan, thus permitting the survival of ice-contact deposits such as kames and kettles. At the same time isostatic uplift of the land was in progress. Subsequently the ice-margin stood at Stirling and fluvioglacial deposits including outwash plains were laid down in the Stirling gap. One of the outwash plains merged with beach deposits, the latter now being represented by a raised beach whose shoreline descends from about 125 ft. near Stirling to between 85 and 90 ft. near Bo'ness. A further shoreline, also merging with outwash has been identified only in the Carron area, where its altitude is between 85 and 90 ft. By the time the ice finally disappeared from the Stirling area and the Carron glacial drainage area, isostatic recovery of the land of at least 60 to 65 ft. had occurred.

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Appendix

Raised Shoreline Heights

For each side of the valley and for each shoreline the heights of the measured points are arranged in the order of their occurrence along the shoreline, commencing with the westernmost or north-westernmost point.

North Side

106·1	NS 9339.9042	106·1	NS 9387.8747	100·8	NS 9544.8638	97·7	NS 9727.8599
108·0	NS 9342.9037	100·4	NS 9473.8690	101·1	NS 9552.8636	97·5	NS 9739.8597
107·3	NS 9342.9032	102·0	NS 9486.8679	101·4	NS 9560.8635	96·9	NS 9747.8598
108·5	NS 9339.9021	100·7	NS 9491.8675	101·9	NS 9569.8632	96·8	NS 9753·8601
105·9	NS 9340.9012	99·6	NS 9503.8662	102·9	NS 9680.8609	98·6	NS 9760.8602
106·3	NS 9348.9005	99·3	NS 9510.8657	101·7	NS 9689.8607	98·5	NS 9767.8603
106·2	NS 9363.8771	99·8	NS 9518.8653	99·8	NS 9696.8606	100·5	NS 9779.8599
105·7	NS 9368.8766	100·9	NS 9520.8648	98·6	NS 9704.8606	100·7	NS 9788.8599
103·7	NS 9372.8758	103·5	NS 9528.8644	98·0	NS 9708.8606	99·4	NS 9794.8600
104·9	NS 9378.8752	101·5	NS 9538.8640	97·8	NS 9719.8605	97·9	NT 0160.8643

96·1	NT 0167.8642	89·2	NT 0330.8521	79·3	NT 0903.8300	70·5	NT 1713.8393
95·8	NT 0174.8644	89·0	NT 0345.8501	78·5	NT 0908.8298	71·9	NT 1719.8391
94·0	NT 0183.8644	88·9	NT 0348.8496	78·5	NT 0912.8298	73·1	NT 1724.8388
94·6	NT 0191.8648	88·6	NT 0352.8488	78·5	NT 0920.8297	71·8	NT 1746.8381
94·6	NT 0215.8646	90·4	NT 0353.8482	79·9	NT 0928.8297	73·5	NT 1849.8440
94·6	NT 0234.8642	85·6	NT 0610.8391	79·0	NT 0930.8297	73·0	NT 1851.8442
93·6	NT 0253.8641	87·6	NT 0613.8430	74·8	NT 1557.8265	71·0	NT 1852.8445
90·7	NT 0252.8640	85·9	NT 0621.8422	76·0	NT 1561.8270	71·6	NT 1855.8449
92·6	NT 0274.8586	87·2	NT 0629.8412	77·6	NT 1567.8277	74·1	NT 1859.8452
92·8	NT 0279.8582	86·2	NT 0629.8400	76·4	NT 1571.8281	70·6	NT 1862.8457
92·2	NT 0284.8578	84·3	NT 0826.8344	75·1	NT 1655.8385	69·9	NT 1889.8473
91·9	NT 0289.8570	83·0	NT 0832.8339	74·3	NT 1663.8388	69·1	NT 1891.8480
92·2	NT 0291.8561	81·8	NT 0842.8332	73·8	NT 1669.8390	69·7	NT 1894.8497
91·6	NT 0296.8553	82·5	NT 0870.8308	73·9	NT 1676.8391	69·6	NT 1897.8506
90·5	NT 0303.8549	82·6	NT 0878.8307	73·4	NT 1683.8393	66·2	NT 2203.8611
88·6	NT 0315.8538	82·3	NT 0886.8304	74·0	NT 1690.8394	68·1	NT 2206.8612
89·6	NT 0320.8532	79·9	NT 0893.8302	71·9	NT 1699.8395	67·2	NT 2210.8614
90·0	NT 0324.8527	80·2	NT 0898.8301	69·2	NT 1707·8395	67·1	NT 2213.8616

South Side: First Shoreline

121·1	NS 8356.8702	102·5	NS 9321.7932	88·3	NT 0415.8019	84·5	NT 0555.7931
120·8	NS 8359.8696	102·0	NS 9326.7931	88·0	NT 0419.8018	84·2	NT 0559.7941
121·1	NS 8362.8690	100·8	NS 9345.7927	87·5	NT 0423.8017	84·1	NT 0564.7941
122·6	NS 8364.8682	102·0	NS 9359.7926	87·0	NT 0427.8016	83·8	NT 0599.7925
120·2	NS 8355.8619	98·8	NS 9427.7941	87·0	NT 0432.8014	84·3	NT 0604.7924
120·5	NS 8354.8614	104·3	NS 9448.7950	86·2	NT 0435.8013	85·3	NT 0608.7923
120·7	NS 8354.8609	103·7	NS 9454.7951	86·7	NT 0438.8012	85·3	NT 0613.7922
120·8	NS 8354.8605	103·5	NS 9460.7953	86·4	NT 0443.8011	76·9	NT 1090.7867
122·0	NS 8352.8600	102·9	NS 9464.7954	85·5	NT 0448.8009	80·1	NT 1094.7867
119·8	NS 8430.8489	102·7	NS 9469.7955	85·4	NT 0452·8008	77·0	NT 1098.7867
119·4	NS 8432.8483	101·2	NS 9479.7957	85·9	NT 0456.8006	78·7	NT 1161.7857
119·2	NS 8435.8476	101·3	NS 9483.7958	86·5	NT 0459.8004	77·4	NT 1165.7855
117·8	NS 8437.8469	90·6	NT 0181.8108	86·2	NT 0463.8002	79·1	NT 1168.7853
116·1	NS 8471·8406	91·0	NT 0198.8105	86·6	NT 0468.7999	78·0	NT 1174.7851
117·1	NS 8474.8402	90·2	NT 0202.8104	86·9	NT 0472.7996	77·5	NT 1180.7849
119·5	NS 8484.7993	90·0	NT 0208.8102	87·3	NT 0476.7993	78·1	NT 1186.7848
116·1	NS 8523.8362	90·7	NT 0214.8100	87·1	NT 0479.7990	79·5	NT 1195.7845
116·0	NS 8530.8356	89·0	NT 0218.8099	86·7	NT 0481.7986	78·2	NT 1202.7844
117·3	NS 8534.8347	88·9	NT 0223.8098	85·7	NT 0525.7938	76·9	NT 1208.7843
118·2	NS 8541.8337	90·4	NT 0228.8097	85·5	NT 0529.7946	76·3	NT 1214.7842
116·4	NS 8546.8332	87·8	NT 0399.8025	85·7	NT 0534.7943	76·7	NT 1220.7840
117·3	NS 8555.8289	88·1	NT 0404.8023	85·2	NT 0538.7942	76·1	NT 1226.7839
116·5	NS 8548.8281	88·1	NT 0408.8022	86·5	NT 0544.7942	76·6	NT 1232.7837
102·9	NS 9317.7932	88·0	NT 0411.8020	86·1	NT 0549.7941	76·5	NT 1237.7835

South Side: Second Shoreline

122·6	NS 8043.9098	114·2	NS 8525.8882	115·7	NS 8483.8754	98·6	NS 9002.7976
122·9	NS 8046.9094	114·3	NS 8529.8878	116·1	NS 8475.8758	97·7	NS 9006.7972
122·5	NS 8050.9090	115·1	NS 8532.8874	116·2	NS 8395.8542	98·6	NS 9009.7970
121·9	NS 8160.9022	113·4	NS 8635.8845	115·6	NS 8398.8538	99·3	NS 9015.7969
121·5	NS 8165.9020	114·3	NS 8631.8841	109·4	NS 8488.8414	102·4	NS 9021.7967
120·5	NS 8170.9019	114·5	NS 8628.8837	110·0	NS 8490.8412	101·3	NS 9027.7965
122·4	NS 8179.9018	114·7	NS 8624.8833	108·2	NS 8585.8344	101·4	NS 9032.7964
115·1	NS 8512.8899	115·2	NS 8556.8776	109·2	NS 8590.8340	98·6	NS 9176.7954
115·2	NS 8515.8895	116·1	NS 8552.8769	107·8	NS 8596.8316	95·5	NS 9192.7955
114·7	NS 8519.8891	116·6	NS 8493.8758	107·8	NS 8594.8313	99·9	NS 9200.7953
114·1	NS 8522.8886	116·4	NS 8488.8755	108·7	NS 8591.8309	85·9	NS 9846.8082
						85·5	NS 9850.8082

14

QUATERNARY

J. B. Sissons



Corries on the northern side of Braeriach in the Cairngorm mountains. Photograph by J. K. St. Joseph. Crown copyright reserved.

In 1840 Agassiz, Buckland and Lyell recognised that the Scottish mountains had nourished glaciers but it was not until 1863, when A. Geikie published a long paper on the glaciation of Scotland, that widespread interest was aroused. There followed a flood of Quaternary literature, which not only provided a sound basis for subsequent Scottish studies but also comprised a major contribution to the study of glacial geology in the world as a whole. This golden age culminated in 1894 with the publication of the third edition of J. Geikie's *Great Ice Age*. Since that time, although much of value has been published, Quaternary studies in Scotland have advanced more slowly.

The landforms

During the Quaternary the Scottish uplands nourished glaciers that coalesced on the adjacent lowlands to form piedmont glaciers and, ultimately, ice-sheets, some of which extended at their maxima to East Anglia, south Wales and south Ireland. The main source areas were in the western mountains, especially in the western Highlands. Here are the great majority of the corries, numerous deep U-valleys forming complex interconnected systems resulting from repeated diffluence and confluence of glaciers and ice-streams, many large and deep rock-basin lakes often following faults or shatter-belts with Loch Morar deepest of all at 1,017 ft, and fjords along the whole coastline. In contrast the eastern Highlands have far fewer corries (although they are large and numerous in the Cairngorms) and no major rock-basin lakes, while the only large indentations of the coastline are the Dornoch, Cromarty and Beauly firths (Fig. 14.1). Although deep U-valleys are cut in the Cairngorms and the mountain mass south of Braemar and several major through-valleys exist, the eastern Highlands generally have suffered only moderate glacial erosion and considerable areas of Tertiary plateaux survive in contrast with the narrow interfluves of the west (Linton 1951, 1959). In harmony with this is the survival in north-east Scotland of granite weathered to depths of 30 ft (Phemister and Simpson 1949) and of marine deposits thought to be possibly Pliocene, the

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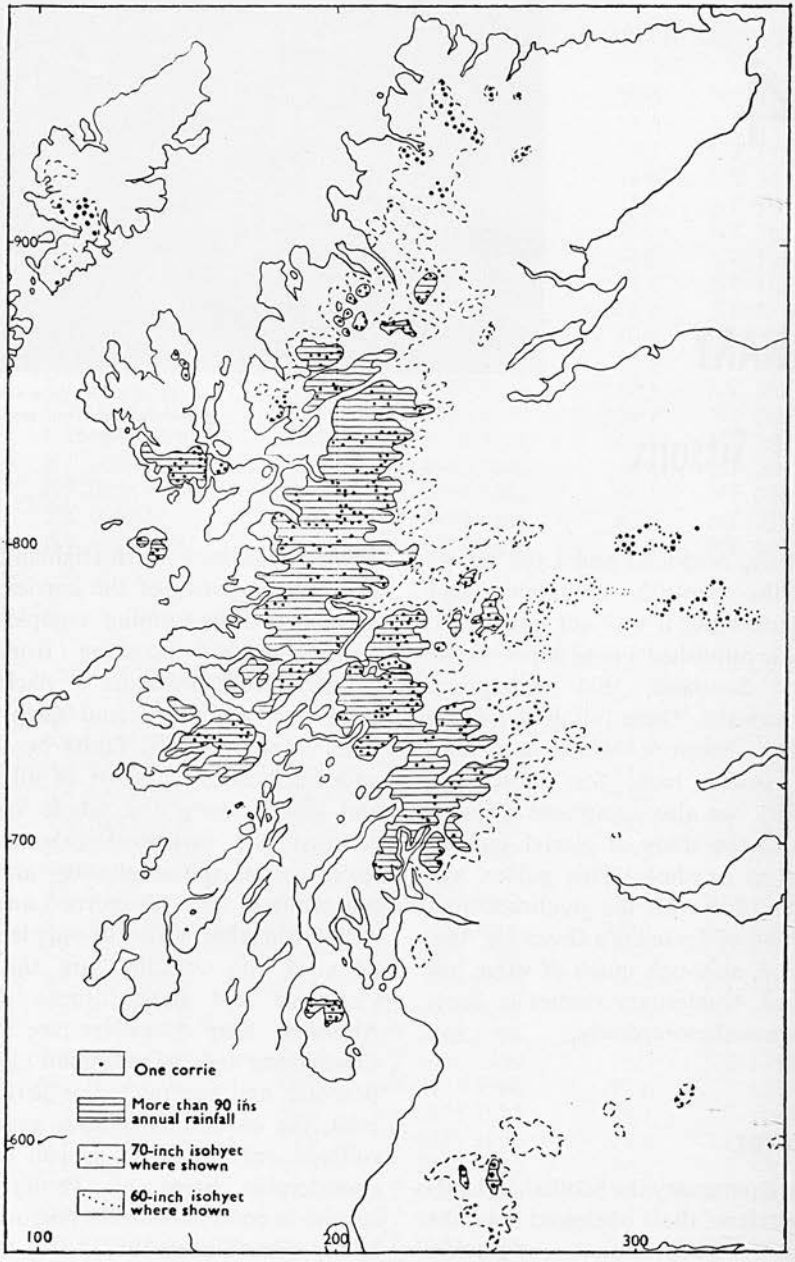


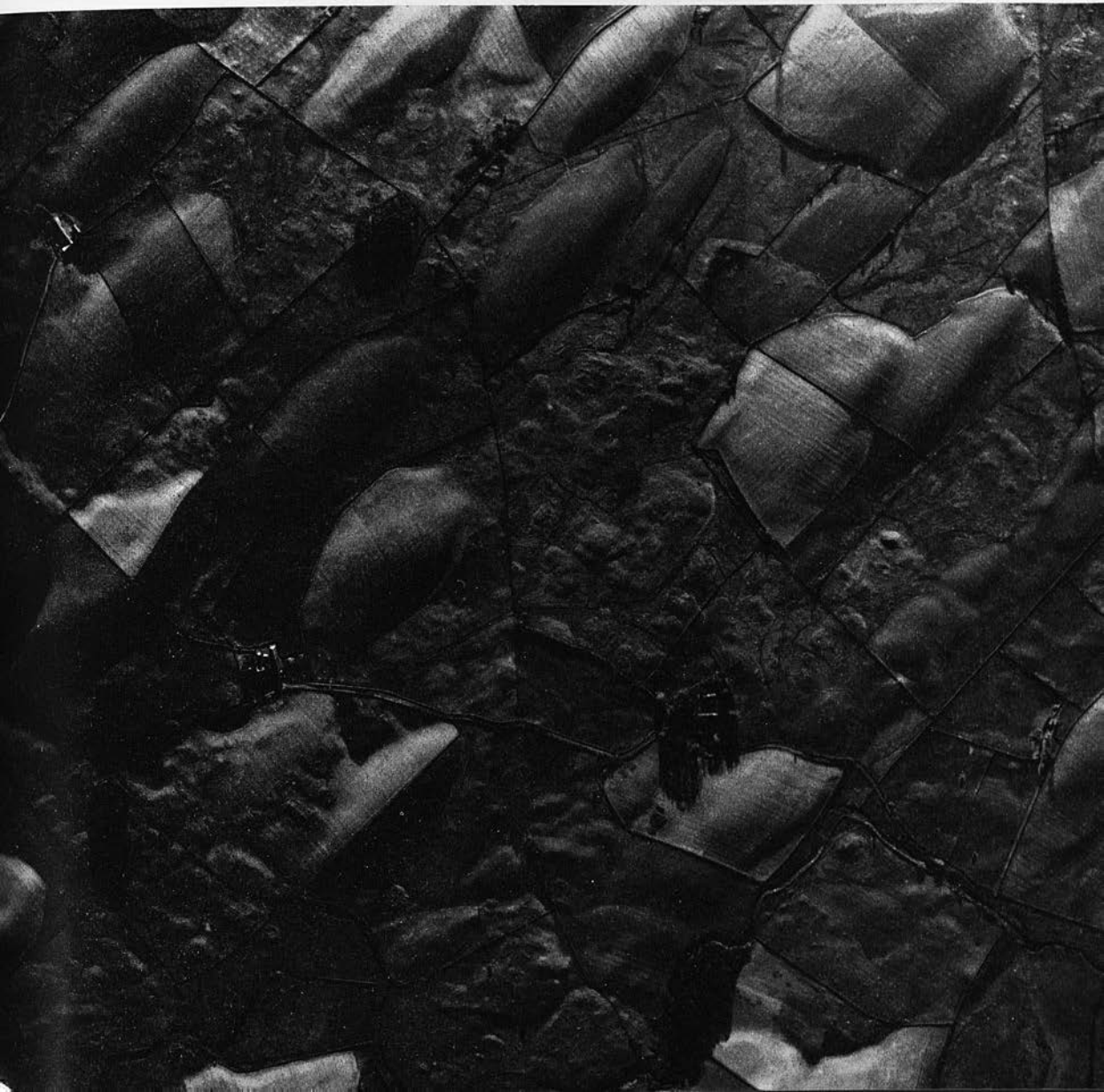
FIG. 14.1. Corries and present mean annual precipitation (after D. L. Linton 1959, p. 22).

greater development of periglacial deposits, and the suggestion by Charlesworth (1955) and Synge (1956) that a part of north-east Scotland was not ice-covered during the last glaciation.

In the Southern Uplands also there is a contrast between west and east, although emphasised by rock type. In the west the granitic and metamorphic Loch Doon region has many corries, several rock basins and much ice-scraped bare rock. In the east, however, the Lammermuirs and Moorfoots, composed of greywackes and shales, lack these features and smooth, rounded slopes prevail.

The form of much of the lower ground in Scotland and some of the higher ground also has been greatly modified by ice-sheets. In the north-west coastlands and the Outer Hebrides a complex relief of rock knobs and small peat-

FIG. 14.2. Drumlins near New Galloway, Kirkcudbrightshire. Royal Air Force photograph. Crown copyright reserved.



or water-filled hollows along with intersecting, structure-guided, linear grooves occupied by streams or lochs, covers extensive areas. In the eastern half of the Central Lowlands almost the whole landscape is strongly ice-moulded along a trend varying from about east to east-north-east. Rock ridges, crag-and-tail forms, rock-cored drumlinoids and occasional true drumlins are all present and the upstanding ice-scoured igneous hills dominate much of the landscape. Although the present form of these various features owes much to the last ice that covered them, in Edinburgh the "tails" of the Corstorphine, Craiglockhart and Blackford Hill "crags" all have bedded sands beneath the surface till (Hall 1812, Croll 1875). In the Tweed Basin the uplands of the Teviot-Ettrick area are strongly grooved along the strike and the trend continues north-eastwards in the large drumlin field around the lower Tweed. The Glasgow area and much of lowland Ayrshire and Galloway are also strongly drumlinised (Fig. 14.2).

The influence of glacial meltwaters is particularly evident in the east, centre and south of Scotland. Meltwater channels over 100 ft deep trench the hill-foot zone of the northern Lammermuirs. Along the Highland margins north of Stirling hundreds of small channels give a rippled appearance to whole hill sides. Deep gorges occupied by present streams, especially in the Highlands, were probably cut mainly by meltwaters. Ice-contact deposits are most conspicuous in the Moray Firth coastlands, the Spey Valley, the northern and southern borders of the Central Lowlands and many of the valleys of south-west Scotland. Long esker systems such as occur in Scandinavia or Canada are lacking, but assemblages of kames and kettles are common, largely a result of the highly-irregular relief facilitating stagnation of the ice.

Moraines composed of angular material (as opposed to kame-moraines) occur in many valleys in the western Highlands and some of the islands but are largely restricted to the valleys of the highest ground in the eastern Highlands and the western and central Southern Uplands. Distinct lateral and terminal features are present in places but more often the moraines occur as innumerable, small, block-littered mounds and hollows that cover the valley floors and sometimes climb high up the hill sides.

Fossiliferous "Interglacial" deposits

In England three interglacials have been so far identified, implying that Scotland has been subjected to at least four full glaciations. No Scottish deposit, however, has been satisfactorily correlated with any of the English interglacial deposits. Excluding the shelly drifts, there are only a few sites where fossiliferous beds have been found between glacial ones. There are also a few places where organic remains have been found incorporated in the glacial deposits.

Near Kilmaurs in lowland Ayrshire the most famous Scottish "interglacial" deposit was discovered about 1816. Stratified deposits up to 36 ft thick, beneath till up to 76 ft thick, yielded 9 or 10 mammoth tusks (including one said to be striated) and a pair of reindeer antlers, and contained molluscs, foraminifers and ostracods indicating arctic or subarctic conditions. The seeds of freshwater plants including *Hippuris*, which is common in British lateglacial deposits, were also obtained. It has been concluded that the mammal and plant remains were washed into the deposits. The evidence points to arctic or subarctic marine conditions (to an altitude of over 100 ft) followed by an advance of ice across the area (Bryce 1865, Young and Craig 1871, Craig 1888, J. Smith 1898, Gregory and Currie 1928, Richey *et al.* 1930).

In Cowdon Glen, 10 miles south-west of Glasgow, a lower till is covered by mud, sand, gravel and peat up to 60 ft thick and these deposits are overlain in places by an upper till. Plant and animal remains from the stratified beds comprise temperate species except for a few plants common in lateglacial deposits. One view is that the upper till is *in situ* and another that it has slipped down the steep slopes of the valley in which the deposits occur (Craig 1874a, Bennie 1893, Gregory and Currie 1928, Richey *et al.* 1930). Craig's detailed descriptions suggest the upper till is not *in situ*.

Several finds have been made near Airdrie: mammoth bones were obtained from bedded sands beneath till (A. Geikie 1863); wisps of peat in till included *Salix herbacea*, now restricted to high mountains in Britain (Bennie 1896); contorted peat in till included *Betula nana* (Dunlop 1888); and bedded sands and laminated clays containing layers of peat, twigs and branches occurred beneath 114 ft of till (A. Geikie 1863). In the Carluke area, 10 miles to the south, a reindeer antler was found in till (Young and Craig 1871) and 48 ft of "surface clay" overlying 4 ft of moss was recorded (Hunter 1882). In eastern Ayrshire a thin layer of plant remains beneath 90 ft of glacial deposits was seen (J. Smith 1898). Near Glasgow remains of woolly rhinoceros and mammoth have been obtained from fluvioglacial deposits and reindeer remains were found in gravel covered by at least 50 ft of till (Proc., 1882; Swinton 1927, Flett 1927, Macgregor and Ritchie 1940). Near the southern end of Loch Lomond a reindeer antler was found at the base of 7 ft of blue clay overlain by 12 ft of till (J. A. Smith 1857).

In the remainder of the country records of this type are few. A. Geikie (1863) described stratified sediments with plant remains between thick beds of till near Hawick and a mammoth tusk in till west of Edinburgh. In Strathmore Campbell (1934) found peaty silt overlying grey till and overlain by several feet of red till: Donner (1960) has concluded that the overlying "till" is a solifluction deposit and the peat beneath probably lateglacial (Zone II) age. Peat in till was observed at two coastal localities north of Aberdeen (Bremner 1938).

Some of the fossiliferous deposits described above have been frequently interpreted as interglacial (and, occasionally, as preglacial) and for the remainder such an origin or derivation has been often implied. When the best-known Scottish deposits were investigated it was not recognised that the decay of the last ice-sheet was interrupted by major readvances. J. Geikie (1894), the main proponent of the interglacial view, wrote at that time, but he emphasised that the "interglacial" deposits immediately precede or are associated with the latest glacial deposits of the areas in which they occur. These glacial deposits are now known to relate to various readvances (Fig. 14.4). Except at the questionable Cowdon Glen locality, the plant and animal remains indicate a cold climate, favouring the view that they relate to the interstadials that separated the later readvances of the last ice-sheet and not to true interglacials. A recent examination of the beetles in a small sample of peat from one of the Airdrie sites accords with this suggestion (Coope 1962).

Glaciation

The sequence of glacial events has been traced farthest back in time in north-east Scotland, owing to its considerable distance from the main ice centres and the labours of Jamieson and Bremner. In this area, near Elgin, a radiocarbon date of 28,140 B.P. has been recently obtained for a podsol strongly affected by solifluction and covered by till (E. A. Fitzpatrick, personal communication).

Outside north-east Scotland no more than two superposed tills of known derivation have been certainly identified in any one area and no deposit has been proved to antedate the late Würm.

A. Geikie (1863) recognised a phase of local glaciation in the Highlands and Southern Uplands following the decay of the *mer de glace*. During the first thirty years of the present century successive stages referred to as the maximum, confluent, valley and corrie glaciations were described in Geological Survey memoirs relating to the Highlands, but no widespread attempt was made to correlate these stages, which do not necessarily have a regional significance. Charlesworth (1926a) was first to identify a major readvance of the ice-sheet, which he termed the Lammermuir-Stranraer Readvance. J. B. Simpson (1933) recognised a readvance in the vicinity of Perth, which he showed to be quite distinct from a later one that produced the end-moraine around the southern end of Loch Lomond. Simpson's readvances were subsequently correlated with events in north-east Scotland by Synge (1956). Meanwhile, Charlesworth (1955) identified two main stages and twenty minor ones throughout the Highlands.

Only Charlesworth has attempted detailed correlations of glacial stages in Scotland as a whole although several variations on his maps have been published by others with little or no explanation. The limits identified by Charlesworth therefore form the basis of Fig. 14.4 but considerable modifications have been introduced. These are necessary because (1) the Perth readvance recognised by Simpson is not considered significant in Charlesworth's scheme; (2) the main correlations in the Highlands were based on the assumption that "100-foot" and "50-foot" raised beaches exist (p. 487); (3) a recent radiocarbon date in the northern Highlands shows that previous correlations in this area are incorrect (Kirk and Godwin 1963); (4) a considerable part of the Lammermuir-Stranraer "moraine" that was interpreted as formed along the margin of Highland ice is now known to be dead-ice deposits of Southern Upland ice (Sissons 1961). The readvance limits shown in Fig. 14.4 will in turn need modification, for much basic mapping has yet to be done, the raised shorelines need to be thoroughly re-studied, and pollen and radiocarbon dating in relation to glacial events have barely commenced.

Maximum Glaciation

The minimum surface altitude of the Scottish ice-sheets can be deduced from records of striae, erratics, and occasionally ice-moulding, on the mountains. It cannot be proved to which ice-sheet this evidence relates but, in view of the past and present intensity of periglacial processes on the mountains, it seems very probable that it relates mainly to the last major ice-sheet.

On the western mountains from Ullapool to Loch Lomond striae and erratics to 3,000 ft or slightly above occur on many summits, the highest record apparently being that of granite erratics over 3,600 ft near Glencoe (*Summ. of Prog.*, 1905). Definite evidence that Ben Nevis (4,400 ft) has been over-ridden has not been found (Bailey *et al.* 1960). Over much of the region north of the Great Glen the ice-shed at the maximum was situated to the east of the mountain backbone, which is breached by about thirty through-valleys up to 2,000 ft deep. In central Ross-shire the movement across the watershed is proved by the north-westward carry to Loch Broom of augen gneiss that crops out not less than six miles east of the watershed. In the same area blocks of Moine schist were transported to altitudes of nearly 3,000 ft on mountains west of the Moine thrust, suggesting an upward carry of as much as 1,500 ft (Peach *et al.*

1913a). The general movement of ice at the maximum, shown by high-level striae, was parallel to the main fjords: north-westerly in the north, becoming westerly towards Mull in the south. The ice-sheet crossed the Minch, incorporating the ice from Skye, and traversed the Outer Hebrides with the same radial trends. It failed to reach St. Kilda 40 miles farther west, but North Rona, 45 miles north-north-east of the Butt of Lewis, was over-ridden and has Torridonian and Cambrian erratics (Wager 1953, Dougal 1932, Gailey 1959). In the mountains of Harris, where the local glaciers were deflected by the ice-sheet, the *roches moutonnées* produced by the latter have a sharp upper limit at 1,600 ft (J. Geikie 1878). In South Uist the upper limit of erratics is at the same altitude (Jehu and Craig 1925). Assuming a maximum surface altitude of the ice-sheet of about 3,500 ft in the northern Highlands, average surface gradients of about 30–35 ft a mile from the mainland to the Outer Hebrides, and over 40 ft a mile beyond the latter, seem likely.

In the Grampians the Rannoch basin was a great centre of ice dispersion at the maximum and for long thereafter. This basin, at about 1,000 ft altitude, is surrounded by 3,000–3,500-foot mountains breached by deep glacial troughs. The Rannoch granite, which mostly crops out well below 2,000 ft, is widely distributed to the west, north and east, reaching altitudes of at least 3,300 ft, and extends far down the Spey Valley (Hinxman *et al.* 1923). South of the Rannoch centre and joining with it, the mountain mass around the heads of lochs Fyne, Long and Lomond sent major streams along these and other lochs into the Firth of Clyde and across Kintyre, Jura and Islay. Over most of the central and eastern Grampians the prevailing movement at the maximum was towards the east and north-east. Ice from the great western centre reached at least 3,000 ft on Schiehallion, striated Ben Vrackie (2,750 ft) to its summit, carried its drift to 2,900 ft on the mountain slopes north of the Garry, deposited Rannoch granite to 2,900 ft east of the Pass of Drumochter and attained 2,850 ft on the northern side of the Cairngorms (Barrow *et al.* 1905, 1913; MacNair 1908; Hinxman *et al.* 1915). These figures suggest a slight eastward or north-eastward gradient of the ice-sheet surface in the central Grampians and imply that the Cairngorms (over 4,000 ft) and many other eastern mountains stood well above its surface. These mountains added their quota to the ice-sheet and supplied especially the ice-stream that flowed along the Dee Valley. This eastern ice, however, was less powerful than that from the west and even the glaciers from the northern Cairngorms were deflected by the western ice.

Ice from the mountains west and north of the Great Glen occupied the Moray Firth. This ice deposited shelly drift with chalk flints and Jurassic rocks in northern Banffshire and Aberdeenshire, where some of the large erratics of Jurassic clay have contained brick pits. The other flank of this ice-stream swung round to move north-westwards across Caithness and Orkney, its drift including marine shells and Cretaceous and Jurassic rocks from the floor of the Moray Firth as well as Highland erratics. A transported mass of Cretaceous sandstone in Caithness measures 240 by 150 yards (Peach and Horne 1880, 1881; Carruthers 1911, Crampton *et al.* 1914).

The Scandinavian ice-sheet is considered to have deflected the Scottish ice across Caithness and Orkney and to have itself over-run Shetland, where one Scandinavian erratic has been found (Peach and Horne 1879, Finlay 1932). Although the zig-zag course of the ice-sheet across Shetland described by Peach and Horne seems rather improbable, as pointed out by Milne-Home (1881), the absence of shelly drifts is not inconsistent with glaciation by foreign ice, for such drifts are also absent

from the Outer Hebrides (except the extreme north). J. Geikie (1878) suggested the Hebridean anomaly was due to deflection of basal ice by the steep submarine slope that follows the eastern side of this island chain, an explanation subsequently applied to Shetland by Bremner (1934a).

The whole of central Scotland was completely ice-covered at the maximum. Along the Highland edge in Perthshire striae reach 3,000 ft and erratics and striae have been found on the summits of the Ochils (2,350 ft), Lomond Hills of Fife (1,700 ft) and Pentlands (1,900 ft). Carboniferous erratics from the lowlands of the Lothians occur on the Lammermuirs (Jamieson 1865, J. Geikie 1894, MacNair 1908, Clough *et al.* 1910). The general eastward and north-eastward movement over most of this area prevailed also over the Southern Uplands from the Loch Doon basin eastwards. The latter was the main centre of ice accumulation in southern Scotland, although a simple radial flow was prevented by the powerful Highland ice to the north and west. Erratics occur on the summit of Merrick, the highest point in the Southern Uplands (2,750 ft), near the Solway on Criffel (1,850 ft) and in Lanarkshire near the top of Tinto (2,300 ft) (A. Geikie 1863, Charlesworth 1926b). It seems likely that the Southern Uplands were entirely ice-covered but that the highest part of the Cheviot (2,700 ft) was not. The granite that forms the central tract of the Cheviot contributed little, but the surrounding porphyrites much, to the English drifts, and erratics appear to be absent from the summit area (Carruthers *et al.* 1932).

Early drifts of north-east Scotland

The earliest glacial deposit identified in the north-east (and in the whole country) is Jamieson's (1906) indigo till containing striated marine shells, which has been observed at several points on or near the coast from north of Aberdeen to south of Stonehaven. In the same area a few scattered Scandinavian erratics have been found and from the basal sands and gravels of a cliff section near Aberdeen more than a dozen have been obtained along with erratics from the north-west. At other points a grey till with north-western or northern erratics overlies the shelly till (Fig. 14.3). From this evidence it has been concluded that Scandinavian ice invaded eastern Aberdeenshire, but that subsequently Scottish ice from the Moray Firth replaced it, this ice being deflected south-eastwards and southwards by the Scandinavian ice still occupying the North Sea (Bremner 1934a, Synge 1956). Synge has suggested that these events were contemporaneous with the Riss glaciation of the Alps.

These early deposits are overlain by till, silt, gravel and sand comprising the "Red Series", which is best developed along the east coast of north Aberdeenshire. Striae, Old Red Sandstone erratics from Strathmore and meltwater channels show that the associated ice-stream moved north-north-eastwards along the coastlands (Fig. 14.3). This Strathmore drift dies out near Peterhead, where it interdigitates with the shelly drift with Jurassic erratics that covers northernmost Aberdeenshire and Banffshire and was deposited by ice moving south-eastwards from the Moray Firth. The interdigitation suggests contemporaneity of the two ice-streams, which are believed to have existed at the maximum of the last ice-sheet (Jamieson 1906, Synge 1956). Some 800 square miles of northern Aberdeenshire, however, remained ice-free. The absence of fresh till from this unglaciated area, along with the existence of till oxidised to depths of over 10 ft and of weathered till overlying deeply-weathered igneous rocks, has led Synge to suggest that the chemical weathering involved occurred during the last interglacial.

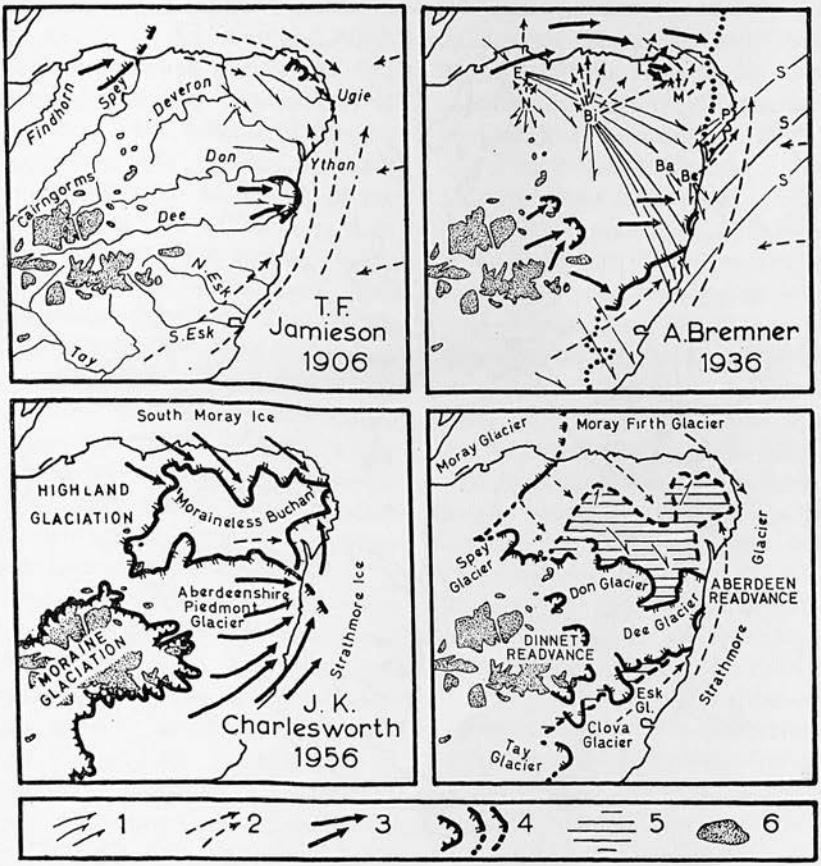


FIG. 14.3. The glacial sequence in north-east Scotland according to different investigators (after F. M. Synge 1956, p. 131). 1. Direction of first glaciation. 2. Direction of second glaciation. 3. Direction of third glaciation. 4. Glacial limits. 5. Areas unglaciated during Würm. 6. Land over 2,400 ft. Bottom right map F. M. Synge's interpretation.

Abbreviations denoting different erratic trains: E, Elgin sandstone; N, Netherley diorite; Bi, Binn Hill gabbro; L, Longmanhill granite; M, Maud norite; P, Peterhead granite; Ba, Barra Hill dolerite; Be, Belhelvie dolerite; S, Scandinavian erratics.

Aberdeen-Lammermuir readvance

The decay of the Strathmore ice in Aberdeenshire was followed after an unknown time interval by an advance of ice from the west down the Dee Valley. This ice extended beyond the present coastline around Aberdeen, where its grey till and associated gravels overlie, replace or incorporate the earlier Strathmore drift. Northward-directed meltwater channels related to the Strathmore ice were wholly or partly obliterated by the later ice (Jamieson 1906; Bremner 1925b, 1934a; Synge 1956). A contemporaneous readvance in the Don, Spey and lesser valleys has been mapped by Synge (Fig. 14.3).

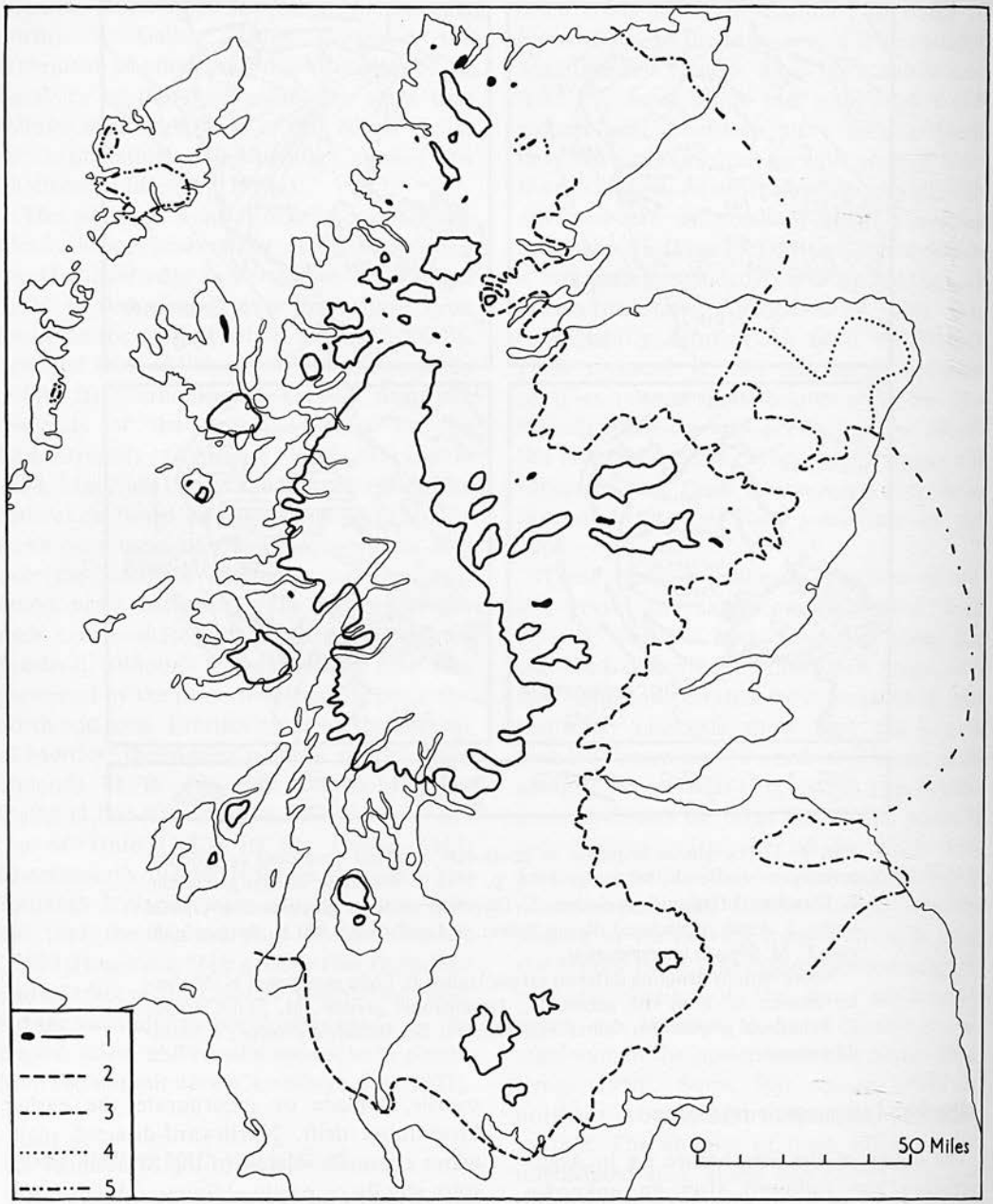


FIG. 14.4. Stages of the last glaciation. 1. Loch Lomond Readvance. 2. Perth Readvance. 3. Aberdeen–Lammermuir Readvance. 4. Northern and eastern limit of "unglaciated" enclave of north-east Scotland. 5. Local glaciation of the Outer Hebrides.

At the same time ice from the Great Glen and the mountains to the west and north occupied the Moray Firth and carried Old Red Sandstone erratics into the uplands of Nairn and Moray. The margin of this ice declined eastwards and appears to have crossed the coastline ten miles east of the mouth of the Spey at Cullen (Fig. 14.4). West of this village the dark, shelly Jurassic drift is replaced or covered by a lighter, more varied drift lacking shells (Jamieson 1906). During the decay of this ice the Spey Valley was blocked to form a lake, which was almost completely filled with horizontally-bedded sands and silts that merge southwards into coarse fluvioglacial materials from the contemporaneous Spey glacier. These deposits have been subsequently cut into numerous river terraces (Bremner 1934a).

In Caithness the shelly till of the maximum glaciation is partly overlain by moraines of local material, while the north-westward-directed striae of the maximum are crossed by later striae pointing north-east and north (Peach and Horne 1881, Crampton *et al.* 1914). This local glaciation appears to be equivalent to the Aberdeen Readvance, for it marks the greatest independent extent of the local ice and seems too extensive for the subsequent (Perth) readvance in view of the eastern location and moderate altitudes.

At the time of the Aberdeen Readvance central Scotland was apparently completely ice-covered (Charlesworth 1955). On the Highland edge of Strathmore, ice from the western Highlands moving north-eastwards deposited red till, up to four miles beyond the boundary fault, implying that the glaciers of the neighbouring mountains were relatively weak (Bremner 1934b). South of Edinburgh a sequence comprising basal Highland till, thick bedded sands with frost wedges at the top, and overlying till that occupies the frost wedges and contains Tinto felsite from the Southern Uplands, covers a considerable area (McCall and Goodlet 1952). It implies deglaciation followed by a readvance during which Southern Upland ice occupied a considerable area formerly covered by Highland ice. The two ice masses were confluent east of the Pentlands. Their limit lay along the northern slopes of the Lammermuirs, declining at 15 ft a mile and passing out to sea at St. Abb's Head. The extension of Southern Upland ice towards Edinburgh shows that a large mass of ice existed on these uplands at this time and suggests that the lower Tweed Basin was occupied by ice. A likely limit is the kame belt in Berwickshire that separates the Tweed drumlins to the south from the frost-shattered kame-free Lammermuir uplands to the north. The southern margin of the drumlin field also corresponds with a large mass of kames and kettles (near Cornhill in Northumberland), beyond which existed a large lake in which was formed a kettle-free delta covering several square miles.

The climatic amelioration that followed the Lammermuir-Aberdeen Readvance caused extensive deglaciation with widespread meltwater activity. Long kame-terraces slant down gently eastwards in Nairn and Moray and are associated with numerous meltwater channels. Similar forms related to the Spey ice occur in the north-western Cairngorms (Bremner 1934c). Dead-ice deposits cover large areas in the Spey Valley and occupy many relatively low areas farther south as in Strathmore, Fife and the Lothians. Over a large part of the central Southern Uplands the meltwater channels trend across the hill tops and ridges on to which they were superimposed from the downwasting ice-sheet. Along the northern flank of these hills in Peeblesshire this ice-sheet wasted back southwards from the Forth-Tweed watershed, resulting in englacial water-tables that were drained with the development of subglacial meltwater flow along the valley floors to the Tweed (Sissons 1958, Price 1960).

The Perth readvance

The glaciers of the Highland valleys bordering eastern Strathmore failed to reach the lowland at this time, but across it their streams formed massive, terraced outwash trains beneath one of which a bed of marine clay with arctic shells has been traced 7 miles inland from Montrose (Howden 1870). The glaciers extended into western Strathmore and Strath Earn, where their outwash overlies varved clay estimated to represent 640 years (J. B. Simpson 1933). The ice covered the lower slopes of the western Ochils, overwhelmed the Campsies except for their highest summits, and in the Forth valley terminated in the sea (then higher than now) below Stirling. Numerous boulders of Lennoxtown essexite on the steep southern slope of the Campsies (Peach 1909) and drumlins on the low ground to the south show that the ice here moved east-north-eastwards. The limit was attained a few miles west of Falkirk. Contemporaneity of this limit with that of the Stirling limit is established by an almost continuous raised beach that merges into the outwash deposits of each (Sissons 1963b, 1964).

The ice that nearly reached Falkirk was part of a mass that flowed up the Clyde Valley past Glasgow and was nourished in the south-west Grampians. Highland erratics (including Garabal Hill granite from the head of Loch Fyne), striae, and the drumlins splaying out from Glasgow prove this flow (Bell 1874, 1893). Under the till deposited by this ice, thick stratified beds have been found at numerous localities from the foot of the Campsies southwards to beyond Lanark. These deposits, especially the stratified ones, infill an extensive system of narrow, steep-sided valleys, those of the Avon and Clyde exceeding depths of 170 and 300 ft, respectively (Stark 1902, Clough *et al.* 1920). In one instance the sides of a buried channel were seen to be considerably weathered (Dick 1870). It follows that before the readvance the ice withdrew from the Clyde Basin and streams cut down deeply. It seems probable, in fact, that the ice completely vacated the Clyde area and allowed the sea to enter the estuary, for marine shells have been occasionally found in the drifts (Crosskey 1865) and ten samples of till from Glasgow all yielded foraminifers (J. Wright 1896). The readvance up the Clyde produced an ice-dammed lake (Bell 1874), in which the stratified sediments, comprising thick beds of sand with interbedded clay and some gravel were deposited, along with occasional organic remains (p. 471). The ice contorted or incorporated these deposits in many places as it advanced, but in others it merely over-rode them and preserved the frost wedges in their upper parts beneath its till. The lake was extinguished when the ice reached the Lanark area and became confluent with Southern Upland ice.

Striae and ice-moulding show that the Highland ice moved south-westwards over the Kintyre peninsula, excepting the southern part, which westerly-directed striae and slopes smoothed by solifluction suggest remained ice-free at this time (F. M. Synge, personal communication). The Highland ice encompassed Arran, deflecting the local glaciers of the northern mountains, and spread eastwards into central Ayrshire. It has been suggested that this eastward movement was not the last in this area (Richey *et al.* 1930), but eastward-directed drumlins, fluvioglacial features, and a widespread upper drift with abundant marine shells and foraminifers to altitudes of over 1,000 ft strongly suggest that the last ice to occupy central Ayrshire moved in from the Firth of Clyde (Craig 1874b, J. Smith 1898, Charlesworth 1926a, Sissons 1964). The shells imply considerable deglaciation and marine incursion before this final readvance.

The shelly drift ends sharply southwards along the borders of the Southern Uplands



FIG. 14.5. Part of the Carstairs esker system, Lanarkshire. Photograph by J. K. St. Joseph. Crown copyright reserved.

and is replaced by a drift with many boulders of Loch Doon granite, showing that south-westward-moving Southern Upland ice existed at the same time as Highland ice deposited the shelly drift (Richey *et al.* 1930). The two ice streams moved parallel with the coast in south Ayrshire and reached their limit near Stranraer (Charlesworth 1926b). Further east the Southern Upland ice spread out on the low ground bordering the Solway, its limit being defined near Wigtown by two sets of mutually-exclusive drumlins that meet at a sharp angle and near Dumfries by a great arcuate belt of kames and kettles spanning the Nith Valley (Charlesworth 1926b).

In north-east Scotland near Dinnet, 35 miles inland from Aberdeen, a large end-moraine occupies the Dee Valley (Jamieson 1874). This has been regarded as equivalent to Simpson's Loch Lomond Readvance, which is generally considered to be the last significant one in Scotland (Charlesworth 1955, Synge 1956, Donner 1957, 1959). However, since well-marked readvance moraines occur near the head of the Dee Valley (Jamieson 1874) it appears that these represent the Loch Lomond Readvance, the Dinnet moraine being equivalent to the Perth Readvance. The massive outwash associated with the Dinnet moraine is typical of this stage.

Further north the position of the Perth limit is uncertain, although the extensive fluvioglacial material in the Beauly, Cromarty and Dornoch firths may well have been deposited mainly at this time. Along much or all of the west as far south as southern Kintyre the ice reached its limit beyond the mainland coastline. In Mull, where the local

valley glaciers appear to have reached their maximum independent limits at the Loch Lomond stage, the last deflection of the local ice by Highland ice revealed by the distributions of local materials and Highland erratics (Bailey *et al.* 1924), can be therefore equated with the Perth Readvance.

The Perth Readvance was followed by rapid downwastage of the ice leading to stagnation in many areas. Dead-ice topography is well developed especially around the heads of the Beauly and Cromarty firths, in the Dee Valley above Dinnet, and along the Tay Valley where terraces above Dunkeld have been successively interpreted as marine, fluvial, lake and kame terraces (Bremner 1939). Extensive kame-and-kettle topography demonstrates ice stagnation on the floor of Strath Allan and in the Campsie valleys, while hundreds of submarginal meltwater channels are cut into the surrounding slopes. In the middle Clyde Basin dead-ice topography is well developed (Fig. 14.5) and shallow lakes formed amidst the decaying ice: the Clyde meanders over lacustrine and alluvial deposits for many miles before plunging into the vertical-sided gorge it has excavated in place of its buried channel. Ultimately the whole country probably became free of ice.

The Loch Lomond readvance

A large arcuate end-moraine surrounds the southern end of Loch Lomond and merges into small lateral moraines that ascend obliquely the mountain slopes bounding the loch. At the same time as this moraine was formed a piedmont glacier occupied the adjacent upper Forth lowlands and produced an arcuate end-moraine more than 12 miles long and over 100 ft high in places. The valleys radiating seawards from the south-west Grampians also contained glaciers, whose limits from near Oban to Loch Awe are defined by end-moraines from which outwash terraces lead down to the coast.

Donner (1957) concluded from pollen studies at several lake sites inside and outside the moraines in the upper Forth and Oban areas that these moraines were produced by a readvance during the Zone III cold phase that followed the milder Alleröd (Zone II) phase. Radiocarbon dates show that the Alleröd lasted from about 12,000 to 10,800 B.P., the succeeding cold phase ending about 10,300 B.P. At Loch Droma, however, on the main watershed inland from Ullapool, a radiocarbon date of 12,800 B.P. has been obtained for a layer with arctic plant remains lying in the meltwater silts of a former lake (Kirk and Godwin 1963). Loch Droma lies far within the readvance limit mapped by Charlesworth (1955) as correlative with the Loch Lomond Readvance and that Donner (1957) considered to be Zone III age. As Godwin and Willis (1961) have pointed out and Donner (1962) has admitted, modification of this view is therefore required.

Less than one mile to the north of Loch Droma a "great terminal ridge, upwards of 100 ft in height" crosses a side valley "like a rampart". In a valley to the south a sea of moraines terminates one mile short of Loch Droma (Peach *et al.* 1913a, p. 101 and Fig. 7). Similar fresh moraine accumulations occupy other mountain valleys of this area and prove an important readvance that failed to reach Loch Droma. If Donner's dating of the moraine limit in the south-west Grampians is correct it therefore seems likely that these moraines near Loch Droma are its northern equivalent. This assumption has been adopted in drawing the limit shown in Fig. 14.4. This limit corresponds with that of the extremely-fresh, hummocky morainic topography that occupies numerous mountain valleys in the Highlands and Southern Uplands (Fig. 14.6) and was recognised as representing a readvance by the Geikies (1863, 1894). In many valleys the moraine limit is very sharply defined. It has been mapped from aerial photographs,

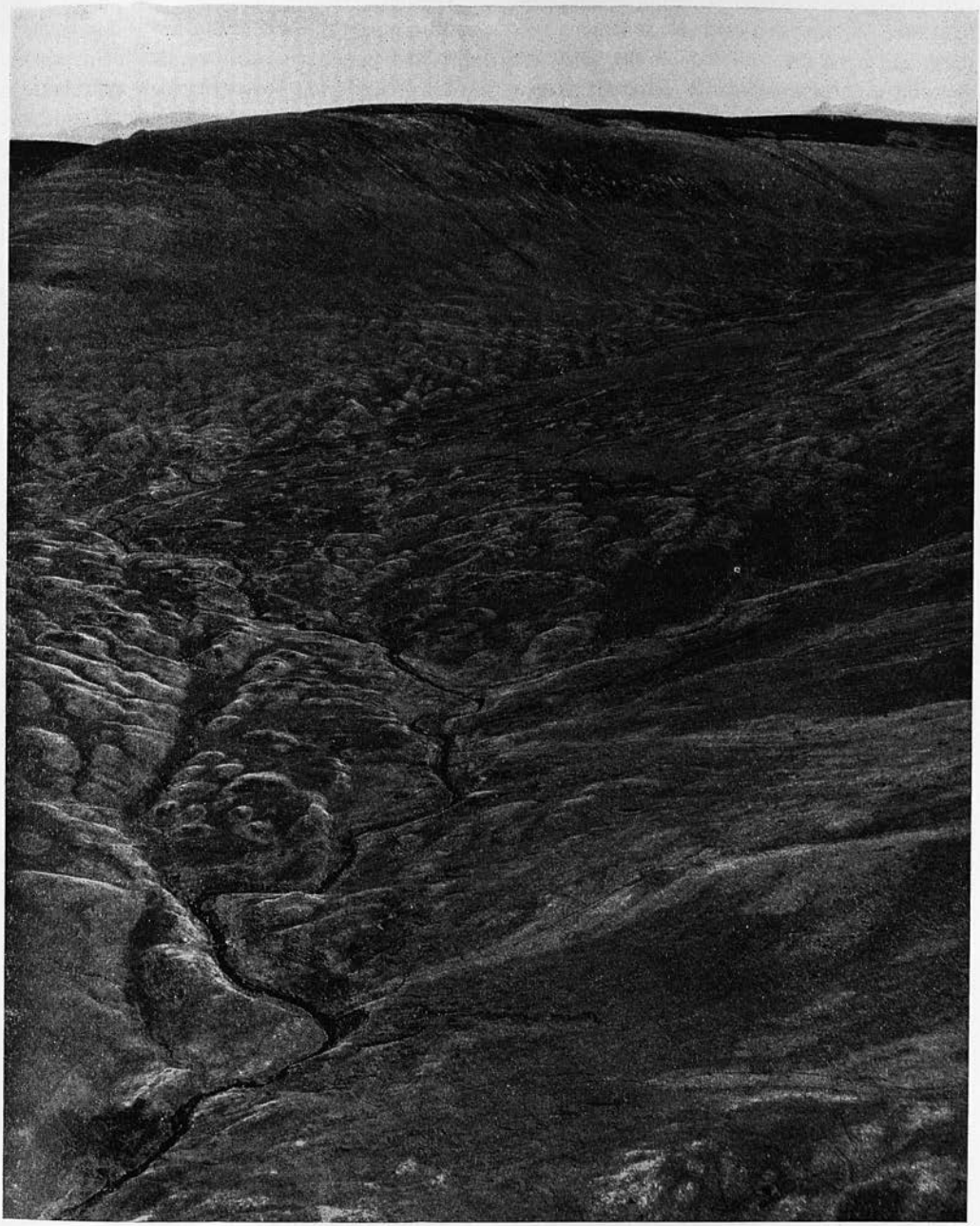


FIG. 14.6. Hummocky moraines of presumed Zone III age, valley of Allt Ollach, Perthshire. Photograph by J. K. St. Joseph. Crown copyright reserved.

field study and descriptions in numerous publications. Where such evidence is lacking the limit has been interpolated on the bases of altitude and of evidence in adjacent areas. Owing to the scale of the map it is not possible to show separately the hundreds of glaciers that existed at this time: probably less than half the area within the limits shown was actually covered by ice.

In the western Southern Uplands some of the glaciers from the Merrick and Kells ranges flowed into the Loch Doon basin, from which ice had radiated at earlier times. Others, flowing outwards, descended as low as 500 ft. In the central Southern Uplands none of the glaciers descended below 1,000 ft. Farther east the Cheviot, although it attains a similar altitude, lacked glaciers, but masses of frost-shattered debris fill the highest valley heads. In the Lake District and northern Pennines where a similar hummocky morainic topography represents the equivalent readvance, the glaciers descended to about 500 and 1,250 ft, respectively, and pollen and radiocarbon dating imply a Zone III age (Manley 1959, Godwin and Willis 1959).

In Arran, Mull, Skye and Jura some of the glaciers descended to sea-level or to within a few hundred feet of it, as did those of the south-west Grampians. Glaciers from the latter centre also occupied the Trossachs to the vicinity of Callander, extended into the upper Tay Valley, and flowed into the Rannoch Basin from the west, carrying with them Glencoe volcanic rocks in the opposite direction to the earlier carry of Rannoch granite (Hinxman *et al.* 1923). Faint shorelines show that the glaciers held up several small lakes in the Rannoch Basin (Mathieson and Bailey 1925). Further eastward in the Grampians the glaciers were much smaller: even in the Cairngorms they did not descend below about 1,500 ft, as shown by the abrupt down-valley termination of fresh hummocky moraines.

Glaciers from the west occupied the southern part of the Great Glen and advanced up Glen Spean to produce the ice-dammed lakes of the Glen Roy area now represented by deltas and a series of shorelines (Jamieson 1863; *Summ. of Prog.*, 1900, 1954). On the west glaciers extended into some of the sea-lochs and appear to have reached the western ends of lochs Morar and Shiel, where outwash fans exist (*Summ. of Prog.*, 1923, 1937; McCann 1961b). In north-west Scotland, however, the glaciers reached only to the heads of lochs Carron, Kishorn, Torridon and Erribol and do not appear to have reached the intervening sea lochs of this coastline. The four glaciers that did reach the sea were small ones descending steeply from adjacent corries (Hinxman 1893, Horne 1899, Peach *et al.* 1913b). Such small glaciers appear to have been numerous in north-west Scotland at the time of the Loch Lomond Readvance. Their lower limits were very variable, depending on precipitation, aspect and size of accumulation area, but many descended to within a few hundred feet of sea-level. Towards the east, as in the Ben Armine range in central Sutherland, the glaciers did not extend far beyond the corries and the lowest limit was about 1,000 ft.

Local glaciers have affected a large area around the mountains of Harris and southern Lewis as well as parts of South Uist, Orkney and Shetland (J. Geikie 1878; Peach and Horne 1879, 1880; Jehu and Craig 1925). The limits of these glaciers probably correspond with a mainland readvance (or readvances) preceding the Loch Lomond one and the extent of the latter is not known.

Within the hummocky moraine limit in and near some of the corries of the Highlands and Southern Uplands small end-moraines exist. Similar moraines in the Lake District are considered by Manley (1959) to be due to severance during downwastage of the lower parts of the Zone III glaciers from their upper

parts, the latter remaining active slightly longer than the former and producing the moraines. Pollen analysis of lake sediments in a corrie at 2,350 ft on the north-east side of Ben Lawers in the central Grampians shows that deposition of organic sediments began in Zone IV and these deposits are overlain by the remainder of the postglacial sequence (Donner 1962). Thus no glacier existed in this corrie (and, by implication, many others) after the Loch Lomond stage.

The pattern as a whole

The distribution of glacial erosion forms, the relative strengths of ice-streams from the various source areas, and the limits of the readvances all reflect the overwhelming importance of the western accumulation areas, especially the western Highlands from Loch Lomond to about the latitude of northern Skye. At the present day this is the area of maximum precipitation with 100–200 inches on the uplands. Northwards present precipitation is considerably less and the areas of high ground far smaller. Eastwards over the whole country present precipitation diminishes rapidly and a sharp eastward rise in the glacial snowline is indicated especially by the size and lower limits of the glaciers at the Loch Lomond stage.

Marine erosion and deposition

"Preglacial" marine erosion

The term "preglacial" has been generally applied to platforms and cliffs that pre-date part (at least) of the last glaciation. This usage is adopted here. The best of these shorelines occurs in Mull, Jura, Islay and adjacent smaller islands and has been also identified in Ardnamurchan and Raasay (W. B. Wright 1911, Bailey *et al.* 1924, D. E. Smith 1961). It attains its optimum development along parts of the western coasts of the larger islands, where it sometimes comprises a striking seaward-sloping rock platform backed by a cliff up to 200 ft high (Fig. 14.7). In western Jura and Islay the platform is largely covered by shingle ridges, and elsewhere is often masked by other marine deposits of lateglacial age. The platform is frequently ice-moulded, occasionally striated and sometimes bears patches of till. Its excellent preservation in certain areas suggests that it was formed in late rather than early Quaternary times. The break of slope between platform and cliff ranges between altitudes of about 80 and 160 ft, a variation that seems most satisfactorily explicable by faulting or warping. Such an explanation is in accord with the apparent absence of the feature at these altitudes from other western areas favourable for its formation and preservation.

Marine erosion forms of some antiquity occur on many other parts of the Scottish coast. In North Rona a wave-cut rock platform, possibly composite, with a notch at 10–25 ft, is overlain by beach deposits and then by solifluction material containing mainland erratics (Gailey 1959). Near Stonehaven an extensive rock platform at about 75 ft is covered by glacial deposits (Bremner 1925b). A rock platform at a similar altitude at Dunbar is shown to pre-date the final occupation of this locality by ice by a perfect crag-and-tail that descends well below its level. Cliffs washed by the sea at the present time are frequently not a product of postglacial marine erosion (Fig. 14.8). In Caithness some of the geos are partly filled with glacial drift and in parts of north-east Scotland the sea is at present exhuming an old coastline with cliffs, geos and stacks from beneath a cover of glacial deposits (Crampton *et al.* 1914, Walton 1959). Along parts of the Scottish coastline the postglacial beach deposits are backed by a steep cliff cut in rock



FIG. 14.7. "Preglacial" marine platform, Islay. "The Grampian Highlands", H.M.S.O. 1948, pl. VIIIB.

and overlie a rock platform up to 100 yards wide. It is often difficult to believe that such features have been cut in the limited time available since deglaciation and a "preglacial" origin seems likely (McCallien 1937). This could help to account for the fact that well-marked erosional features are associated with the postglacial beach deposits along parts of the coasts of eastern Mull and the western Highlands south of Mallaig, whereas along the coast of the north-west Highlands only limited erosion of rock accompanies these deposits (Bailey *et al.* 1924; *Summ. of Prog.*, 1937). Stephens (1957) has suggested that the rock platform and cliff of the south-west Highland coast may be equivalent to the similar features widely developed in Ireland and there overlain by glacial deposits as well as by postglacial beach deposits.

Deep depressions below sea-level occur in the central lowlands, as along the lower Clyde Valley where bedrock has been reached at more than 200 ft below sea-level in several bores; along the foot of the southern slopes of the Campsies; in the Forth Valley above Stirling where rock-head is 360 ft below sea-level at one point; and in the Forth Valley below Stirling where drift at 675 ft below

sea-level has been recorded (Macgregor 1940, 1941; *Summ. of Prog.*, 1962). These depressions have been often interpreted as of fluvial origin and sub-drift contours constructed accordingly. "Preglacial" sea-levels of 200, 300 or even 500 ft below present have been claimed (Croll 1870, Cadell 1913, Clough *et al.* 1925). However, Soons (1959) has attributed to glacial erosion a closed basin in the Devon Valley (near Stirling) that descends to more than 300 ft below sea-level. Other deep buried troughs of the Central Lowlands were also probably formed by glacial erosion.

Marine deposits of uncertain age

The most puzzling of these deposits is situated at about 500 ft at Clava, a few miles from Inverness. A bed of blue clay 16 ft thick and extending for at least 190 yards is overlain by 20 ft of sand and this in turn by 40 ft of till. The clay contains a great variety of foraminifers and ostracods as well as *Littorina littorea*, *Natica groenlandica*, *Leda pernula* and other shells, which are mostly well preserved and not striated, although sometimes crushed. The majority of the British Association committee that studied the section concluded the clay was *in situ*, but a minority favoured transport by ice, although only because a glacial sea-level 500 ft above present seemed highly improbable (Rept. 1893). The minority view is still supported by the absence of glacial marine deposits *in situ* at this altitude or at any altitude approaching it in the rest of Scotland.

Along the coast of western Kintyre fossiliferous clays up to 200 ft above sea-level covered by thick till are exposed in several glens. The shells range from temperate types such as *Turritella terebra* to arctic types such as *Pecten groenlandicus* (Rept. 1896). The lower part of the fossiliferous clay has been interpreted both as transported by glacier ice and as *in situ*. Munthe (1897) and Jessen (1900) agreed, however, that the upper part is *in situ*, having been deposited in an arctic sea and subsequently covered by the advancing ice. Shell-bearing clays and sands, up to 200 ft altitude, between tills, also occur in southern Arran (Watson 1864).

Fossiliferous clays in the Outer Hebrides are restricted to a part of northern Lewis and comprise a lower till with temperate shells picked up by the ice, a marine deposit with warm forms such as *Ocenebra erinacea* and *Sipho jeffreysianus* and a cold fauna including *Trophon clathratus*, the whole being overlain by a thick upper till (J. Geikie 1878, Baden-Powell 1938). Baden-Powell suggested a sea-level of at least 56 ft above present to account for the middle bed.

Lateglacial seas

Marine deposits, which were accumulating while parts of Scotland were still ice-covered, are widespread in many coastal areas of the mainland and inner islands, but are absent from Shetland and Orkney (except perhaps for a few patches of cemented gravel at 25 ft in Hoy: *Summ. of Prog.*, 1930) and almost the whole of the Outer Hebrides. The associated shorelines show little rock erosion, although till was often cut into or removed. They are essentially depositional features comprising beds of gravel, sand, silt and clay. Contemporaneity of glaciers and raised shorelines is demonstrated by the merging of beaches and glacial outwash and by the absence of certain beaches from areas where they would otherwise have been developed and preserved as in many western sea lochs. Arctic or subarctic marine clays, usually covered by later deposits, extend over large areas in the Forth, Tay and Clyde lowlands. The coldness of the seas is shown by the fossil remains as well as by boulders dropped by icebergs or ice-floes. Some boulders in the clays have barnacles on their undersides or



FIG. 14.8. Bowdun Head, Kincardineshire. Indented cliffed coastline in Old Red conglomerates partly exhumed from beneath a cover of Strathmore drift. Photograph by Aerofilms and Aeropictorial Ltd.

girdling them in the wrong direction, suggesting that they were floated from a shore by sea ice (Jamieson 1865, Robertson 1877b). In the Dee Valley near Aberdeen solifluction appears to be responsible for the interdigitation of marine clays with beds resembling till (S. Simpson 1955). Contortions in marine deposits, sometimes due to glacier ice, but more often to sea ice, have been frequently recorded. In various parts of Scotland both chalk and chalk flints have been obtained from marine deposits. In the eastern parts of the Central Lowlands these erratics are absent from the glacial drifts, thus suggesting transport by floes or bergs from the North Sea area (Smyth 1880, A. Geikie 1902).

For the last eighty years two lateglacial shorelines, usually termed the "100-foot" and "50-foot" have figured in Scottish Quaternary literature. The higher shoreline is usually considered to be isostatically warped and attempts have been made to construct isobases for it (Movius 1942; Donner 1959, 1963). It was considered by Donner to have been formed contemporaneously with the Perth Readvance, but Charlesworth (1955) correlated it with the Loch Lomond Readvance. The "50-foot" beach has been almost always described in rather vague terms and W. B. Wright (1937) suggested that what has been termed the "50-foot" beach may not be everywhere the same feature. It has been sometimes described as locally tilted (e.g. Horne *et al.* 1914) but Donner (1959) concluded it is horizontal. Donner correlated it with the Loch Lomond Readvance but Charlesworth related it to a slightly later stage.

It is now apparent that the division of the lateglacial shorelines into "100-foot" and "50-foot" is incorrect. To a large extent this division has arisen as a result of the frequent practice of referring shorelines to map contours. Furthermore, it has not been sufficiently appreciated that, as ice terminating in the sea decays, debris-laden meltwaters issuing from it will cause the most extensive raised beaches to be formed in the immediate vicinity of the ice-margin. With continued retreat of the ice many such beaches may be produced which, taken together, may appear to form a true shoreline but in fact form a feature that is of constantly varying age along its length. The so-called "100-foot" beach is of this character in many areas (Sissons 1962). At the present time a thorough re-examination of lateglacial shorelines in Scotland is required, and consequently the following account can be only tentative.

Marine deposits laid down at a time when most of Scotland was still ice-covered occur in the north-east, especially in the Aberdeen-Peterhead area, up to about 100 foot altitude. They comprise beds of clay that are typically red, contain scattered stones, are often finely laminated but also interbedded with sand and gravel, and occasionally include arctic shells. S. Simpson (1955) concluded they were laid down during the retreat of the Strathmore ice from the eastern coastlands of Aberdeenshire because large meltwater channels related to this ice terminate at about 100 ft. The absence from the lower ends of these channels of extensive gravel spreads at this altitude suggests, however, that the marine transgression occurred after the channels were formed and favours Synge's (1956) view that the red clays near Aberdeen were deposited by a sea at about 80 ft as the ice began to recede from the Aberdeen Readvance limit. Thereafter relative sea-level fell rapidly in this area, for kettles with rims as low as 50 ft and meltwater channels descending to 25 ft occur in or near Aberdeen (Bremner 1925a, Synge 1956).

In the eastern part of the Central Lowlands, south-west Scotland and the Moray Firth area the extensive deglaciation that followed the Aberdeen-Lammermuir Readvance was accompanied by marine incursion. Kettles and kames at levels below the highest adjacent marine deposits indicate the decay of dead ice

in the sea as in east Fife (A. Geikie 1902, Rice 1961), Morayshire and the Lothians. The diachronous upper marine limit associated with this period of deglaciation falls in height rapidly in the peripheral parts of Scotland. Near St. Andrews the sea was at about 100 ft at the time the ice decayed there (Rice 1961), near Dunbar the marine limit does not exceed 60 ft, while in northern Northumberland lateglacial beaches are absent or at a very low level. In the Rhinns of Galloway and near Glenluce beaches are said to reach 100 ft (Charlesworth 1926b), but on the Cumberland coast 36 ft is the upper limit (*Summ. of Prog.*, 1922). In north-east Ireland also a marine transgression was associated with or followed deglaciation and its upper limit declines westwards and southwards from a maximum of 90 ft in north Antrim (Stephens 1963). In east Sutherland shorelines reach 90 ft around Golspie but are absent from northern Caithness (Crampton *et al.* 1914, Read *et al.* 1925). In the west a sharp westward decline in the upper marine limit is indicated by the absence or low altitude of lateglacial beaches in the Outer Hebrides (beaches at about 20–30 ft are known at a few localities: Baden-Powell and Elton 1936–7, M. C. Chisholm, personal communication).

In east-central Scotland, especially in the Forth and Tay lowlands, marine silts and clays of this period, sometimes red, and containing ice-transported stones, are widely developed. These silts and clays are far more arctic than those of the west of Scotland. *Leda arctica*, *Pecten groenlandicus*, *Cardium groenlandicum* and *Thracia myopsis* are common in the east but rare or absent in the west. Differences in sea temperature such as exist at present are not adequate to explain this contrast, which is essentially a result of the earlier deglaciation of the east (Jamieson 1865). The highly arctic fauna appears to be confined to areas outside the limits of the Perth Readvance.

At the limit of the Perth Readvance outwash sands and gravels merge at a few points into raised beaches formed largely of sand, silt and clay. In the Beauly Firth the highest shoreline associated with the readvance is 80–90 ft (Ogilvie 1923), near Perth between 85 and 90 ft, and south-east of Stirling between 120 and 125 ft. S. Simpson (1948) concluded that sea-level was below 30 ft at Aberdeen. When the ice was at its maximum extent or shortly after, the relative level of the sea fell considerably. In the Beauly Firth, kettles with rims as low as 40 ft occur; near Perth, channels cut by meltwaters from the retreating ice descend to 50 ft; in the Forth Valley the highest shorelines are absent west of Stirling; and in the Carron Valley west of Falkirk the floors of some kettles are 30 ft below the surface of the marine deposits. In the west of Scotland marine deposits between 80 and 130 ft are found at various points on the coasts of the mainland and inner islands from Ayrshire northwards to Loch Carron and central Skye. Their lack of continuity, height relationships and relations to fluvioglacial deposits suggest that many of them represent local accumulations composed essentially of meltwater deposits, sometimes reworked by the sea, laid down originally at or near the margin of the wasting ice during the deglaciation that followed the Perth Readvance.

Fossiliferous marine clays were deposited after the Perth Readvance especially in the Firth of Clyde and the lower Clyde Valley. A thin reddish clay sometimes forms the basal deposit (Robertson 1877b, Macgregor 1941), but the thick deposits of arctic red clay that occur in the east are lacking. The basal clay is sometimes more arctic than the remainder, occasionally containing *Leda arctica* and *Leda pygmaea*. Much the greater part is subarctic, however, and is typified by shells such as *Pecten islandicus*, *Astarte borealis* and *Tellina calcarea* (Robertson 1877a). In the

Glasgow-Paisley area the upper marine limit is 70–80 ft (Dougall 1867, Blair 1893, Mitchell 1952) and pollen analysis at Drumchapel, near Glasgow, indicates a late Zone I age for this sea-level. At this locality the marine clay is overlain by freshwater mud with *Betula nana* and other arctic plants corresponding with the Zone II period, and this in turn by a thin Zone III solifluction deposit (Mitchell 1952).

A relationship between the Loch Lomond (Zone III) Readvance and sea-level has been established at a number of points. Towards the head of the Forth lowlands, near Lake of Menteith, outwash poured through a breach in the Zone III moraine demonstrates that sea-level was less than 40 ft while the ice was still near its maximum extent. The site of Loch Lomond was invaded by the sea before the readvance, as is proved by numerous shells in the moraine. Outwash and a meltwater channel show that shortly after the maximum of the readvance sea-level here was less than 50 ft (J. B. Simpson 1933; *Summ. of Prog.*, 1938). Terraced outwash trains lead down towards the coast from the Zone III ice limit between Loch Awe and Oban and some of these are said to merge at their seaward ends into raised beaches (Bailey *et al.* 1924, Peach *et al.* 1909). The lower terraces may indicate a relative fall of sea-level as the ice retreated. North of Oban at the entrance to Loch Etive a large pitted outwash fan, truncated by post-glacial marine erosion, indicates a sea-level less than 40 ft (McCann 1961a). This fan appears to have been formed slightly within the Zone III limit when the ice-margin had become temporarily stabilised on the rock bar at the entrance to the loch.

In eastern Mull outside the end-moraine of the Zone III piedmont glacier a shoreline at 75–80 ft is said to occur, but inside the moraine this shoreline is absent (Bailey *et al.* 1924). Further west in Mull outwash fans of valley glaciers show that sea-level was below 35–40 ft and at least in one instance (Glen Forsa) the fan was formed after the maximum of the readvance. Similar outwash fans occur at the entrance to Loch Creran, at Corran on Loch Linnhe, and at the western ends of lochs Shiel and Morar (*Summ. of Prog.*, 1923, 1937; McCann 1961a, b) and relate to a sea-level no more than 20–40 ft above present. In north-west Scotland presumed Zone III moraines by lochs Kishorn, Carron, Torridon and Erribol were also formed when sea-level was relatively low (Hinxman 1893, McCann 1961b).

Buried and submerged peat beds

The relatively low sea-level associated with the decay of the Zone III glaciers ushered in a phase partly represented by a bed of peat, whose significance, lying between the deposits of the cold lateglacial seas and those of the warmer postglacial ones, was first appreciated by Jamieson (1865). The peat layer, which varies in thickness from a few inches to a few feet and includes remains of oak, hazel and birch, is best developed beneath the carse clays of the Forth and Tay. Pollen studies in these two areas show that the peat is of Boreal age (Erdtman 1928, Godwin 1961) and radiocarbon dating has given results of 8,300–8,400 years B.P. Although the altitude of the peat bed varies locally, in general it declines away from the upper Forth valley on the east side of Britain. For example, it is generally about 15–30 ft in the Stirling area, about 6–9 ft around Perth, below sea-level in the lower reaches of the Tay and at or below sea-level at Montrose and in Aberdeenshire (J. Geikie 1894). A radiocarbon date of 8,425 years B.P. for peat from the floor of the southern North Sea shows that this area was still land at this time and world sea-level at least 90 ft below present (Godwin and Willis 1959). In north-east Ireland the peat bed, buried beneath postglacial raised beach deposits, has

been radiocarbon-dated as 8,100 years B.P. and in Dumfriesshire, where it is covered by carse clay, as 8,100 years (Godwin and Willis 1962).

The above results present a consistent picture, but a slight complication is introduced in Ayrshire. At Girvan samples from two sites where a peat bed is overlain by raised beach deposits have given dates of 9,000 and 9,350 years, while one site at Irvine has given 9,500 and 9,600 years (Godwin and Willis 1962, Jardine 1964). The rather earlier age for the Ayrshire peats may be due to slightly delayed isostatic recovery here as compared with the other sites, resulting in an earlier postglacial marine transgression and burial of the peats. Only the Ayrshire sites of those mentioned lie within the limit of the Perth Readvance.

Further north in Scotland peat below high water mark has been reported at scattered points including Oban, Loch Linnhe, Loch Alsh, Tiree, Coll and the Moray Firth area (J. Geikie 1894, Read *et al.* 1925) and has been frequently observed in Orkney, Shetland and the Outer Hebrides (e.g. J. Geikie 1867). A peat bed in the island of Soay, near Skye, has been pollen-dated as Zone VI and the overlying marine clays as Zone VII (Harrison 1948). This age seems unlikely for some of the peats observed in the outer islands since the latter lie at the margins of the isostatically affected area or beyond it and probably stood well above sea-level in early postglacial times. In some instances the peats passing below high water mark in these islands appear to be quite recent (Jehu and Craig 1925).

Postglacial raised shorelines

The transgression produced by the final great eustatic rise of ocean level in postglacial times is now represented in the larger coastal indentations by nearly-flat spreads of carse clay, especially extensive in the lowlands of Forth, Tay and Clyde, and along more exposed shores by beaches of sand and shingle that form conspicuous terraces for many miles. Wave-built sand and shingle ridges are common and include elaborate systems of complex recurved spits in the Moray Firth area (Ogilvie 1923). Behind the beach deposits a well-marked cliff is frequently present but often appears to be much older than the deposits (p. 484). Shells, most of which are similar to those of the present shoreline, are abundant in the beaches and some beaches are largely composed of them. A climate at one time slightly warmer than the present is demonstrated by *Scrobicularia piperata* in the carse clays of the east of Scotland and the discovery of *Trochus lineatus*, *Tapes decussatus* and *Patella vulgata* at several points between northern Sutherland and the Firth of Clyde (Jamieson 1865; Crosskey 1867a; Robertson 1877b, c; J. Smith 1893; Peach and Horne 1893; Baden-Powell and Elton 1936–7). Evidence of the presence of man is found in or on the postglacial beach deposits or in caves associated with them. Among other remains the most notable are the whales, especially in the Forth carse clays.

The features described above are usually referred to as the "25-foot" beach although a lower beach, sometimes called the "15-foot" has been identified at many points. Recent work in the Forth Valley shows, however, that there are several postglacial beaches. As long ago as 1867 Dougall described marine terraces in the Glasgow area at 39–40, 31–32 and 20 ft. Stephens and Synge (personal communication) have recently identified three postglacial shorelines in the west of Scotland, which are at 37–41, 31 and 18 ft around Oban.

The first postglacial shoreline, which corresponds approximately with the "25-foot" beach of W. B. Wright (1937), represents the culmination of the major postglacial transgression. The peats beneath, as already noted, indicate that the marine and estuarine deposits

began to accumulate between 9,600 and 8,100 years ago in localities so far dated. The terminal date of formation at one point in the Forth Valley is given by the radiocarbon date of 5,500 years obtained for the base of the thick peat deposits that overlie the estuarine clay at Flanders Moss (Godwin and Willis 1962). This date accords with pollen and radiocarbon evidence in England and eastern Ireland, where the eustatic rise of sea-level has been shown to have been completed by about 5,500 B.P. (Godwin and Willis 1961).

Continued isostatic uplift in Scotland resulted in the abandonment of the first

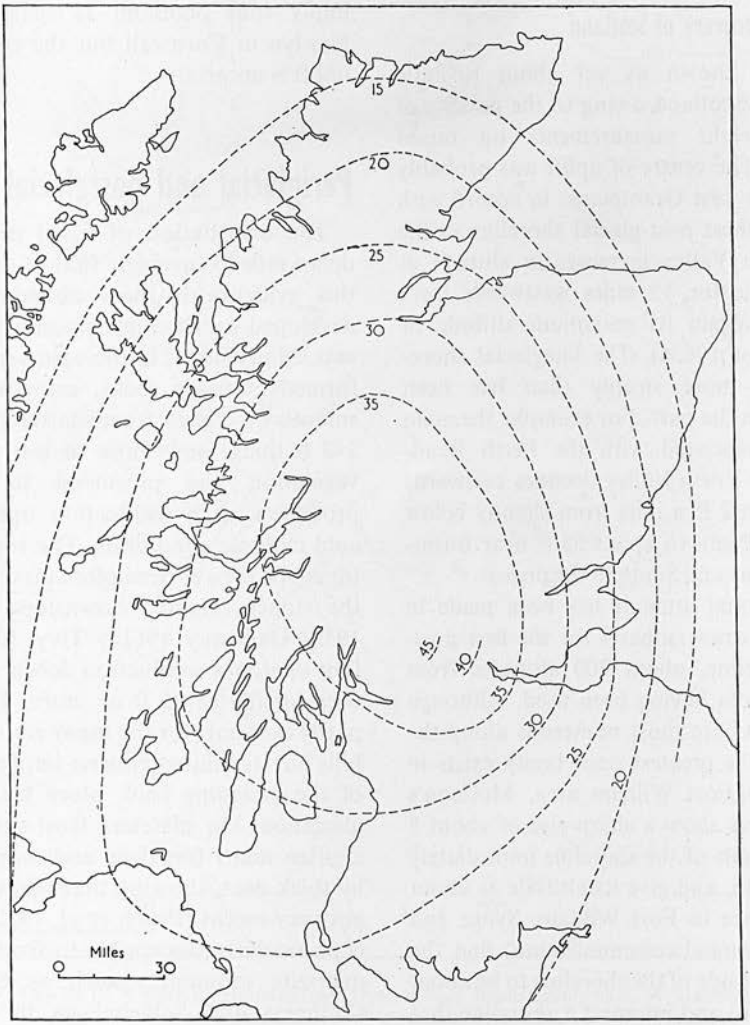


FIG. 14.9. Provisional isobases (ft O.D.) for the first postglacial raised shoreline. Based on numerous sources, especially on unpublished heights obtained by R. A. Cullingford, S. B. McCann (1961b), J. B. Sissons, D. E. Smith, N. Stephens and F. M. Synge.

postglacial shoreline and other shorelines, not yet dated, were subsequently formed. The transgressive nature of the lowest of these shorelines in the Forth area is demonstrated by carse clays interbedded with coarse fluvial deposits and by occasional thin peat overlain by carse clay.

Isostatic recovery of Scotland

Little is known as yet about isostatic recovery in Scotland owing to the paucity of accurate height measurements on raised shorelines. The centre of uplift was probably in the south-west Grampians: in accord with this, the highest post-glacial shoreline in the upper Forth Valley increases in altitude at 0·4 ft a mile for 15 miles westwards from Stirling to attain its maximum altitude of 49 ft (Sissons 1963a). The lateglacial shorelines slope more steeply than has been recognised in the past. For example, the main shoreline associated with the Perth Readvance in the Forth Valley declines eastwards at more than 2 ft a mile from slightly below 125 ft near Plean to about 65 ft near Burntisland (Sissons and Smith in the press).

A provisional attempt has been made in Fig. 14.9 to draw isobases for the first postglacial shoreline, about 200 altitudes from various sources having been used. Although measurements are most numerous along the west coast, the greatest uncertainty exists in the Oban to Fort William area. McCann's (1961b) figures show a sharp rise of about 5 ft in the altitude of the shoreline immediately north of Oban, and give its altitude as about 45–49 ft thence to Fort William. Synge and Stephens (personal communication) find the maximum altitude of the shoreline to be about 41 ft (at Oban) and interpret a shoreline that is at 48–49 ft at Oban as a separate, earlier feature. Synge and Stephen's view has been adopted in constructing Fig. 14.9.

It is not certain whether isostatic recovery in Scotland has ended or is still continuing slowly. Hafemann (1954) concluded that there has been no isostatic recovery since the first century A.D., but Valentin (1953) claimed on the basis of tide-gauge records that Scotland is still rising and produced a very questionable map of the rate of uplift. The geodetic levellings of the Ordnance Survey imply that Scotland is rising relative to Newlyn in Cornwall but the rate of relative uplift is uncertain.

Periglacial and postglacial

The distribution of fossil periglacial evidence reflects inversely that of the glacial, for this evidence is most abundant and best developed on the high mountains and in the east. Many slopes in the east, especially those formed of fissile rocks, are covered with a smooth mantle of frost-shattered rock debris 2–3 ft thick, now more or less protected by vegetation and presumed to have been produced when solifluction operated under cold climatic conditions. The top few feet of till deposits have been often modified also and the stones oriented downslope (Fitzpatrick 1958, Galloway 1961a, Tivy 1962). At the foot of slopes solifluction debris may thicken occasionally to 25 ft or more. This material partly accounts for the many smooth rounded hills but its limited volume implies that much of the rounding took place before the last glaciation. On plateaux frost-shattered rock is often many feet deep and may be covered by thick peat, showing that the weathering is not very recent (Peach *et al.* 1912). Quartzite is particularly susceptible to frost action and quartzite mountains such as Scaraben in Caithness and Schiehallion in the central Grampians are buried in blocky debris (Crampton *et al.* 1914). On slopes, distinctive solifluction forms such as lobes and terraces often occur. Large features of this type, with

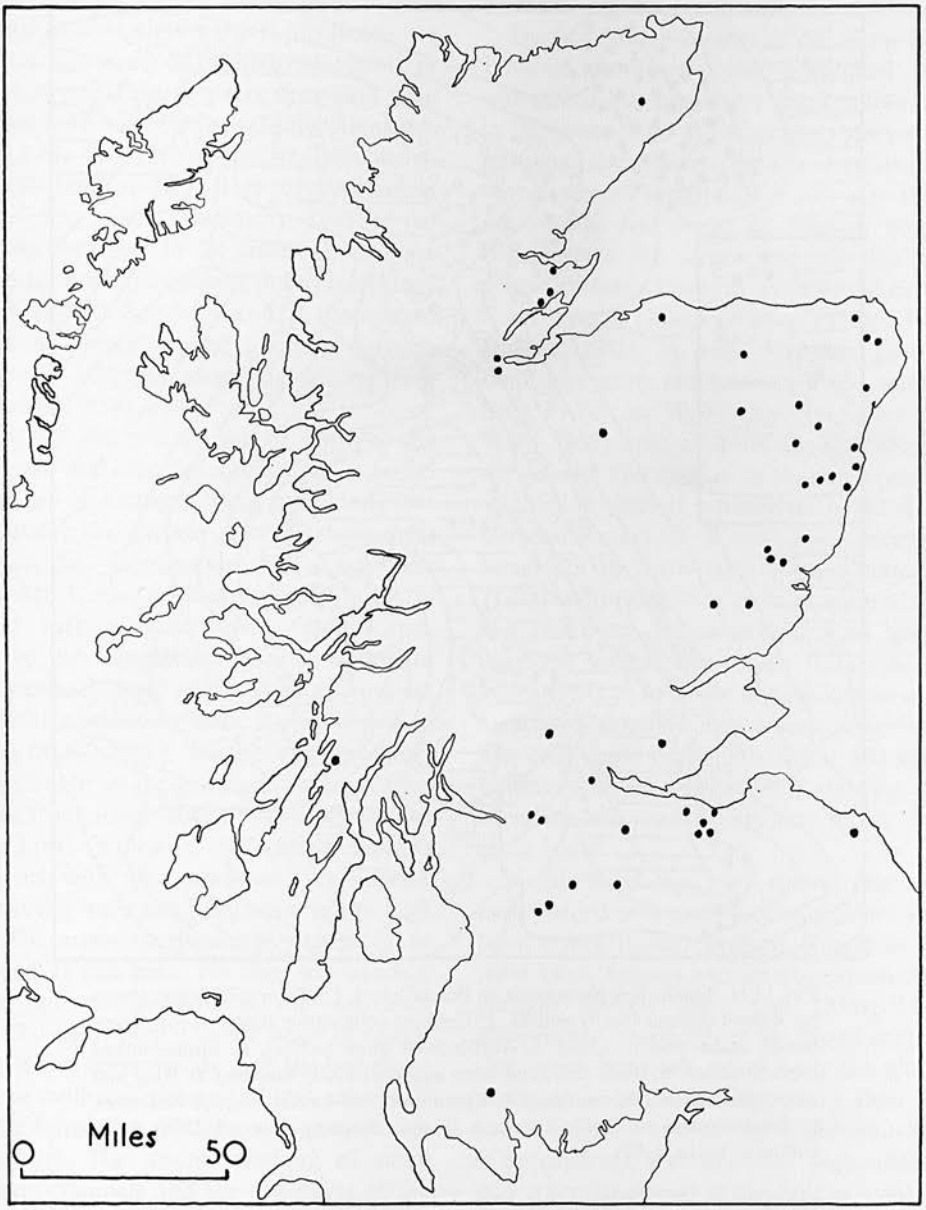


FIG. 14.10. Known locations of fossil frost-wedges. Based mainly on R. W. Galloway (1961b). E. A. Fitzpatrick and J. C. C. Romans (personal communications).

frontal slopes 3 or 4 ft high composed of stones are widely distributed above altitudes of about 1,800–2,000 ft: on the granite mountains of the Lochnagar group they attain heights of 15 ft with boulders up to 4 ft long. These large features are believed to

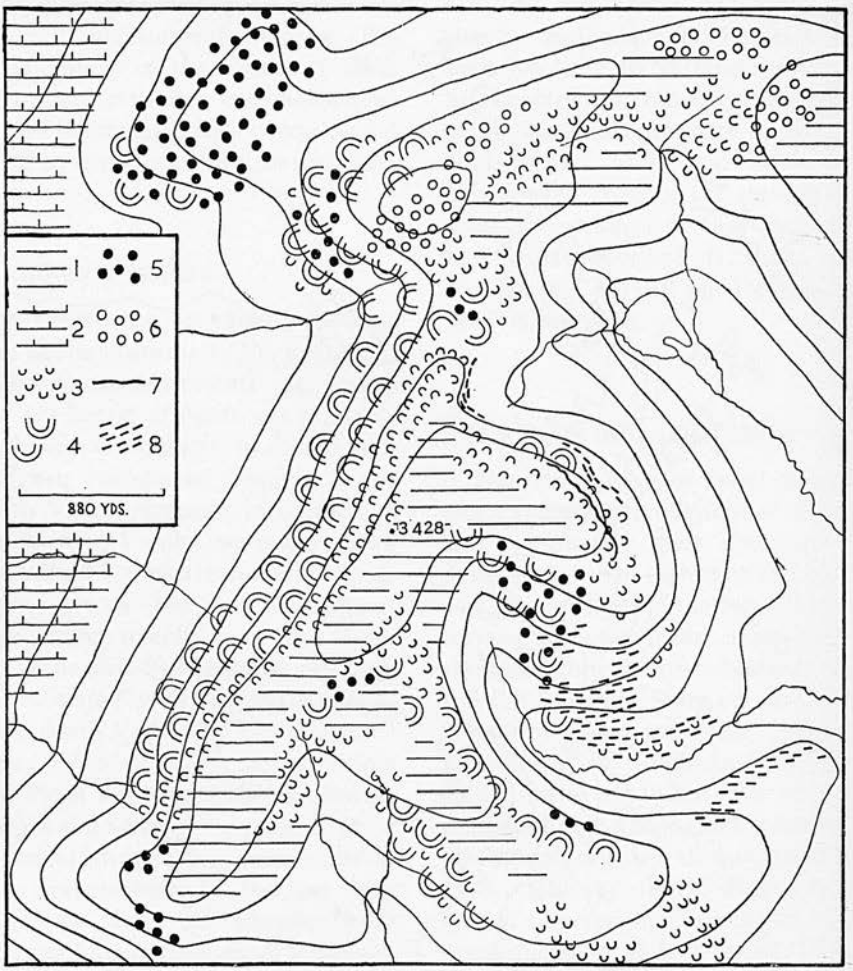


FIG. 14.11. Solifluction phenomena on Ben Wyvis. 1. Uniform solifluction sheets on summit plateau (partly active). 2. Uniform solifluction sheets at low levels buried under peat (inactive). 3. Turf-banked lobes (active). 4. Stone-banked lobes (inactive). 5. Block fields and stone streams (mostly inactive). 6. Ring and stripe patterns in *Rhacomitrium*. 7. Continuous turf-banked terrace (inactive). 8. Discontinuous turf-banked terraces (active). Contour interval 250 ft (after Galloway 1961a, p. 80).

be mainly related to past conditions (Galloway 1961a). Block fields are common where rocks are well-jointed and are now often buried by peat, but slow movement on some slopes is shown by turf ridged up in front of individual boulders. Tors occur especially on granite uplands and are very well developed on the Cairngorms.

Fossil frost wedges, extending to depths as great as 12–15 ft, have been observed in sand and gravel deposits in low ground from the Central Lowlands to north-east Scotland and

Caithness. Their known distribution in part reflects that of sand and gravel pits (many are known in and south of Edinburgh), yet out of more than 40 localities where they have been observed only one is in the western Highlands (Fig. 14.10). Involutions are similarly distributed (Richardson 1874, Tait 1910, Anderson 1940, Gailey 1961, Galloway 1961b). An indurated horizon 18–24 inches below the surface is widely developed in freely-drained soils in eastern Scotland and is thought to have been formed at the junction between permafrost and the overlying active layer (Fitzpatrick 1956).

Although the precise age of most of the periglacial features is unknown, a broad pattern can be deduced. Eastern uplands that stood above the surface of the last ice-sheet such as the Cairngorms, the Lochnagar group, Mt. Keen and Ben Wyvis (Fig. 14.11) abound with well-developed fossil forms. Likewise the "unglaciated" area of north-east Scotland has suffered more intense periglacial weathering than the surrounding country (Synge 1956). The last period of very severe climate in the lowlands corresponded with the Perth Readvance. Frost wedges occur in the Forth Valley in the marine deposits associated with this readvance or formed before it but have not been observed in later ones. The arctic marine clays ceased to be deposited at this time, for they are overlain by the outwash deposits at the readvance limit. The equivalent Dinnet stage in north-east Scotland corresponds with the last phase of severe solifluction there and was associated with the formation of stone polygons almost at sea-level. The limited infilling of small meltwater channels and the sharpness of the fluvioglacial forms within the Perth Readvance limit demonstrate the absence of severe solifluction since they were formed. The distribution of the few finds of mammoth and woolly rhinoceros suggests these animals became extinct during the climatic amelioration that followed the readvance.

The final cold phase that produced the Loch Lomond Readvance was associated with widespread but apparently slight solifluction in the lower parts of the country. In certain lacustrine sites this solifluction is believed to have formed a layer of pink silty clay up to one or two feet thick. At Whitrig Bog in Berwickshire the upper part of this clay contained many leaves of *Salix herbacea* and *S. reticulata* (Mitchell and Parkes 1949, Donner 1957). A silty *Hypnum* peat in Banffshire gave radiocarbon dates ranging from 11,900 to 10,800 B.P. (Godwin and Willis 1959), indicating its accumulation in the Alleröd and suggesting that an overlying solifluction deposit equates with the Loch Lomond Readvance. A similar sequence near Langholm in Dumfriesshire gave dates of 11,800 and 11,600 for a layer of organic mud and 10,800 B.P. for an organic mud lens in overlying solifluction gravels (Godwin and Willis 1964). A few frost wedges appear to be associated with the Loch Lomond Readvance. The cold phase lasted only about 500 years, however (10,800–10,300 B.P.), and the surrounding seas had a subarctic rather than arctic fauna.

In Edinburgh and Fife several lake sites have yielded cold-climate remains from their basal layers, including *Dryas octopetala*, the three arctic willows and the crustacean *Apus glacialis* (Bennie 1891, Tait 1934). These deposits are lateglacial (in the broad sense of this term) but their relation to the glacial sequence is uncertain. The Giant Deer has been found in one of the Edinburgh sites (Corstorphine), the only other authenticated find in Scotland being at Maybole in Ayrshire. This animal became extinct during or soon after the last cold phase (Mitchell and Parkes 1949). In a limestone cave near Inchnadamff in north-west Scotland remains of arctic lemming and hundreds of ptarmigan were overlain by a thin layer of presumed morainic

material and this in turn by abundant animal remains including reindeer and red deer (Peach and Horne 1917). The middle layer may represent the Zone III cold phase.

With the rapid warming of early postglacial times trees, especially birch and pine, spread over much of Scotland and during the climatic optimum forests existed in some mountains to altitudes considerably exceeding 2,000 ft (Durno 1959). Yet periglacial processes were probably active on the highest mountains throughout postglacial times. At present they operate above about 2,000 ft over much of Scotland producing turf-banked lobes and terraces up to 4 ft high, small stone polygons, stone stripes and movement of blocks. Exceptionally these processes are active at lower levels, as on Tinto in Lanarkshire, where stone stripes are forming down to 1,300 ft on the north-facing side. Towards the north-west of the country the level down to which these processes operate descends considerably, probably owing mainly to the high winds inhibiting vegetation growth: present periglacial activity in part of Shetland at 1,200–1,400 ft has been compared with that in the Cairngorms at 3,500–4,000 ft (Miller *et al.* 1954, Fitzpatrick 1958, Galloway 1961a).

Below the periglacial zone the thick vegetation cover slowed down mechanical weathering and mass-movement in the first part of postglacial times, but over the last few thousand years, owing partly to slight climatic deterioration and especially to man's influence, these processes have been accelerated by extensive deforestation. Burning and overgrazing have recently assisted in the reactivation of screes and the production of numerous deep gulleys in the Southern Uplands (Tivy 1957). In the Cairngorms a single storm caused gulleys 30 ft deep in till. Exceptionally heavy rains on deforested slopes of unconsolidated material produce many small landslides and coarse fluvial deposition on adjacent lowlands in many parts of the country (Common 1954, Baird and Lewis 1957).

Postglacial modifications of the landforms as a whole, however, have been slight and the effects of glacial and periglacial processes remain dominant. A mere 10,000 years separates us from glacial times and at the present day the theoretical snowline passes only a few hundred feet above the summit of Ben Nevis. A slight deterioration of climate would therefore establish glaciers once again.

N.B. All altitudes are related to Ordnance Datum.

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Relative Sea-Level Changes between 10,300 and 8,300 B.P. in Part of the Carse of Stirling

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The visible raised beaches of Scotland have been studied for over a century. Another group of raised beaches—the invisible or buried ones—have not been studied at all, for their existence appears not to have been appreciated. This paper is concerned with some of these buried raised beaches in a small area, namely, the western part of the Carse of Stirling.

The term 'buried raised beach' is here applied to raised beaches that are of similar form and composition to some visible raised beaches, but differ from them in that they have been covered by the deposits associated with a later transgression. In the Forth valley burial was accomplished as a result of the major transgression that, according to H. Godwin and E. H. Willis (1962), culminated about 5500 years ago and was associated with the widespread deposition of carse clay. Beneath the carse clay peat frequently occurs, and beneath this are usually found fine sediments comprising grey or pink clay, silt and sand. These fine sediments were interpreted by T. F. Jamieson (1865) as Late-glacial marine deposits. This threefold sequence of carse clay, peat and fine grey or pink sediments is of widespread occurrence in the western part of the Carse of Stirling.

Early accounts and diagrams of the sub-carse peat in the Forth and Tay valleys (for example, Jamieson, 1865; J. Geikie, 1894) invariably represent it as resting locally on a horizontal surface. They thus suggest the possibility that the peat rests on a buried raised beach. Furthermore, it is clear from these accounts that the altitude of the sub-peat surface falls eastward from twenty or thirty feet above sea-level west of Stirling to below sea-level in the eastern parts of the Firths of Forth and Tay. This suggests the existence of a buried beach that has been isostatically tilted in a similar direction to the visible raised beaches of south-east Scotland. To test these possibilities an investigation into the form of the sub-peat surface by hand boring was undertaken. The western part of the Carse of Stirling was selected for study for three reasons: first, the Mentieth moraine, pollen-dated by J. J. Donner (1957) as of Zone III age, crosses the area; secondly, the distance of this area from the open sea should have favoured the preservation of a buried beach; thirdly, a buried beach should here be closest to the carse surface and therefore most easily reached by boring. The explanation of this last point is that the amount of isostatic recovery increases westward in the Forth valley, so that a buried beach, owing to its greater age and consequent greater inclination than the present carse surface, should approach this surface more and more closely in a westerly direction.

Fieldwork

A Hiller peat sampler of the type frequently employed by palynologists was used for the boring. This equipment is only suitable for use in soft or fairly soft deposits, has a limited

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depth of penetration, and being hand operated, is often very laborious to use in some of the deposits encountered in the present area. The sampler is stopped abruptly by coarse gravel, although it can be forced a short distance into sand with small gravel. The sub-carse peat was often difficult to penetrate where the overlying carse clay was more than about twenty feet thick, owing mainly to its being greatly compressed by the weight of the clay, and in some instances it could not be bored through. Owing to the difficulties of extracting the borer and to the likelihood of its breaking, no bore was carried beyond a depth of 33 feet. In order to reduce wear on the equipment, the upper part of the carse clay, which is usually much tougher than the lower part, was opened to a depth of three feet with a soil auger. Despite these and other precautions the equipment had to be repaired from time to time and all the rod connections strengthened with steel bands.

These disadvantages of the Hiller sampler were more than outweighed by the advantages so far as the present study is concerned. The equipment is very cheap compared with power-operated boring rigs, is easily transported, and requires only one person for its operation. Especially important is the fact that it enables the stratigraphy to be determined very accurately. Furthermore, the equipment permits fairly rapid sampling: on a typical day 70 to 100 feet of deposits were bored through and the maximum for a day was over 200 feet. All the boreholes were accurately levelled from Ordnance Survey bench marks, the bench mark heights used being those tied to the second geodetic levelling.

In addition to about 360 boreholes with a total footage of 5800 made by the writer, a large number of bores were put down about fifteen years ago by the Scottish Peat Survey, during an investigation of the surface peat that covers the carse clay over a considerable part of the area dealt with in this paper. Many of these Peat Survey bores are not relevant to the present study since they were stopped on reaching the underlying carse clay. However, in two areas the carse clay is absent and the surface peat and buried peat merge to form a single deposit (J. B. Sissons and D. E. Smith, 1965). In these two areas the Peat Survey bores, numbering altogether about 100, are directly relevant to the present investigation. A few of these bores are located in the north-eastern part of Figure 1. The rest, comprising the great majority, are in the south-western part, where they are located at approximately regular intervals on north-west to south-east lines.

The Highest Buried Step

The general result of the investigation by boring was to show that, beneath the carse clay and peat, is a buried landscape whose essential characteristic is its flatness. Over much of the area the buried surface closely follows the present carse surface and, were it possible to view it, would for the most part present a similar monotonous scene. However, there are small but significant variations within this dominantly flat surface: in fact three distinct steps occur.

The highest step occurs in the south-eastern part of Figure 1, that is, in the area east of the Menteith moraine and south of the River Forth. It forms a continuous feature extending for one and three-quarter miles in a west–east direction with a width of between 200 and nearly 600 yards. At its southern margin the step ends abruptly against the steeply rising ground that also limits the carselands at a higher level. The altitude of the buried step close to its southern limit, as determined at seventeen points, lies between 36.0 and 40.4 feet. Most of these points lie within a much smaller range of altitude, thirteen of them being between 38.7 and 40.4 feet. The surface of the step is interrupted by valleys a few feet deep in places (section GH, Fig. 2).

High Buried Beach

Main Buried Beach

Low Buried Beach

Buried channel of Forth

Outwash gravel

Meltwater channel

X 23·5 Section, with altitude in feet O.D. of top of buried beach deposits

• 31·8 Borehole site, with altitude of top of buried beach deposits

• 17·1 Borehole site, where buried beach deposits, if present, are below altitude given

•(22·7) Borehole site with approximate altitude of top of buried beach deposits

• Borehole site where buried beach deposits not encountered and believed not to be present at greater depth

0 1/4 1/2 3/4 1 Mile

MENTEITH

MORAINE

Present course of River Forth

Arnprior

A B C D E F G H J K L M

99 98 97 96 95

100 200 300

57 58 61 62 63 64 65

FIGURE 1—Buried morphology in the vicinity of the Menteith moraine. The map shows the buried channel of the River Forth and the distribution and altitude of the raised beaches and outwash deposits that would be seen if the carse clay and peat were removed from the area. Letters refer to sections in Figure 2. Where the symbols for outwash and buried raised beaches are shown together the former is known or believed to underlie the latter.

[*facing page 20*

As a whole, however, it usually shows a gentle decline in altitude northward, from typical heights of 39 to 40 feet at its southern margin to about 36 feet at its northern margin.

The buried step is almost everywhere covered by a layer of peat which normally has a thickness of only a few inches. The peat rests on a few inches of grey silty sand, beneath which is much thicker sand that is pinkish or pale brown in colour. The persistent thin grey layer forming the surface of the buried step is presumed to owe its colour to weathering. The pink or brown deposits, which were penetrated to a maximum depth of 13 feet, show variations but are essentially fine-grained water-laid sediments. No laboratory examinations of these or other sediments were made, but visual inspection in the field indicates that fine to medium sand is probably the dominant constituent. In some samples the sediments were seen to be distinctly laminated, comprising numerous alternating layers of clay and fine to medium sand. These rhythmites, which varied in thickness between about 0.3 and 0.8 inches, may well be varves.

The Main Buried Step

At the northern margin of the high buried step, the buried surface falls more rapidly through a vertical interval of a few feet before levelling out as another step, which varies in width up to a maximum of about 450 yards. This main step, as it will be called for the moment, normally slopes gently northward. At its southern margin the altitude varies from about 31 feet in the extreme east to about 34 to 35 feet two miles farther west. The comparable heights of its northern margin are 28 to 29 and 30 to 31 feet.

The main buried step is much more extensively developed in other parts of the area shown in Figure 1. In the north-eastern part of this area it extends one and a half miles in a north–south direction and at least two and a half miles in a west–east direction. Throughout this large area it shows remarkably little variation in altitude, all the heights on its surface except two lying between 30.1 and 34.8 feet.

The main step is also widely developed to the south of the River Forth within the Menteith morainic arc (Fig. 1). Here it varies in width up to three-quarters of a mile and has been followed in an east–west direction for two and a half miles. In this distance its southern margin rises westward from 33–36 feet to 36–38 feet, and its northern margin from about 29 to 31 feet. In addition to this slight rise towards the west there is a gentle rise in altitude southward.

The peat resting on the main buried step is usually at least a foot thick and in some areas is very much thicker than this. The step itself is almost everywhere composed of fine grey sediments. The top few inches to a foot are usually formed of silt or clay with occasional thin bands of silty fine sand. The latter become more numerous downward until they comprise the whole deposit apart from occasional thin bands of silt or clay. The thickness of these deposits is unknown since the boring equipment used would not penetrate more than a few feet into them and no commercial borehole records are available.

The Low Buried Step

Towards the River Forth the buried surface descends from the middle step to level out again as a third and final step. The descent is in some places quite gentle but in at least part of the area east of the moraine and south of the Forth, where lines of closely spaced boreholes

were put down, the descent is sometimes extremely abrupt and was not encountered even when boreholes were spaced only a few yards apart (sections CD and EF, Fig. 2).

Within the morainic arc, the low step is a rather narrow feature in the area studied and lies between about 25.5 and 28 feet. East of the moraine it is much better developed, and in the southern part of this area forms a continuous feature typically about a quarter of a mile broad, lying almost entirely between 21 and 25 feet. North of the River Forth it is rather higher, lying between 25.5 and 27 feet.

The low step is overlain by very compressed peat which is normally one and a half to two feet thick. In some instances this peat could not be bored through, partly owing to its compressed nature and partly owing to the numerous branches and trunks of trees it contains. In these instances the altitude of the step has been estimated, using the average thickness of peat in the vicinity as a guide, and these figures are shown in brackets in Figure 1. It is thought unlikely that these estimated figures are more than a foot in error, The low step itself usually consists of sticky grey silt and clay for the top few feet. In the top two feet or so of this deposit there are innumerable vertical stems of reeds that merge upward into the overlying peat. Downward, the grey clay merges by alternation into grey silty fine sand similar to that forming the main buried step.

At many points along certain stretches of the banks of the Forth the deposits of the low step and the peat that overlies it are visible when the river is at its normal low water level.

The Central Belt

Along a belt varying in width from about a quarter to three-quarters of a mile, and for the most part following the trend of the present meander zone of the Forth, neither buried peat nor the fine grey sediments were encountered in any of the boreholes. Within this belt the carse clay is much thicker than elsewhere. In the lower parts of the strata investigated by boring this clay alternates and is partly mixed with layers of sand. Also in these lower layers marine shells and shell fragments were obtained at several points.

Origin of the Features

One result of the borings described so far is to show that the buried morphological features of the western part of the Carse of Stirling have a simple pattern, comprising three steps and a central linear depression. The three steps are differentiated not only on the basis of their altitude but also by their composition. Thus the highest step is composed of pinkish or brown deposits that, on the whole, are coarser than the deposits of the other two levels. The middle step is formed of fine grey deposits usually with a thin layer of clay in their upper part, while the low step is composed of similar deposits that normally have a much thicker layer of clay in their upper part.

These extensive steps of fine water-laid material appear to be explicable only as lake deposits or as marine (or, more precisely, estuarine) deposits. The former origin can be immediately excluded since the erection of a barrier to produce the required lake would raise insuperable difficulties. In any case, marine penetration into the western part of the Carse of Stirling even before any of the three steps was formed is demonstrated by the occurrence of marine shells in the Menteith moraine, part of which is composed of ice-transported Late-glacial

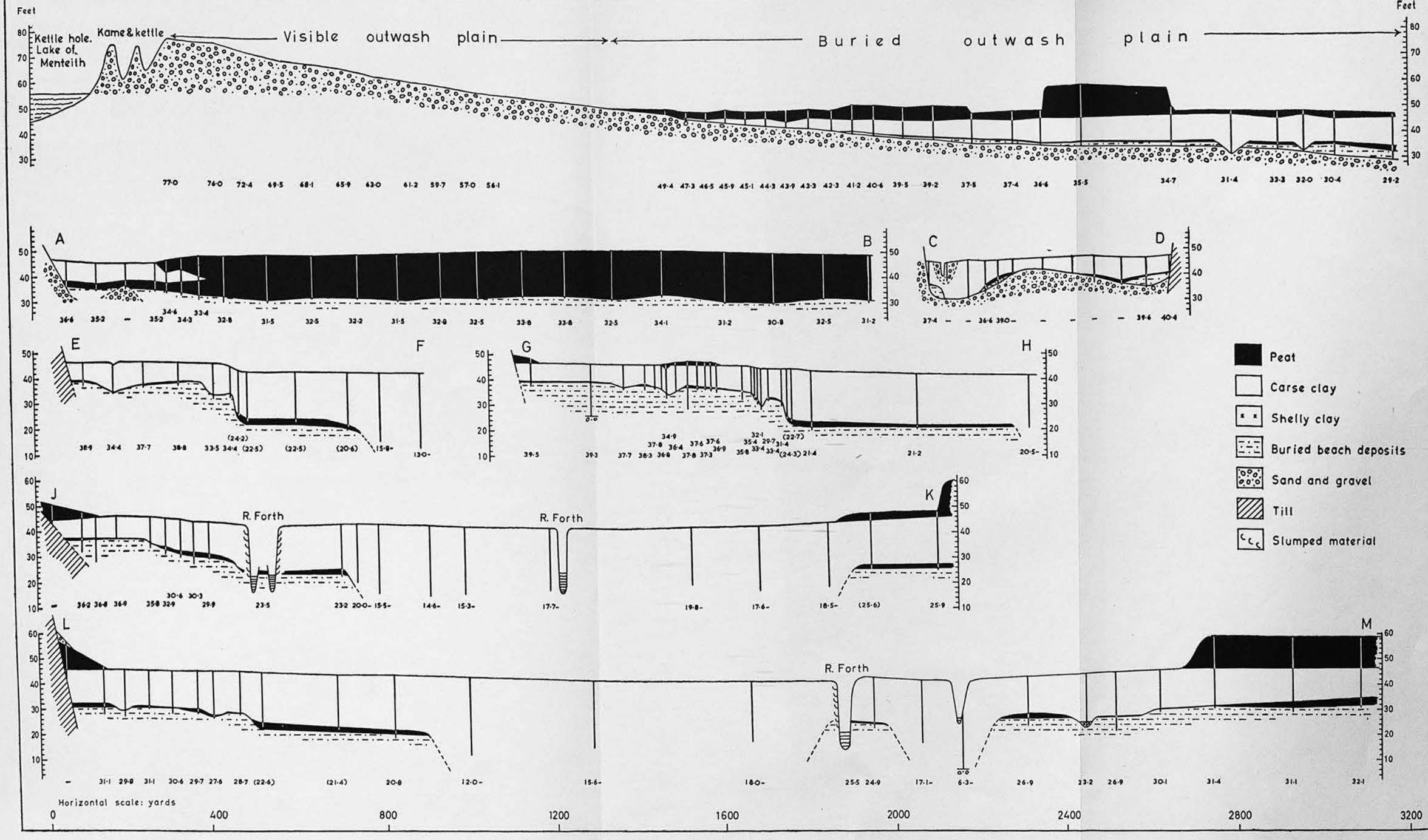


FIGURE 2—Sections of outwash deposits and the buried beaches. Section lines lettered AB to LM are indicated in Figure 1. The figures beneath these sections are altitudes in feet O.D. using the same notation as in Figure 1. In section GH additional data, not shown on Figure 1 owing to the scale of the map, are given. The top section shows the outwash plain that occurs east of Lake of Menteith and the figures here are altitudes in feet O.D. of the surface of the outwash gravel. The irregular distribution and thickness of the surface peat is mainly a result of peat digging by man.

[facing page 33

marine clays. No marine organisms have been found in the deposits of any of the three steps, although microscopic examination has not been attempted.

There can be little doubt that the central belt in which the buried peat is absent and in which the fine grey deposits have not been encountered represents a buried estuary of the Forth. The mixing of clays and sands and the occurrence of marine shells accord with this origin. This old estuary probably looked very much like the present Forth estuary for a few miles upstream of Kincardine Bridge. The uppermost clays of the low buried step as well as the overlying peat have been analysed for pollen by W. W. Newey. The results of this study are described in detail elsewhere in this volume. In the present context it is relevant to note that Newey finds clear evidence of marine influence in the clay and in the basal part of the overlying peat. It appears that the lowest step represents an ancient estuarine mud flat on which salt-marsh developed. Newey also finds evidence of salt-marsh vegetation in the base of the peat overlying the main buried step, thus demonstrating its origin. The thin peat on the highest buried step has not been analysed for pollen, but it would be surprising if this step had an entirely different origin in view of the morphological and stratigraphic evidence. These three steps are therefore regarded as elevated estuarine or marine deposits and will be referred to as (buried) raised beaches. The term 'raised beach', although not strictly accurate, is used since this term has long been applied to both elevated marine and estuarine features in Scotland. The three features may therefore be called the High Main, and Low Buried Raised Beaches.

Altitudes of the Buried Raised Shorelines

The boreholes located closest to the buried shoreline of the Main Beach show that this shoreline increases in altitude from east to west. The relevant altitudes are plotted in Figure 3 and the best-fit line calculated by linear regression analysis is shown. The calculated west–east slope is 1.3 feet a mile (but see p. 15). One may presume that the tilting of the shoreline is largely a result of isostatic warping consequent on glacial unloading. As would be expected, the gradient of the shoreline is greater than that of the younger, visible shoreline marking the upper limit of the carse clay, which slopes eastward at about 0.4 feet a mile (Fig. 3).

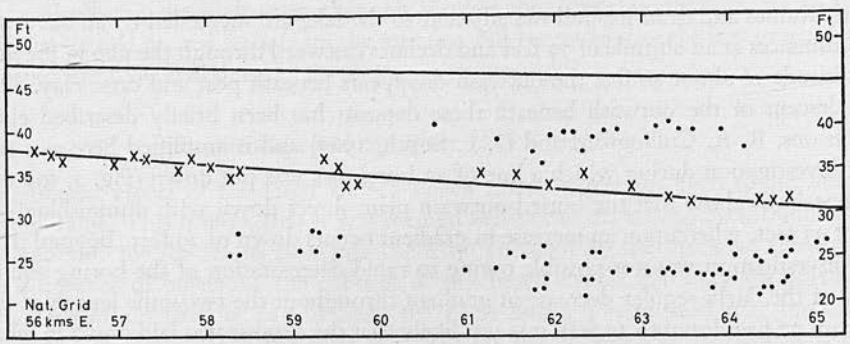


FIGURE 3—Certain altitudinal data relating to the three buried beaches projected into a vertical plane aligned west-east. Crosses give the altitude of the shoreline of the Main Buried Beach as located approximately in boreholes, and the regression line calculated for these points is shown. The upper group of dots relates to the shoreline of the High Buried Beach, while the lower group of dots gives all heights, and estimated heights of the Low Buried Beach. The shoreline associated with the overlying carse clay is shown for comparison, this line being part of a line calculated from 160 levelled heights between Stirling and the western end of the carselands.

The altitude of the shoreline of the Low Buried Beach could be determined at only a few points owing to the difficulty of boring through the overlying buried peat. For this reason all heights (and estimated heights, as explained earlier) obtained on the Low Beach are entered on Figure 3. A decline in altitude eastward is evident, but it is not possible to quantify this in the limited area considered here owing to the discrepancy in altitude of the feature on opposite sides of the Forth. The reason for this discrepancy is not known.

The shoreline of the High Buried Beach is almost horizontal. One possible explanation of this unexpected result is that debris brought by streams from the south contributed to the buried beach deposits in sufficient quantities to mask the effects of isostatic tilting in the small area studied. This possibility is suggested since at the present time four streams drain into this area from the high ground to the south and each has deposited a considerable alluvial fan on top of the carse clay where it emerges from the high ground. It is therefore not unlikely that debris deposited by these streams contributed to the High Buried Beach.

Outwash Associated with the Menteith Moraine

The Menteith moraine comprises a complex series of mounds formed largely of fluvioglacial debris and glacier-transported marine clays that can be traced continuously (apart from one stretch where bedrock and earlier glacial deposits are present) from the northern edge of the Forth carselands towards the southern edge, with an approximately north–south trend. South of the Forth, the morainic belt gradually swings round to an east–west trend, as shown in Figure 1. The moraine is often one-half to three-quarters of a mile wide and its higher parts exceed 100 feet O.D., thus rising some 60 feet above the adjacent carse plain. The moraine is breached by several deep gaps cut by meltwaters, the largest of which is utilized by the Forth. This gap was the only one through which estuarine waters were able to penetrate the area enclosed by the moraine when the buried raised beaches were being deposited and, also, at a later date, when the carse clay was being laid down.

A mile and a half north of the Forth gap, the moraine is breached near the south-eastern corner of the Lake of Menteith, which occupies a large kettle hole immediately within the moraine. Kames and dead-ice hollows adjacent to the lake are succeeded by an outwash plain that commences at an altitude of 77 feet and declines eastward through the gap in the moraine. At an altitude of about 50 feet the outwash disappears beneath peat and carse clay. The continued descent of the outwash beneath these deposits has been briefly described elsewhere (J. B. Sissons, R. A. Cullingford and D. E. Smith, 1965) and is amplified here as a result of further investigation during which a line of 23 boreholes was put down (Fig. 2, top section). These boreholes show that the buried outwash plain slopes down with diminishing gradient to about 33 feet, whereupon an increase in gradient occurs down to 29 feet. Beyond this point further investigation was not possible owing to rapid deterioration of the boring equipment. In view of the fairly regular decrease in gradient throughout the two-mile length of the outwash from 77 feet down to 33 feet, it seems likely that the deposit was laid down in relation to a sea-level at about the latter height. It thus appears that the sea-level to which the outwash was related lies below that associated with the High Buried Beach.

Immediately adjacent to the Forth gap on its south side the moraine is composed largely of fluvioglacial sands and gravels often disposed as kames and kettles. These imply considerable meltwater activity even before the breach now utilized by the Forth began to be excavated. The outwash deposits east of this gap have been located by boring (Fig. 1), but unfortunately

do not permit inferences to be made about contemporary sea-level since they appear to have no simple buried morphological expression. This is probably because, once the ice had wasted back from the moraine, the Forth gap was the only one through which meltwaters could escape eastward, so that any outwash plain that existed was probably largely destroyed. This destruction continues very slowly at the present time, for outwash gravel forms the bed of the river today for a short distance.

A third outwash area, which is entirely buried, occurs near Arnprior (Fig. 1). West of this village the moraine is breached by large meltwater channels up to sixty feet deep. These combine to form a single channel, north-eastward from which the buried outwash deposit extends for about a mile. The outwash has a somewhat irregular surface, but on the whole falls gently north-eastward down to an altitude of 37 to 38 feet. Its surface then falls away sharply. Leading out from the meltwater breach in the moraine and extending the length of the buried outwash plain is a channel up to 200 yards broad and 10 feet deep. In this distance of approximately a mile, the floor of the channel declines from 38 to 29 feet. This large channel is interpreted as a meltwater feature owing to its size and its relation to the breach in the moraine. This suggestion is strengthened by comparison with the present small surface stream, which occupies a depression in the soft carse clay not much greater than its own width despite the five and half thousand years it has existed. The sharp change in gradient of the Arnprior outwash at 37 to 38 feet may represent the convex break of slope at the margin of a delta and suggests that the outwash was related to a sea-level at about this altitude. The full extent of the Arnprior outwash gravel is not known owing to the difficulty of boring through the buried beach deposits. In one borehole, however, gravel was encountered at 26.0 feet O.D. below the deposits of the High Beach (section CD, Fig. 2), and it is therefore possible that there exists an area of outwash gravel at a lower altitude than that already described.

Together the Arnprior and Lake of Menteith outwash deposits thus indicate a sea-level in the range 33 to 37–38 feet O.D., with the possibility of sea-level having been even lower during part of the time they were being deposited. This sea-level must have existed at a time when glacier ice extended up to the moraine. Thus it is clear that the Lake of Menteith outwash could only have been laid down at a time when the site of the present lake was occupied by glacier ice. Similarly, the meltwater channels feeding the Arnprior outwash are so located that, once the ice had wasted back from the moraine and begun to bare the low ground within it, they could not have functioned.

Relationships and Age of the High Buried Beach

A significant aspect of the High Buried Beach is that it has not been found inside the Menteith moraine. This implies that it was formed before the ice wasted back sufficiently far from the moraine for the sea to penetrate the basin enclosed by the moraine. On the other hand, in a number of boreholes the deposits of this beach overlie the outwash gravels near Arnprior. This means that, in part at least, the High Beach was formed after the outwash deposits were laid down. Thus the time at which the beach ceased to be formed is strictly limited, and is restricted to the period after the moraine had already been built up (for the channel that fed the Arnprior outwash cuts deeply into the moraine) but before the ice had retreated from the moraine. If we accept Donner's pollen dating of the moraine as a product of the Zone III readvance, then it would appear (in terms of available radiocarbon dates elsewhere) that the High Buried Beach was probably formed shortly after 10,300 years ago.

Since the outwash deposits near Lake of Menteith and Arnprior suggest a sea-level of no more than 33 to 38 feet when they were being accumulated, whereas the High Beach has a shoreline usually at 39 to 40 feet, it appears that, while the ice-margin stood at the Menteith moraine, there was a slight relative rise of sea-level. Since no evidence of buried beaches above the Main Beach has been found inside the moraine, it follows that relative sea-level subsequently fell and had already dropped below the level of the Main Beach before the ice had withdrawn from the moraine sufficiently to allow the sea to penetrate the basin within the moraine. Thus a marine transgression to 39–40 feet followed by a regression below 35 feet seems to have occurred while the ice-margin was standing at the moraine.

The transgressive nature of the sea that deposited the High Buried Beach is also indicated by the sharp upper limit to the deposits of this beach. Had the beach instead been produced during a regression from an altitude above 39 to 40 feet, one would have expected marine deposits to veneer the slopes above this altitude. A similar relationship is evident in the Arnprior outwash area, for where the surface of this outwash exceeds 41 feet it was in no instance found to be overlain by the deposits of the High Beach. The comparable altitude on the Lake of Menteith outwash is about 42 feet.

The extensive development of the High Beach eastward from Arnprior as compared with its feeble development elsewhere east of the moraine (see below) suggests local circumstances favourable to its accumulation, in turn suggesting that it owes its extent to the meltwaters that flowed into the area from the west. This interpretation is supported by three types of evidence. First, as already noted, the High Beach is composed of coarser material than the other two buried beaches. Secondly, the pinkish or brown colour of the deposits indicates their derivation from the Old Red Sandstone bedrock or drift that forms the rising ground to the south, or from the Old Red drift that is abundant in the west–east part of the moraine. Since the meltwater streams cut through the latter it seems likely that they supplied most of the material for the beach, while a smaller proportion may have been provided by streams draining from the high ground to the south. The third type of evidence relates to the channel that extends the length of the Arnprior outwash plain. The grey silty fine sand of the Main Buried Beach extends along the floor of this channel for almost a mile, to the vicinity of the right-angle bend in the channel (Fig. 1). These grey deposits appear to be identical with those that form this beach elsewhere, and their surface shows a very gentle increase in altitude up the channel, reaching 37.1 feet near the right-angle bend. It is clear, therefore, that this channel was a long, narrow arm of the sea in which quiet sedimentation occurred at the time that the Main Buried Beach was being formed. Since, however, the shoreline of the Main Beach lies several feet below that of the High Beach this immediately leads one to inquire why the channel had not been previously filled by the deposits of the High Beach. The simplest answer to this question is that, at the time the High Beach was being deposited, the volume of meltwater flowing through the channel was sufficient to prevent deposition of the finer sediments in it, except towards the margins of the channel where the pinkish deposits sometimes occur beneath the grey ones (thus, incidentally, demonstrating their relative age).

It thus appears that the High Buried Beach owes its strong development eastward from Arnprior essentially to the material provided by meltwaters from the west. It also seems, as has been noted already, that the beach ceased to be formed at a slightly later date than the deposition of the main mass of outwash. Away from the Arnprior area the High Beach appears to be only feebly developed. It may have been encountered in a few boreholes put down by D. E. Smith (1965) along the northern margin of the Carse of Stirling. East of the Lake of

Menteith its deposits, here usually grey in colour, form a thin layer overlying the buried outwash plain up to an altitude of about 42 feet O.D. (Fig. 2, top section).

Relationships and Age of the Main and Low Buried Beaches

The Main Beach was formed at a distinctly later date than the High Beach, for it is widespread both inside and outside the Menteith moraine. Within the moraine the fine grey sediments of which this beach is composed have a sharp upper limit, so that, as with the High Beach, it seems probable that the Main Beach is the result of a marine transgression. S. E. Durno (1956) analysed the pollen content of peat resting on this beach and showed that the lowest metre or so belongs to Zone V of the pollen sequence. From this one can infer that the age of the Main Beach is probably around 9500 years (Sissons and Smith, 1965). This date is now supported by Newey's pollen analyses of the peat at another point, which show that the base of the peat belongs to the very end of Zone IV or to the transition from Zone IV to Zone V (p. 58).

Although the steep bluff that in places separates the Main and Low Beaches probably represents erosion by the ancient Forth, the Low Beach cannot be regarded simply as an erosional step cut into the Main Beach. That its upper part at least is a true depositional feature related to a distinctly lower sea-level is shown by the presence here of the grey clay with contained aquatic plant stems in position of growth. Newey's pollen analyses demonstrate that the top part of this clay and the lower part of the overlying peat belong to Zone V, the remainder of the peat belonging to Zone VI. The Zone V/VI transition has been elsewhere radiocarbon-dated by Godwin and Willis (1965) as 8816 ± 192, 9009 ± 194 and 9090 ± 150 years B.P. This suggests that the Low Buried Beach was formed in or slightly before the period 8800 to 9100 years ago. However, a sample three-quarters of an inch thick from the base of the peat (23.6 feet O.D.) in the section analysed by Newey gives a radiocarbon age of 8690 ± 140 years B.P. (I–1839). Although the difference is slight (and the statistical errors in part overlap), it seems that the base of the peat is slightly younger than one would expect from the pollen analyses. This difference might be due to the pollen zones in the Forth valley not being exactly synchronous with those in northern England and north-eastern Ireland dated by Godwin and Willis. It is also possible that the trees that are represented in the Forth valley peat by abundant trunks and branches, and which flourished after the initial swampy phase, sent roots down into the basal peat and thus slightly reduced its apparent age as given by radiocarbon assay. Taking this latter possibility into consideration, and bearing in mind also that even the basal peat must be slightly younger than the deposit on which it rests, it seems reasonable to regard the Low Buried Beach as about 8800 years old on present evidence.

Thus it appears that both the Main and Low Beaches were formed within about fifteen hundred years of the time at which the margin of a piedmont glacier stood at the Menteith moraine. This suggests the possibility that the character of these beaches is related to the wasting away of this ice mass or perhaps to a later minor readvance in the mountains, for their deposits contrast markedly with the overlying carse clay. The lack of visible plant remains in the fine grey deposits (except in their upper parts in association with the overlying peat), their often distinct banding, their apparent lack of marine macro-organisms, and their close similarity to some raised beaches that are certainly composed of the finer debris provided by glacial meltwater streams, all make this suggestion feasible.

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Late-glacial and Post-glacial Shorelines in South-East Scotland

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THE WRITERS have been studying the Late-glacial and Post-glacial shorelines of south-east Scotland for several years. Some of the results of these studies have been published in various papers dealing with particular shorelines or aspects of the area, but no attempt has so far been made to present a statement of the pattern of sea-level changes in the area as a whole. It is felt that such a statement is now required, and the purpose of this paper is to fulfil this requirement as briefly as possible.

It is not intended to discuss here the field evidence in detail, since this is far too abundant to be considered in a single paper. The nature of the evidence will be apparent, however, from the two papers presented by the authors elsewhere in this volume and from the references to their work at the end of this volume. The main methods of study of the area (which comprises the Forth and Tay lowlands together with the intervening coast of East Fife) and progress to date in their application are as follows.

(i) Mapping on a scale of 1 : 10,560 of all identifiable estuarine and marine landforms and related fluvioglacial landforms. This is completed for the area specified.

(ii) Accurate instrumental levelling of raised shorelines and related outwash deposits and kame terraces at intervals of fifty to eighty yards where practicable. The summits of some kames and the rims of some kettle holes have also been levelled. The measurement of all the principal features and of most of the minor ones is complete, over 10,000 heights having been obtained.

(iii) Study of commercial borehole records, of which about 2000 have been collected so far. The distribution of these boreholes is very variable, some parts of the area having none at all, but one part (in and around Grangemouth) having a very large number. Borehole records continue to be collected as and when they become available and the authors are grateful for the co-operation of the Geological Survey and of many firms and organizations.

(iv) Hand borings to maximum depths of about thirty feet with a strengthened Hiller borer. So far nearly 700 boreholes have been made, some at critical localities scattered through

the area, but most of them in the western part of the Forth valley. It is planned to put down many more boreholes, and a detailed boring programme linked with laboratory analyses of the sediments has been recently begun by D. E. Kemp.

(v) Pollen analyses of certain deposits are being carried out by W. W. Newey, some of the results being described elsewhere in this volume.

(vi) The relation of the Post-glacial raised beaches to archaeological evidence is being investigated by I. A. Morrison. This work is not yet complete and will not be described here.

(vii) Radiocarbon dating is being carried out by the National Physical Laboratory and by Isotopes Incorporated.

As the above outline indicates, work on sea-level changes in south-east Scotland is still in progress: hence, this paper is a statement of the pattern of changes in the light of evidence so far obtained. It will be noted that no mention has been made of studies of marine organisms. Such studies are needed and it is hoped that specialists in macro- and micro-organisms will turn their attention to the area.

The relations of the various shorelines to one another are best shown by a height-distance diagram. Such a diagram appears as Figure 1 and shows the principal shorelines and some of the minor ones. The diagram has been constructed by projecting shoreline heights into a plane aligned N. 72° W.–S. 72° E. This line was selected as being the one most nearly at right-angles to the isobases so far as they are known at present. It will be seen from the place-names given in Figure 1 that, apart from the East Fife coast between St. Andrews and Fife Ness, the diagram depicts only the Forth valley. This is because the Forth valley extends much farther west than the Tay valley, so that a more widespread and complete sequence of changes can be elucidated here.

It must be pointed out that the term 'raised beach' is used throughout this paper even where the term 'raised mud flat' or 'raised estuarine deposit' would be more appropriate. This usage has been common in Scotland and is adopted here partly for brevity of expression and partly because in specific instances it is often difficult to say precisely where one type of feature ends and the other begins.

Late-glacial Shorelines formed before the Perth Readvance

The earliest Late-glacial shorelines represented in Figure 1 are those of East Fife, for this area, along with the eastern part of East Lothian, was the first to be freed of glacier ice during the period of glacier decay that followed the Aberdeen-Lammermuir Readvance (Sissons, 1965). The six principal shorelines of East Fife are discussed in detail on pages 31–51, so that only brief mention of them is needed here. The relations of the shorelines to glacial outwash and to one another show that, during the period of their formation, the ice-margin was, in general, retreating, thus allowing the sea to extend westward into the Forth area and north-westward along the north coast of Fife towards the site of St. Andrews. The shoreline gradients in East Fife range from 6.7 feet per mile for the oldest to between 3.1 and 3.2 feet per mile for the youngest. It has not yet been found possible to correlate the six Late-glacial shorelines of East Fife with shorelines in East Lothian. This probably results partly from the nature of the local relief in East Lothian and partly from spreads of blown sand in places. The steepest shoreline found in East Lothian has a gradient of slightly over five feet per mile.

Following the formation of the easternmost group of features, the ice continued to decay

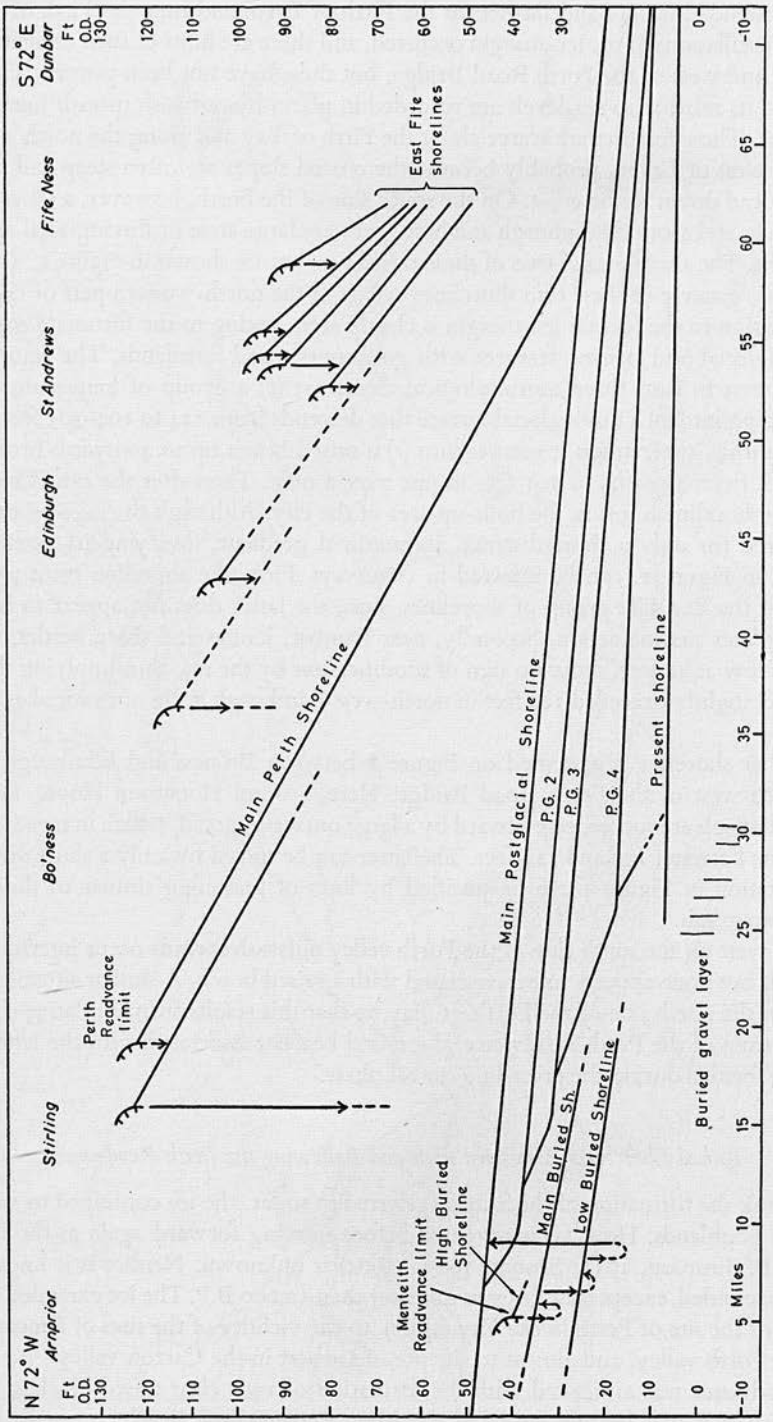


FIGURE 1.—The principal raised shorelines of south-east Scotland projected into a plane aligned approximately at right-angles to the isobases (so far as these are known at present).

and the sea extended farther and farther up the Firth of Forth and into the Firth of Tay. It is possible that oscillations of the ice-margin occurred, and there are hints of such oscillations near St. Andrews and west of the Forth Road Bridge, but these have not been proved. The retreat of the ice and its relation to sea-levels are recorded in places by outwash spreads merging into raised beaches. These features are scarce along the Firth of Tay and along the north side of the Forth valley west of Leven, probably because the coastal slopes are often steep and few large river valleys lead down to the coast. On the south side of the Forth, however, a series of rivers enters the sea in and around Edinburgh and here there are large areas of fluvioglacial and raised beach deposits. The shorelines of two of these raised beaches are shown in Figure 1.

The more easterly of these two shorelines occurs in the north-western part of Edinburgh, where its relation to the former ice-margin is clearly seen, owing to the fortunate coincidence of the fluvioglacial and marine features with golf courses and farmlands. The features comprise, from west to east, three morphological elements: (*a*) a group of kames and dead-ice hollows, leading into (*b*) a fluvioglacial terrace that descends from 124 to 104–105 feet in three-quarters of a mile, whereupon it merges into (*c*) a raised beach up to 400 yards broad whose shoreline falls from 104–105 to 101 feet in just over a mile. Thereafter the raised beach continues but its shoreline is lost in the built-up area of the city. Although the raised shoreline has been measured for only a short distance, its minimal gradient, justifying its extension as a broken line in Figure 1, can be inferred in two ways. First, the shoreline must pass below the lowest of the East Fife group of shorelines, since the latter does not appear to have been modified by later marine action. Secondly, near Dunbar, kames and sharp kettles, the latter with rims as low as 44 feet, show no sign of modification by the sea, thus implying that at the time sea-level slightly exceeded 100 feet in north-west Edinburgh it did not exceed 44 feet near Dunbar.

The other shoreline represented on Figure 1 between Bo'ness and Edinburgh occurs a short distance west of the Forth Road Bridge. Here, around Hopetoun House, kames and meltwater channels are succeeded eastward by a large outwash spread, which in turn is followed by a shoreline between 110 and 114 feet. The latter can be traced for only a short distance but its representation in Figure 1 can be justified by lines of reasoning similar to those of the preceding paragraph.

Farther west on the south side of the Forth valley outwash spreads occur intermittently as far as Falkirk but none appears to be associated with a raised beach. A similar situation appears to prevail on the north side of the Forth. It may be that this results from a relative rise of sea-level at the time of the Perth Readvance, the raised beaches associated with the latter having buried those formed during the preceding retreat phase.

Raised Shorelines Associated with and Following the Perth Readvance

Following the formation of the features referred to so far, the ice continued to waste back towards the Highlands. How far it retreated before moving forward again as the Perth Readvance (J. B. Simpson, 1933; Sissons, 1963b, 1964) is unknown. Neither is it known when the Readvance ended, except that this was not later than 12,000 B.P. The ice extended to within a few miles of the site of Perth in the Tay region, to the vicinity of the sites of Kincardine and Plean in the Forth valley, and almost to the site of Larbert in the Carron valley.

The readvance was associated with the formation of very clear raised beaches, of which the highest, which will be referred to as the Main Perth Raised Beach, is usually the most

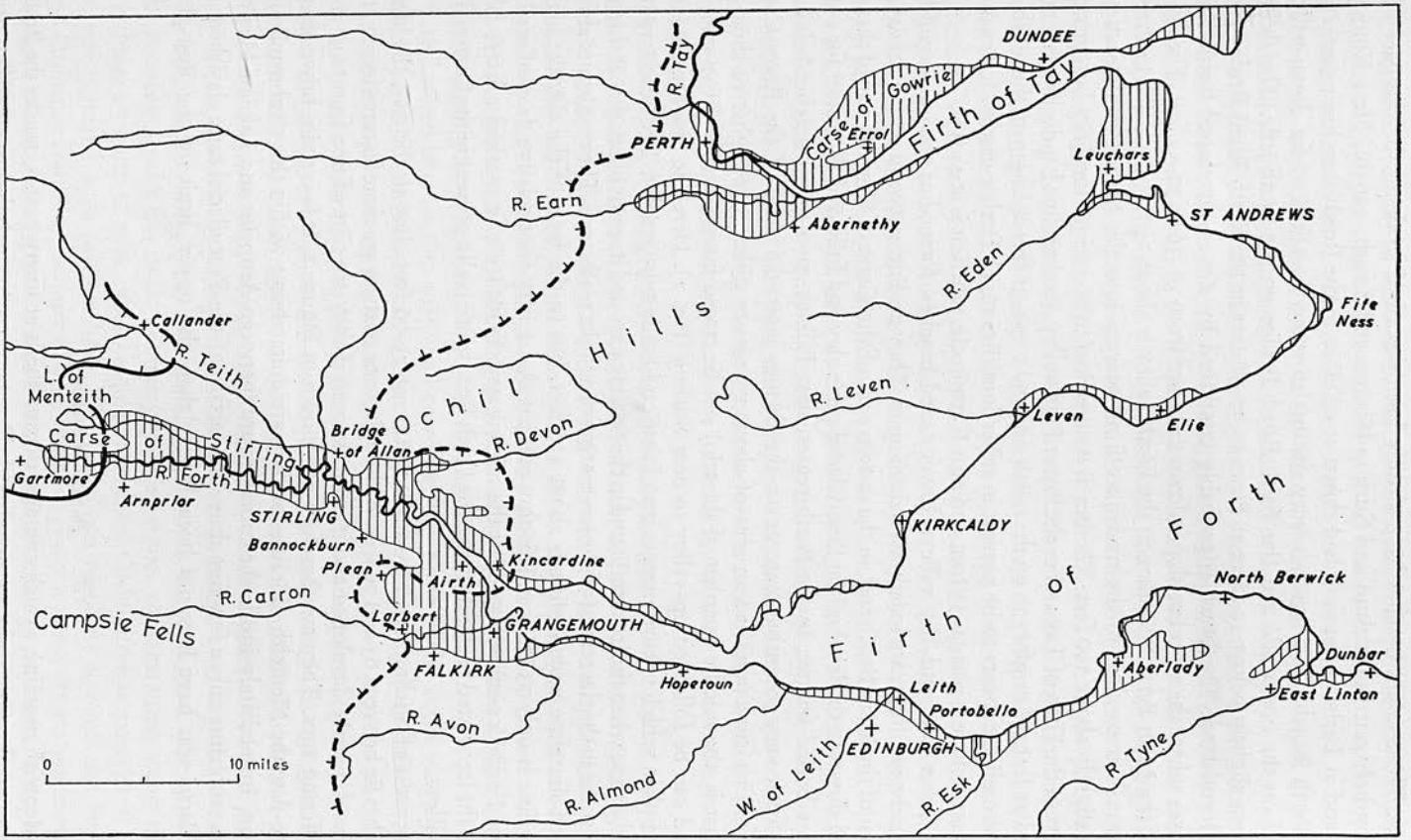


FIGURE 2—Extent of raised beach deposits (small areas omitted) and principal place-names referred to in the text. The limits of the Perth Readvance (dashed line) and the Menteith (Loch Lomond) Readvance are also shown.

conspicuous. These raised beaches have already been described in detail for the part of the Forth valley between Burntisland and Stirling (Sissons and Smith, 1965b). Near Kincardine and Plean and at Larbert, outwash laid down at or close to the Readvance limit merges into the Main Perth Beach. This beach is very distinct in many localities as far down-valley as Burntisland on the north side and the Forth Road Bridge on the south side. The shoreline declines from slightly below 125 feet at Plean to 77–78 feet at the Forth Road Bridge and 67 feet near Burntisland. The beach is probably continued by a well-developed feature in the Aberlady area where the shoreline slopes down eastward from 38–39 to 30–31 feet. The average gradient of the Main Perth Shoreline in the Forth valley is about 2¼ feet per mile. In the Earn valley, outwash at or close to the readvance limit merges into the Main Perth Beach at an altitude of slightly above 100 feet. The beach is conspicuous at various localities bordering the western part of the Firth of Tay and in the lower Earn valley. Its shoreline has declined to 77 feet at Errol and 71 feet at Longforgan on the north side and to 70–71 feet at Logie on the south side. The same shoreline appears to be present in other localities much farther east as, for example, near Leuchars, where it is at 51–53 feet, and near Carnoustie, where it is at 43 feet.

In the lower Earn and Tay valleys, lower raised beaches, formed as the ice wasted back from the readvance limit, are locally well developed. They decline eastward with a similar gradient to that of the Main Beach and, in the eastern part of the Carse of Gowrie, one of them has been traced down to the level of the Post-Glacial carse clay and followed thereafter for a short distance as a buried feature. In the Forth region near Falkirk, two raised beaches below the main one form very extensive features on the former intertidal delta of the River Carron which, before its dissection, had an area of about six square miles. The higher of these two beaches (that is, the middle member of the trio) can be traced intermittently down-valley to Bo'ness and can be followed up-valley to near Stirling (Fig. 1). Here the beach merges into an outwash plain which in turn merges into kame and kettle topography. The beach was thus formed at the time when the ice terminated in the Stirling gap, and it seems likely that at this stage there was a pause in the the retreat of the ice-margin or a slight readvance. The maximum altitude of the raised shoreline near Stirling is about 125 feet, but to the west of the city the highest raised shoreline is only 73–76 feet. It is thus evident that a fifty-foot relative drop of sea-level occurred while the ice-margin stood at the Stirling gap. Evidence of a marked relative fall of sea-level as the ice wasted back following the Perth Readvance is also present in the Carron, Earn and Tay valleys.

To the west of Stirling, apart from the feature at 73–76 feet close to the city, the highest raised beaches lie between 65 and 70 feet. Although some of these are clear features locally, they are, in general, poorly developed and it is not known if they are part of one feature or are of slightly differing ages. They are therefore not shown in Figure 1. It is certain, however, that they all pre-date the Menteith moraine for they are quite absent within this twelve mile-long arcuate loop. Immediately inside the morainic arc (between Arnprior and Lake of Menteith) kettle holes with rims only a few feet above the surface of the Post-glacial carse clay show that the Late-Glacial seas have here not been higher than the upper limit of the Post-glacial seas.

Buried Shorelines

The Menteith moraine, which contains marine shells at many points, marks the limit of an important readvance (Simpson, 1933). J. J. Donner (1957) concluded from pollen studies

that the readvance occurred in Zone III of the pollen sequence, implying that the moraine was formed about 10,300 years ago. Although there is no reason at present to doubt Donner's pollen dating, it should be noted that it is not conclusively proven.

Relative sea-level changes as recorded by buried shorelines and outwash deposits in the neighbourhood of the Menteith moraine are described in detail on pages 24 to 25 and need be mentioned only briefly here in relation to Figure 1. In this diagram the three buried features that have been identified beneath carse clay and peat are shown. The High Buried Beach was formed while the ice extended to the Menteith moraine and does not occur inside the moraine. As explained on pages 25 to 27, the evidence indicates that, while the ice-margin stood at the moraine, sea-level rose to about 39–40 feet (the altitude of the shoreline near the moraine) and then fell below 35 feet. The shoreline gradient shown on Figure 1 is not proved: the shoreline is depicted as sloping down towards the Main Buried Shoreline since it has not been possible to follow it more than a few miles down the Forth valley from the Menteith moraine.

The Main Buried Beach, formed about 9500 years ago (page 27) and associated with a transgression, has been traced by boring from the area inside the moraine to within a few miles of Stirling and again between Stirling and Bannockburn. Throughout this distance, its shoreline shows a decline in altitude towards the east, falling from about 37 feet in the westernmost locality where it has so far been identified, to about 24 feet near Bannockburn. This gives an average gradient of 0.8 feet per mile, as compared with a gradient of 1.3 feet per mile for the part of the shoreline in the vicinity of the Menteith moraine (page 23). The shoreline is shown with a uniform gradient (0.8 feet per mile) in Figure 1, since it is not yet known if the steeper gradient towards the west is of general significance or is merely a local anomaly (for instance, because of the influence on estuarine sedimentation of the marked constriction caused by the Menteith moraine).

The Low Buried Beach, formed around 8800 years ago, has been traced from the area inside the Menteith moraine to a point a few miles short of Falkirk. Between Stirling and Airth it is a very extensive feature, often a mile broad. Its shoreline has been approximately located at a considerable number of points. Near Bannockburn, for example, it is at about 18 feet, a figure that may be compared with about 27–28 feet for the probable shoreline altitude in the area so far studied within the Menteith moraine. The average gradient of the Low Buried Shoreline is slightly less than that of the Main one, as shown in Figure 1.

Buried raised beaches that may well correlate with those in the Forth valley have been identified by hand-borings in the lower Earn valley. The evidence is not yet sufficient, however, to permit correlation between the two areas.

At this point it is necessary to depart from the sequential treatment of the various shorelines to consider the stratigraphy of the Grangemouth area. In and around this town and extending up the Forth valley to Airth, many hundreds of commercial boreholes have been put down. These reveal a very consistent stratigraphic sequence, which is summarized in Figure 3. The most significant element of this sequence in the present context is a widespread layer of gravel that lies between sediments composed mainly of clay, silt and fine sand. The gravel layer is usually between 2 and 5 feet thick, has many rounded stones, contains many stones of boulder size, includes a variety of rock types ranging from local Carboniferous material to Highland schist, and is mixed with marine shells. The bed slopes regularly towards the Forth, from around Ordnance Datum at its inner margin where in places it meets the steeply rising ground at the back of the carselands, to about 20 feet below O.D. close to the Forth. The gravel layer is known to occur over an area of many square miles and its width at right-angles to the Forth

often exceeds a mile. These facts show that the gravel layer represents a buried beach lying at a distinctly lower altitude then the buried features considered so far.

Since the gravel layer is markedly different from the fine sediments above and below it (as well as from most of the other features referred to in this paper) some consideration of its origin is necessary. It is likely that the gravel was in part supplied by the Avon, Carron and smaller streams, for recent artificial excavations near the points where these streams enter the carse plain show spreads of coarse gravel fanning out from the mouths of the valleys. Some of this river gravel was certainly deposited as the carse clay accumulated, for the two deposits interdigitate, but it may well be that the lowest gravel passes into the widespread gravel layer of Figure 3 and contributed to its accumulation. Whether this is so or not, however, it is clear that an important factor in the production of the gravel bed of Figure 3 was erosion by the sea of drift previously deposited. This is shown in several ways. First, the ground at the back of the carse plain often rises steeply, sometimes for a hundred feet, and strongly resembles a marine

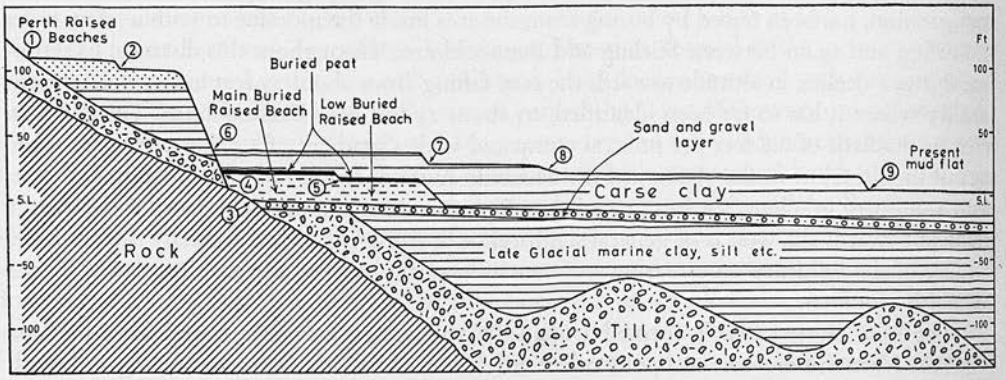


FIGURE 3—Diagrammatic section of the Grangemouth–Airth area to show the main elements of the stratigraphy in relation to the visible and buried morphological features. Circled numbers refer to shorelines in order of formation. The diagram is generalized to enhance clarity; thus, only three visible Post-glacial raised shorelines are shown.

cliff. That this steep slope is in some instances a result of marine erosion is suggested by its being best developed in places that would have been most exposed to former wave attack, whereas in re-entrants a gentle slope often backs the carse plain. Secondly, the till that mantles the solid rock almost everywhere in the Grangemouth area is often missing (beneath the carse clay) in a belt up to several hundred yards wide along the foot of the steep slope, thus suggesting that it has been removed by the sea. Thirdly, farther out from the old cliff the till is sometimes bevelled and on this planed surface rests the buried gravel layer (Fig. 3). In other instances, the higher parts of the strongly undulating ice-moulded buried surface of the till have been truncated and again the gravel layer rests directly on the truncated till. Fourthly, as shown in Figure 3, the marine deposits associated with the Perth Readvance have been considerably eroded to produce a steep bluff in which the horizontally bedded layers of Late-glacial clay, silt and sand crop out.

It thus appears that, while the materials of the gravel layer may have been in part supplied by streams, they were mainly provided by the erosion of till. Some material was probably derived also from the erosion of the Late-glacial marine sediments, for these contain ice-rafted stones. It is relevant to note that the present coast of the Forth eastward from the vicinity

of Bo'ness has suffered considerable marine erosion in relatively recent times where it is composed of drift: in places, erosion is in progress today. The result of this erosion is that the present shore is often littered with a residue of stones and boulders of all sizes mixed with sand, mud and shells, this beach deposit closely resembling the buried gravel layer of the Grangemouth area. The close similarity was clearly demonstrated by H. M. Cadell (1883) for he classified in terms of rock type several hundred stones on the shore near Bo'ness, and compared the results with a similar analysis of stones from the buried gravel layer where it was then artificially exposed in Bo'ness. He found that the stone counts were virtually identical, a result that strongly favours the view that the present shore deposits and the buried gravel layer have a common origin.

Commercial borehole data show that near Airth the buried gravel layer passes beneath the deposits of the buried raised beaches already referred to (Figure 3). Thus the layer is older than these beaches, but, from evidence described above, must be younger than the Perth raised beaches. Thus a period of low sea-level, hitherto unrecognized in Scotland, appears to have occurred after the Perth Readvance but not later than the Menteith Readvance, and the erosional evidence suggests this period of low sea-level was of considerable duration. The precise altitude of the shoreline is not yet established and boring is being carried out to determine this. Provisional evidence suggests that the shoreline is approximately at Ordnance Datum south-west of Grangemouth and between six and ten feet below O.D. in Bo'ness (where numerous closely spaced commercial boreholes have been put down). The feature is therefore represented by a band of shading in Figure 1.

Visible Post-glacial Raised Shorelines

Following the formation of the Low Buried Beach, which ended about 8800 years ago, sea-level fell to expose this beach. As pointed out on page 27, the minimal sea-level probably occurred about 8500 years ago. Thereafter sea-level rose relative to the land during the major Post-glacial transgression. The early part of this transgression appears to have been rapid. Whether the relative rise of sea-level was continuous, or was interrupted by temporary regressions, is not at present known. The result of the transgression was that peat accumulating on the now buried beaches was covered by carse clay over extensive areas in the Forth, Earn and Tay valleys. In a few places, however, peat growth managed to keep pace with the rising sea-level, so that occasionally the carse clay is replaced by peat for its full thickness. The two largest areas where this occurred lie near the Menteith moraine and comprise two roughly circular peat bogs, each about a mile across, that continued to accumulate as the carse clay was deposited around them (Sissons and Smith, 1965a).

According to H. Godwin and E. H. Willis (1962), the transgression culminated about five and a half thousand years ago, for wood from a few inches above the base of the peat that extensively overlies the carse clay in the western part of the Forth valley gave a radiocarbon date of 5492 ± 130 years B.P. The associated shoreline (the Main Post-glacial Shoreline) falls in altitude down the Forth valley from 49 feet in the extreme west to 19–20 feet at Dunbar (Fig. 1). Up-valley from Falkirk and in the more sheltered localities farther east, as near Aberlady and in the Tyne estuary, the deposits consist mainly of carse clay (elevated mud flats) and on more exposed coasts they comprise mainly sand and gravel. In the Tay area the carse clay of the Carse of Gowrie was almost entirely deposited at this time. The shoreline at the back of the carselands reaches 36 feet in the Earn valley, falls to 28–29 feet immediately west

of Dundee, and has declined to 26 feet in the Eden valley a few miles from St. Andrews. South-eastward from the Eden estuary, the beach is composed of shelly sand and gravel, and the shoreline has been traced intermittently to Fife Ness, where it is at 21–22 feet.

After the formation of the Main Post-Glacial Raised Beach, a series of lower beaches was produced. These are most clearly preserved in the carselands of the Forth valley upstream of Kincardine and Grangemouth, where three features below the main one have been identified. At its western end each shoreline increases slightly in gradient and merges into the one above it, as shown diagrammatically in Figure 1. This is the result of sedimentation during the periods of falling sea-level between the times when sea-level relative to the land was more or less stationary. It is not known if the second and third Post-Glacial shorelines were associated with a slight transgression, but the lowest feature appears to be transgressive since very thin peat is occasionally found beneath the associated carse clay.

For purposes of comparison, the present shoreline as levelled on mud flats is shown in Figure 1. This is at about 5 feet in the Eden estuary. In the Forth valley it declines from 6.6 feet immediately north-west of Grangemouth to about 4.5 feet at the mouth of the Tyne near Dunbar. This faint eastward slope is probably the result of tidal influences, for Admiralty Tide Tables (1964) show that high water mark is 10.3 feet at Alloa, 9.6 feet at Grangemouth and 8.3 feet at Dunbar. It will be noted that the feature measured on the modern mud flats, which is comparable with that measured on the elevated features, is 3 to 4 feet below high water mark, so that one may suggest that the elevated features may have been similarly related to former high water marks.

Conclusion

In the preceding account, the purpose has been to present the pattern of relative sea-level movements in Late-glacial and Post-glacial times in south-east Scotland so far as they are known at present. Discussion of the causes of these movements has intentionally been avoided, as also has discussion of the evidence in relation to sea-level changes in the rest of Scotland and adjacent parts of the British Isles. These subjects are far too large to be dealt with here and it is planned to consider them at length elsewhere.

NOTE

All altitudes are expressed in terms of Ordnance Datum and are related to the second geodetic levelling of the Ordnance Survey.

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Glacial stages and radiocarbon dates in Scotland

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SYNOPSIS

New radiocarbon dates and previously published ones related to Scottish glacial events are discussed. Donner's conclusion that the Loch Lomond and Menteith end-moraine complexes were produced at the limit of the Zone III Readvance is confirmed. It is suggested that the Perth Readvance maximum occurred between 13 500 and 13 000 years ago, that the Aberdeen-Lammermuir Readvance post-dates a long interstadial, that an ice-movement during which shelly drift was deposited near Berwick pre-dates it, and that during the interstadial the British Isles were largely or entirely free of ice.

NEW RADIOCARBON DATES

Four radiocarbon dates closely related to glacial events in Scotland have so far been published (Kirk and Godwin 1963, p. 247; Fitzpatrick 1965; Bishop 1963, p. 125; Rolfe 1966, p. 255). The purpose of this paper is to record four additional dates and to discuss them in relation to the previously published ones and to a map of glacial stages. The four dates are as follows.

1. Drymen. Shells and shell fragments, some water-worn, from a gravel band in a sand-pit at [NS/483880]. These deposits form part of the Loch Lomond end-moraine complex (Simpson 1933, pp. 641-4). 11 700±170 B.P. (ref. I-2235).

2. Menteith. Shells, some crushed, from contorted glacier-transported marine clay interbedded with fluvioglacial sands and gravels forming part of the Menteith end-moraine complex at [NN/589000]. 11 800±170 B.P. (ref. I-2234).

3. Kilmaurs. Part of a tusk of *Elephas primigenius* presented to the Hunterian Museum, Glasgow, in 1829 and obtained from a bed of clay below till and an arctic shell bed, 34 feet from the surface, at Woodhill Quarry (Gregory and Currie 1928, p. 2). 13 700(+1300, −1700) B.P. (ref. GX0634).

4. Berwick. Shell fragments from a gravel layer underlying 30 ft of till and overlying sands, which in turn rest on laminated clay. The whole forms a bluff 80 to 90 ft high on the north side of the Tweed at [NT/982532]. Greater than 41 100 B.P. (ref. I-2135).

These dates and others obtained in Scotland are best discussed in relation to a map of the limits of glacial readvances. Such a map is shown in Fig. 1 and is

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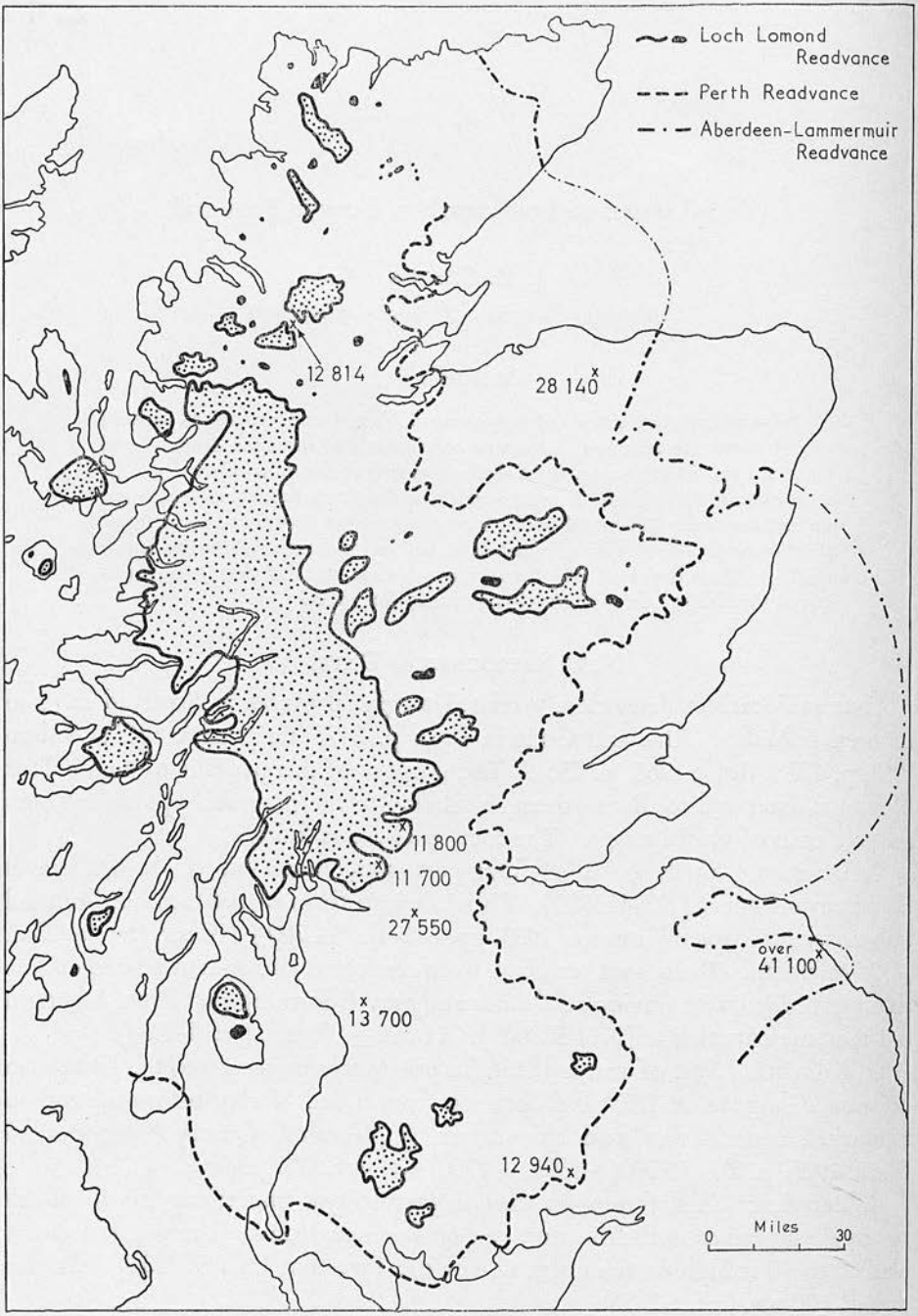


FIG. 1. Glacial readvance limits in relation to radiocarbon-dated sites in Scotland.

almost the same as one published and discussed recently (Sissons 1965, fig. 14.4 and pp. 474-83). It must be emphasised that this map, like others that have preceded it (e.g. Charlesworth 1955), can only be regarded as extremely tentative and will have to be repeatedly modified as knowledge accumulates. That the radiocarbon dates now available do not necessitate such modification reflects the inadequacy of the information available (not only of radiocarbon dates, but also of morphological, stratigraphical and palynological data) rather than the validity of the map.

Loch Lomond Readvance

The radiocarbon dates for shells from the Loch Lomond and Menteith end-moraine complexes place in the Alleröd the marine incursion that preceded the readvance whose limits the moraine complexes represent, implying that the readvance itself took place during Zone III (10 800 to 10 300 B.P.). This confirms Donner's (1957) interpretation based on pollen analyses.

Perth Readvance

A date of 27 550(+1370, −1680) B.P. has been obtained for a water-worn bone of *Rhinoceros antiquitatis* found near Bishopbriggs in deposits associated with the Perth Readvance (Rolfe 1966). This date is far too old for the interstadial immediately preceding this readvance and it appears that the bone has been derived from an earlier phase. The situation around Kilmaurs in lowland Ayrshire is very different, however, for here the abundance and variety of the organic remains (especially when considered in relation to the number of points at which the buried organic deposits have been located) and the stratigraphy make it clear that these remains lie in a distinct stratified deposit over which the ice advanced and laid down its till (Bryce 1865; Young and Craig 1871; Craig 1888; Richey *et al.* 1930, pp. 325-30; Sissons 1963, 1964).

The error given for the Kilmaurs date allows a rather long period of time—12 000 to 15 000 years ago—during part of which the *Rhinoceros antiquitatis* might have lived. However, since the latter existed before or at the time of the Perth Readvance, and the Kilmaurs site lies some distance within the readvance limit, it seems likely that the culmination of the readvance lies in the later rather than the earlier part of this 12 000 to 15 000 year time span. On the other hand, it can hardly lie in the period 13 000 to 12 000 B.P. for the following reasons.

1. Certain evidence on the Continent and in Britain indicates that the Alleröd period was preceded by a short cold period (Zone 1c) and this by the milder Bölling period (Zone 1b). The Zone 1c cold period was apparently of short duration (perhaps 300 to 500 years) and appears insufficient to allow the large accumulation of ice that was associated with the Perth Readvance. The time of the Bölling period is not firmly established but dates between about 12 400 and 12 000 B.P. have been obtained for it in Denmark and Poland (Tauber 1960,

pp. 14-15; 1962, pp. 28-9). In Britain it has been suggested that earlier dates relate to this period including one as early as 13 190±230 B.P. in South-east England (Kerney 1965). Thus, while there is uncertainty concerning the dates of the Bölling amelioration, it seems very improbable that the Perth Readvance limit was achieved between 12 500 and 12 000 B.P. and fairly improbable that it was between 13 000 and 12 500 B.P.

2. A radiocarbon date of 12 814±155 B.P. has been obtained for late glacial deposits at Loch Droma that have not been subsequently covered by glacier ice (Kirk and Godwin 1963, p. 247). The position of this site in relation to the Perth limit indicates that the Perth Readvance maximum is earlier than this date.*

3. Along the course of the Annan near Lockerbie, peat interbedded with silts, clays and sands has been shown to contain late-glacial pollen and has given a radiocarbon date of 12 940±250 B.P. (Bishop 1963, p. 125; Moar 1963). A peculiar feature of the sediments is that they are gently warped into broad synforms and antiforms. This peculiarity is discussed at some length by Bishop (1963, pp. 128-31), who tentatively concludes that the warping may be the result of a glacial readvance. It seems preferable, however, to interpret these deposits as having accumulated on top of buried glacier ice during the last deglaciation of the area, the warping having occurred as the ice melted out. On this interpretation the deposits imply that deglaciation was in progress here (and probably in the rest of Scotland) 12 940±250 years ago. In the light of the evidence considered above it seems probable that this deglaciation was that following the maximum of the Perth Readvance. Taking the evidence as a whole it seems likely that the maximal extent of this readvance was attained between 13 500 and 13 000 years ago.

Aberdeen-Lammermuir Readvance

At Teindland in Morayshire a fossil podsol, covered by 6 to 8 ft of material interpreted by Fitzpatrick (1965) as sandy till and outwash gravel, has given a radiocarbon age of 28 140 (+480, −450) B.P. Since this site lies within the limit of the Aberdeen-Lammermuir Readvance shown in Fig. 1, one may infer that this readvance occurred after this date. The date falls within the time of a long interstadial whose existence is now well established by radiocarbon dates in North America and Europe, the succeeding glaciation culminating between 17 000 and 20 000 years ago. The Aberdeen-Lammermuir maximum may therefore have occurred in the period 17 000 to 20 000 years ago or it may be younger than this.

Long Interstadial

While the Bishopbriggs bone (27 550 (+1370, −1680) B.P.) does not help to date the Perth Readvance, it is of considerable significance in relation to the long

* The ice limits shown in Fig. 5 (A and B) of Kirk and Godwin's paper appear to relate neither to the Loch Lomond nor to the Perth Readvance and revision of these maps is necessary.

interstadial that preceded the last great build-up of glacier ice in the British Isles. Bishopbriggs is situated no more than 50 miles from the heart area of the most important centre of ice accumulation in the British Isles, that of the south-west Grampians. Since there is little chance of the bone having been transported from a point more distant from the ice centre than that at which it was found (although it could well have been transported from a point nearer to the ice centre), it is clear that around 27 000 to 28 000 years ago the Bishopbriggs area was ice-free. This in turn must mean that the whole of the British Isles away from the highest mountain areas was also ice-free at this time and it may be that the mountain areas themselves were free of ice. The build-up of glacier ice that succeeded the long interstadial was associated with an ice limit at least 200 miles south of Bishopbriggs, for near Northwich in Cheshire glacier-transported shells derived from the Irish Sea have been dated at 28 000 (+1800, −1500) B.P. (Boulton and Worsley 1965).

Earlier Glacial Advance

Although the site of the shells near Berwick dated at more than 41 100 B.P. lies just within the limit of the Aberdeen-Lammermuir Readvance shown in Fig. 1, the date does not relate to this readvance. That the last ice to cover the area around Berwick came from the west and south-west is shown by the strongly ice-moulded relief of the low ground of the Tweed basin, which near Berwick is aligned almost due west-east. The dated shells could therefore have been picked up by this eastward-moving ice from an area west of that at which they were found, or they could have been transported by an earlier ice-mass moving into the Berwick area across the present coastline. The second alternative is indicated for the following reasons.

1. No marine shells have been found at any point in the Tweed basin west of the locality from which the dated shells were obtained (apart from shells transported by human agency).

2. The stratigraphy described on p. 375 suggests that the advancing ice held up a lake in which fine sediments were first deposited, followed by increasingly coarser sediments as the ice approached, ultimately to cover the site. This sequence of events is difficult to reconcile with an eastward movement of ice.

3. Gunn (1897) recorded that the shelly deposits contain fragments of Belemnites. There is no place inland of Berwick from which the latter could have been derived.

The Belemnite fragments are interesting in that they suggest the association of the shelly deposits at Berwick with those in the east of Scotland, for the shelly drifts of eastern Scotland from the Orkneys southwards are always associated with Mesozoic erratics derived largely or entirely from the present sea-bed. At least in the area from eastern Aberdeenshire to Berwick the shelly drifts with Mesozoic erratics imply very different directions of ice movement from those associated

with the Aberdeen-Lammermuir Readvance. They also imply a more extensive ice-cover than at that stage and in some areas suggest that the movement of the Scottish ice was modified by the Scandinavian ice-sheet. While the evidence is not conclusive, it seems likely that these events relate to some time preceding the long interstadial referred to above, a view that is encouraged by the minimal age of 41 100 years obtained for the Berwick shells.

Conclusions

The sequence of more recent glacial events in Scotland so far as this can be elucidated in relation to the few radiocarbon dates at present available, appears to be as follows.

1. A long interstadial, during which the country was largely or entirely ice-free, drawing to a close perhaps around 25 000 years ago.

2. An advance of ice culminating about 17 000 to 20 000 years ago, the limit being represented either by the Aberdeen-Lammermuir line of Fig. 1 or by a more extensive ice-cover whose limits are not known.

3. Deglaciation, followed by the Perth Readvance, which probably culminated between 13 500 and 13 000 years ago.

4. Deglaciation, followed ultimately by the Loch Lomond Readvance between 10 800 and 10 300 years ago.

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Drift Stratigraphy and Buried Morphological Features in the Grangemouth–Falkirk–Airth Area, Central Scotland

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ABSTRACT—Mainly on the evidence from about 1900 boreholes, the Quaternary deposits and buried morphological features of a small area in the mid-Forth valley are described and discussed. The buried bedrock surface is contoured and a deep trench that crosses the area is attributed to meltwater and glacial erosion. Thick late-glacial sediments that begin the marine succession extend up to the Main Perth Raised Shoreline at 29–36 m. Following a major relative fall of sea level, extensive marine erosion occurred in late-glacial times to produce a planation surface, veneered by gravel and often backed by a cliff. During the production of this surface the sea probably rose relative to the land, subsequently standing at or slightly above its present level for a considerable time to produce the Buried Gravel Shoreline. Erosion continued in places as sea level first rose to about 6 m and then remained relatively stable for a time. Thereafter the Main and Low Buried beaches were formed, with maximum recorded shoreline altitudes in the present area of 7.4 and 5.1 m, and terminal dates of formation of about 9500 and 8800 BP respectively. Regression of the sea until about 8500 BP was followed by a transgression during which the carse deposits began to accumulate. Two systems of buried valleys related to the two periods of relatively low sea level are identified.

IN and around Grangemouth, Falkirk and Airth, on the south side of the Forth, the extensive and often thick Quaternary deposits have been pierced by many hundreds of commercial boreholes, supplemented by several hundred hand boreholes. The purpose of this paper is to describe and, so far as possible, explain the sequence of deposits as revealed by these boreholes, dealing especially with buried morphological features including the bedrock surface, the widespread and significant buried gravel layer, buried raised beaches and buried valleys.

Published material on the deposits of the area consists mainly of brief descriptions and explanations in the older literature. Exceptional are the important contributions made by H. M. Cadell (1883, 1913), who identified some of the main stratigraphical units and put forward explanations, some of which are still acceptable. A recent paper by J. B. Sissons, D. E. Smith and R. A. Cullingford (1966) includes a summary diagram of the main elements of the stratigraphy and buried morphology that is very relevant to the present contribution: it is recommended that the reader study that paper before reading the present paper.

Surface Form and Drift Succession

Over much the greater part of the area considered, the ground surface is composed of raised estuarine or marine deposits. An extensive area bordering the Forth is covered by raised post-glacial mudflats (carse), whose natural surface for the most part slopes gently towards the Forth with a gradient between 1 in 1000 and 1 in 2500. The carse surface is mostly below an altitude of 8 m except in the north-west, where a considerable area, forming part of the Main

Post-glacial Raised Beach,[1] slopes gently from 12.0–12.8 m at the former shoreline to 11.0–11.5 m at its eastern margin. In places this gentle slope is concealed beneath the remains of a formerly extensive blanket of peat. Included with the carse in Figure 1 are areas of reclaimed land such as that to the east of Airth, the Grangemouth docks peninsula, and the low ground bordering the coast that extends into Bo'ness. Most of the carse deposits are between 5 and 8 m thick, but attain 10–12 m over part of the area covered by the Main Post-glacial Raised Beach. In the latter area the carse deposits usually consist almost entirely of a mixture of silt and clay throughout their full depth, but in other parts such material alternates with or is mixed with layers of sand, fine gravel and very abundant shells. This paper is not specifically concerned with the carse deposits, but with the materials that lie beneath them; the intercalated sands, gravels and shells will be included in the general term 'carse deposits'.

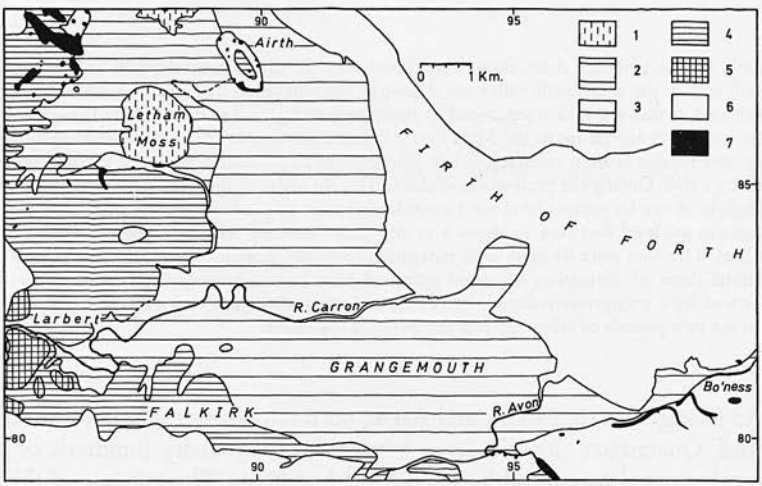


FIGURE 1—Ground surface material. 1. Peat; 2. Mainly carse deposits, but also including made ground and very small areas of fluvial deposits; 3. Carse deposits in the principal area of occurrence of Main Post-glacial Raised Beach; 4. Late-glacial marine sand and clay; 5. Outwash gravel associated with Perth Readvance; 6. Other fluvioglacial deposits and till; 7. Bedrock

In the western part of the area, the carse deposits often rest on peat and this in turn frequently overlies the grey micaceous silts and sands that constitute the buried raised beaches previously described for an area 15–24 km west of Stirling (Sissons, 1966). Elsewhere, the carse deposits usually rest on a layer of gravel or on the planated surface of till or bedrock associated with the gravel layer. Beneath the gravel layer and the buried beaches, the most common material is laminated late-glacial clay.

As well as occurring extensively beneath the carse deposits, the late-glacial clay appears at the surface landward of the margin of the carse, where it occurs up to an altitude of over 30 m. In these higher areas the clay is usually covered by sandy clay and this in turn by sand. Around Larbert, the sand is overlain by outwash gravel interpreted as laid down at or near the limit of the Perth Readvance. The surface of the sand often slopes only very gently and here constitutes the three Perth raised beaches recognized in this area as occurring between about 24 and 37 m (Sissons and Smith, 1965). These raised beaches were originally much more extensive but have been greatly dissected by the River Carron and lesser streams as well as by numerous gullies

that are now streamless or contain underfit streams. The Perth raised beaches have also been much reduced in area (and in some places probably completely destroyed) by later marine erosion at a much lower altitude. Consequently, in localities most exposed to former marine attack, the ground rises steeply along the landward margin of the carse, this steep slope being developed in bedrock or till, capped in some areas by the late-glacial marine deposits.

Beneath the late-glacial marine deposits, fluvioglacial sand and gravel occurs in places. Especially eastward from Falkirk, this material rises to the ground surface from below the late-glacial sediments and forms the Polmont esker system and associated features. The next lower element in the drift succession—the glacial till—occurs widely (though not universally) below the marine sequence (or, locally, beneath fluvioglacial deposits). In places the till alternates with

TABLE I

Drift Succession for Three Types of Area (see text)

	A	*B*	*C*
Peat		(×)	(×)
Carse deposits		×	×
Peat		×	
Micaceous silty sand (buried beaches)		×	
Gravel layer		(×)	×
Gravel (outwash)	(×)		
Sand (late-glacial marine)	×		
Sand and clay (late-glacial marine)	×		
Laminated clay (late-glacial marine)	×	×	×
Sand and gravel (fluvioglacial)	(×)	(×)	(×)
Till	×	×	×
Sand (fluvioglacial)			(×)
Bedrock	×	×	×

or overlies fluvioglacial sand and gravel and, in the deep trench that underlies the River Carron, it rests on a great thickness of sand that in turn lies directly on bedrock.

Owing to various causes, including marine erosion of previously deposited marine sediments and the fact that the various marine sediments have different upper altitudinal limits related to sea level at the times they accumulated, the full drift succession does not occur at any point. It would therefore be misleading to present the drift succession given in the first column of Table I without the qualifications given in the remaining three columns. In the column headed *A* the normal succession in areas where the Perth raised beaches occur is indicated, in column *B* the normal succession in the part of the carse beneath which the buried raised beaches occur is given, and column *C* is typical of carse areas beneath which the buried beaches are absent. In each of columns *A*, *B* and *C* a cross signifies that a particular deposit is normally present, while a cross in parentheses signifies that a particular deposit is present in a small but significant part of the area to which the column applies.

Borehole Data

Commercial Boreholes

About 1300 commercial borehole records, from various sources, have been obtained for the area. These records vary greatly in amount of detail, some of the older ones giving only the

total thickness of superficial deposits, whereas the many recent foundation bores are usually recorded in considerable detail, the descriptions of the strata being based on laboratory analyses the full results of which are often presented in lengthy reports. As is evident from the accompanying maps, the distribution of commercial boreholes is irregular. On some building sites they are so closely spaced that it was possible to plot only selected typical ones on the maps. Much of the petrochemical area in the eastern part of Grangemouth is particularly well covered by very detailed foundation bores, while the ground south and south-west of Airth is covered by a close network of mineral bores for some of which drift records are fairly detailed. The main area for which information is inadequate includes the Firth of Forth and the artificial peninsula of Grangemouth docks, where rockhead descends to great depths and some of the deposits differ in certain respects from those beneath the adjacent land and cannot readily be correlated with the latter on the basis of present evidence.

In plotting and interpreting the borehole data, it became apparent that a very few records were erroneous, the most probable cause being incorrect location of the bore site. In some instances drift records were clearly highly generalized and could not be used. In a number of bore logs the buried gravel layer, which is sometimes mixed with clay, is referred to as 'boulder clay': as a literal description this may be correct, but since 'boulder clay' is used in bore records in a genetic sense this represents misinterpretation (which is usually readily apparent). In some bores a stratum is described as 'clay with stones', a description that could refer to till or to late-glacial marine clay containing ice-rafted stones. Where adjacent more detailed records were available this uncertainty could often be resolved. On the other hand, it was assumed that the description 'hard clay with stones' referred to till, for this description is not applicable to the late-glacial marine sediments. No significance was attached to the fact that a particular thin stratum was not recorded: hand boring at points very close to the sites of commercial bores not infrequently showed that the presence of a thin stratum (such as buried peat, or the buried gravel layer where poorly developed) had been overlooked in the commercial investigations.

It must be emphasized that these and other limitations apply to only a small proportion of the commercial bore records, for comparison of records for closely spaced bores normally reveals very good agreement, even though the bores may have been put down by different firms for different purposes over a time span exceeding 100 years.

Hand Boreholes

Using a strengthened Hiller sampler the writer put down 599 boreholes to a maximum depth of 10.4 m, the cumulative total bored being 2970 m. The limitations of this equipment have been described elsewhere (Sissons, 1966). Its advantages include cheapness, single-handed operation, speed, and very accurate recording of the stratigraphy. The last point is particularly important since even the most detailed commercial bore records, owing to the nature of the equipment and the methods of operation, cannot be as accurate as data obtained with the Hiller sampler. The small errors in altitude and thickness of beds inherent in the commercial records are not usually significant, but where it was necessary to establish accurate altitudes of deposits or to determine detailed stratigraphical relationships the Hiller was essential. It was also used extensively to study particular buried morphological features and to obtain data in areas where commercial records were not available.

Two spiral augers were also used in fieldwork. One, of larger diameter than the Hiller sampler, was used to penetrate the strong surface 'crust' of the carse deposits, in order to reduce wear on the Hiller sampler. An auger of smaller diameter than the sampler was welded to a

Hiller rod and used to penetrate deposits into which it was inadvisable or impossible to force the Hiller sampler. This small auger was also sometimes used as a percussion instrument to produce a hole through layers of gravel or coarse sand, this hole then being enlarged by the Hiller sampler and boring with the latter being continued to greater depths than would otherwise have been possible.

Altitude Determinations

All altitudes quoted relate to Ordnance Datum. The hand boreholes were all accurately levelled from Ordnance Survey bench marks, the latter being checked against one another. Many of the logs for commercial boreholes, including almost all the foundation bores, give ground surface altitude as determined from bench-marks. Where the altitudes of bores were not given they were estimated. It is believed that most of the estimates for bores in the carse are no more than 0.3 m in error. Estimation to this degree of accuracy is possible owing to the gentle gradients and simple form of the carse surface, the morphology of which has been mapped, and to the large amount of levelling data available. The latter comprise measurements along former shorelines and along traverses perpendicular to these shorelines, Ordnance Survey altitudes abstracted from 1 : 2500 plans and from all editions of 1 : 10,560 maps (allowance being made for any discrepancy between such altitudes and the natural ground surface), known altitudes of commercial boreholes, altitudes of hand boreholes, and a large amount of altitudinal information obtained from other levelling by the writer. From these data a map showing contours at 0.6 m intervals was constructed for most of the carse and used in estimating the altitudes of commercial boreholes for which this information was not available.

The most serious difficulty relating to altitude has been caused by mining subsidence. Fortunately this has not occurred at all in the carse south of a line extending from the mouth of the River Carron to the northern part of Falkirk and has taken place only in small parts of the area covered by the Main Post-glacial Raised Beach (Fig. 1). It is most serious north of the Carron between grid lines 89 and 91 E., extending eastward to the coast south of grid line 84 N. Here it amounts to 3 m in very small areas. That subsidence has occurred can usually be readily seen from the surface form of the carse and it was therefore often possible to avoid subsidence areas when selecting sites for hand boreholes. In some instances it was necessary to put hand bores down in such areas (although the worst affected locations were avoided) in order to supplement commercial bore records or to investigate particular morphological or stratigraphical relationships.

The possibility of correcting all the bore altitudes in subsidence areas was considered, since the amount of subsidence can usually be determined accurately from the distortion of the ground surface, of the buried gravel layer, and of a widespread and faintly inclined shell bed that occurs in the carse deposits beneath much of the lower part of the carse. Since, however, all the information needed for such corrections throughout the area involved is not available, they were not attempted. Instead, attention will be drawn to instances where the data presented in maps and sections may be misleading if subsidence is not taken into account.

Maps and Sections

All maps are based on data plotted originally at a scale of 1 : 10,560 or at a larger scale. In constructing the sections it was not always possible (except in Fig. 9) to use bores located in straight lines: appropriate vertical planes were therefore selected and bores in a belt up to about 150 m wide were projected into them, excepting only bores where marked horizontal variation

in the strata in a direction perpendicular to a projection plane would have produced a misleading representation of the strata. Where a change in the stratigraphy is not recorded in a commercial bore log but is nevertheless shown in a section at the point (or projected point) where the bore was put down (being based on evidence from adjacent bores), this is indicated by a break in the vertical line representing the borehole. Although the sections differ in scale, the same vertical exaggeration (17.6) has been used for all of them.

The Bedrock Surface

The bedrock of the area consists almost entirely of slightly folded but considerably faulted Carboniferous sedimentaries, comprising relatively thin beds of sandstone, shale, limestone, fireclay and coal. The principal igneous occurrence is a thick east–west dyke that occurs a short distance north of the River Carron but is not exposed at the ground surface. Outcrops of the sedimentary rocks are of very limited extent in relation to the total area involved and are usually sandstone. Many of these outcrops are in the slope that backs the carse (Figs. 1 and 2), a result mainly of removal of superficial deposits by marine erosion. Other rock outcrops in the area south and west of the Forth have resulted from river incision (for example, where the River Avon emerges from the higher ground on to the carse, and the River Carron near Larbert) or from artificial excavations, while glacially eroded sandstone emerges from beneath the late-glacial marine deposits in the extreme north-west.

The form of the buried rock surface, so far as it can be inferred from available data, is shown in Figure 2. Only limited interpolation between control points has been allowed in the construction of this map and it therefore fails to bring out the remarkable trench that underlies the Forth. Rockhead altitudes recorded in bores in this area are consequently shown on the map and comparison of them, along with other data shown, enables an impression of the general form of the trench to be gained. It is evident from Figure 2 that the main rock features have a strong trend in a direction that is slightly south of west to slightly north of east. It is probable that this trend is mainly a result of glacial erosion, for it is approximately parallel with the trend of ice-moulded features in the ground immediately to the south (Sissons, 1967a, Fig. 38) and with the course of the Lennoxtown boulder train (A.M. Peach, 1909). Where fault lines parallel the trend they appear to have determined in some areas the locations of belts of maximum erosion. Thus the axis of the west–east depression that extends from the western margin of Figure 2 to pass south of the Hill of Airth corresponds for the most part with one or other of two faults. The Carron trench, extending from near the south-west corner of Figure 2 to Grangemouth (where it merges into the much deeper and wider trench under the Forth), corresponds very closely with a major fault for much of its length. This fault and several east–west faults to the north of it are heading for the area of very deep erosion beneath the Forth where they cross the present coastline, but lack of evidence for the area beneath the Forth precludes correlation.

The origin of the deep trenches beneath the Carron and Forth, and of other parts of central Scotland, has been discussed for a century (for example, J. Bennie, 1871). H. M. Cadell (1883, 1913), although mentioning glacial erosion, constructed maps showing former courses of the Forth and Carron now deeply buried beneath superficial deposits. He concluded that the buried trenches related to a sea level several hundred feet below present, a view that still receives qualified support (for instance, T. N. George, 1965, pp. 45–6). On the other hand, J. M. Soons (1960) argued that the east–west trench that descends to at least − 102.4 m beneath

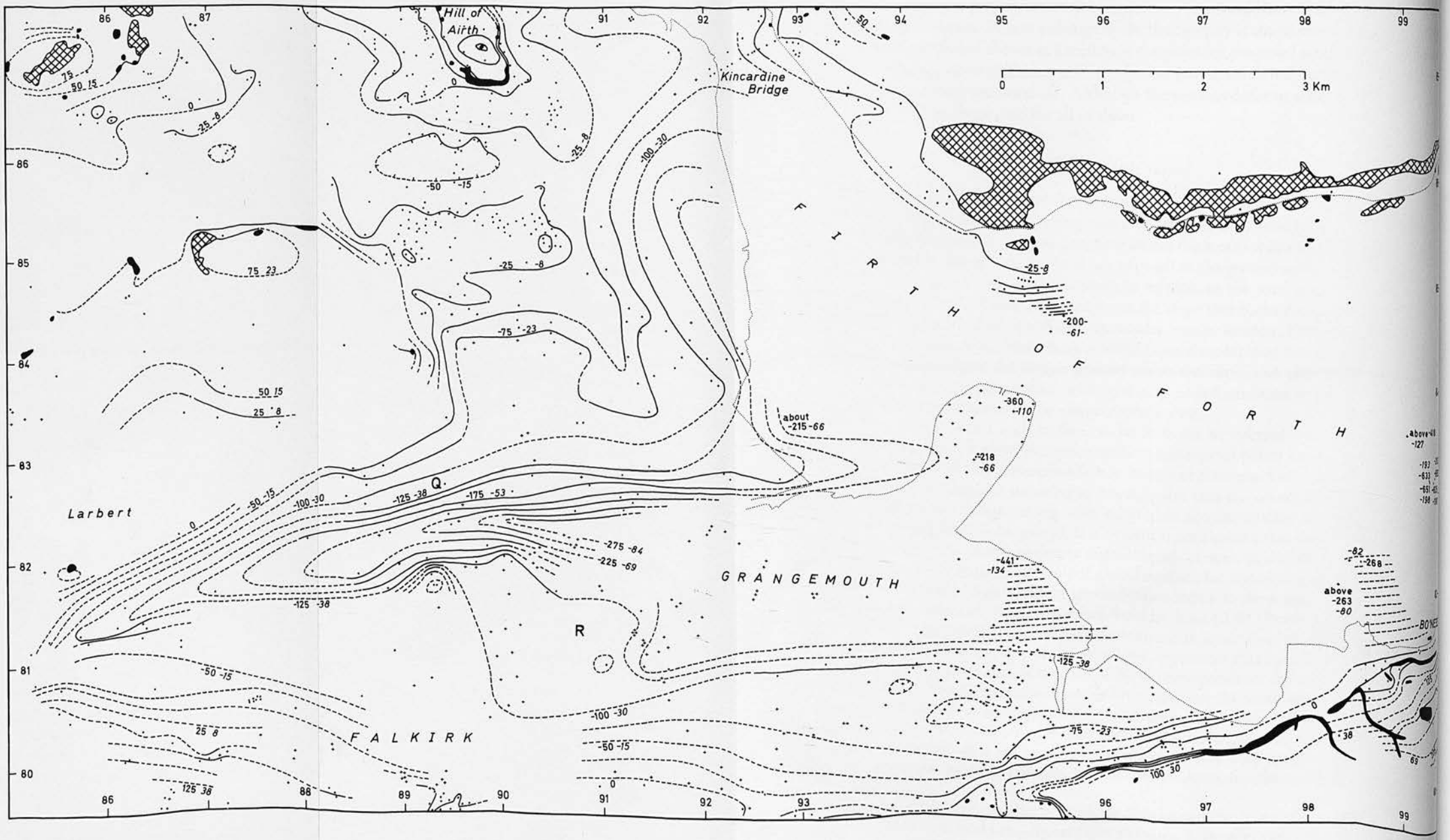


FIGURE 2—Form of the buried rock surface. Rockhead altitudes are shown by contours at intervals of 25 feet (7.6 m). Altitudes are given in metres (italic print) and feet. Bores reaching rockhead are shown by dots; bores not reaching rockhead are shown by crosses. Rock outcrops are shown in solid black or by cross-shading. Q and R mark the end points of the section given as Figure 3.

the lower Devon valley (a few kilometres north of the area covered by Figure 2) has been produced by glacial erosion, while D. L. Linton (1963) and the present writer (1967a, p. 53) have applied this interpretation to similar features elsewhere in the Central Lowlands.

Subaerial river erosion is excluded and glacial erosion strongly favoured by evidence that the deep hollow under the Forth is a closed rock basin. Rockhead altitudes of −190.5, −192.9, −198.4 and −202.7 m have been proved in four closely spaced up-bores from coal workings under the Forth (Fig. 2), while about 2 km farther east rockhead has been encountered at −205.7 m. On the other hand, a seismic refraction traverse across the Forth immediately east of Edinburgh indicated that rockhead is not below −60 m (W. S. Drysdale, 1956).

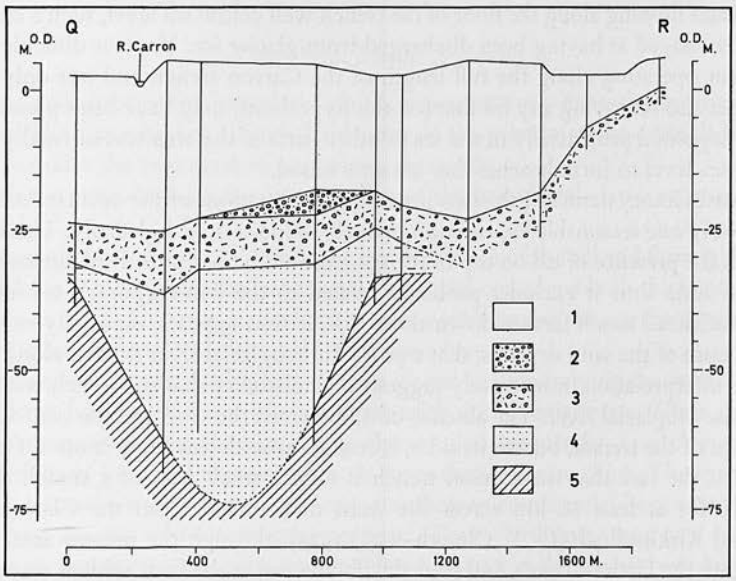


FIGURE 3—Section across the Carron trench between points Q and R in Figure 2, based on commercial borehole data. 1. Late-glacial clay and overlying deposits; 2. Sand and gravel; 3. Till; 4. Sand; 5. Bedrock

The bores in the northern part of Figure 2 are very relevant to the origin of the trench under the lower Devon valley. The closely spaced bores on the line of Kincardine Bridge and the bores to the east and west of this line, coupled with rock outcrops, entirely exclude the possibility of a deep buried channel existing here. The lowest point located on the rock surface is only −21 m, in marked contrast with the figure of −102.4 m in the lower Devon valley and also with a figure of −109.4 m in the Forth lowlands west of Stirling (*Summ. of Prog.*, 1962, p. 48).

While such evidence strongly favours glacial erosion, there is also evidence supporting the older idea of river erosion if one substitutes subglacial for pre-glacial rivers. In the Carron trench, for which there is a considerable number of old but detailed bore records, almost all the bores give the same general sequence, comprising post-glacial and late-glacial marine clays and associated deposits, resting on till, which in turn overlies sand (with occasional gravel) that is as much as 45 m thick (Fig. 3). The flow of water that deposited this sand cannot have taken

place subaerially since (i) the deepest part of the trench lies well below the minimum altitude of rockhead reported by Drysdale (1956) much farther east, and (ii) raised shorelines show that when glacier ice was last decaying in this area, sea level was as much as 90–120 m above the bottom of the trench. In relation to the latter point it might be argued that, since the thick sand is covered by till, the sand may have been deposited subaerially during a period of deglaciation prior to the last one. On the other hand, it seems unlikely that, during a preceding deglaciation, isostatic and eustatic factors combined in a manner so different from that in which they combined during the last deglaciation as to expose the Carron trench (or part of it) to subaerial processes.

Another possible explanation of the presence of the thick sand is that it was deposited by a current of water flowing along the floor of the trench well below sea level, such a current being most easily envisaged as having been discharged from glacier ice. Yet it is difficult to imagine such a current operating along the full length of the Carron trench and not only depositing thick sand but also removing any till that (on this hypothesis) may have been present. Furthermore, sand deposited proglacially in the sea in other parts of the area was normally built up to the existing sea level to form beaches that are now raised.

The unsatisfactory nature of the two preceding explanations of the occurrence of the thick sand leaves only one reasonable explanation, namely, deposition subglacially. This explanation accords with the presence of till on top of the sand and with the linear distribution of the sand, while at the same time it excludes problems related to the contemporary sea level and the altitude of rockhead much farther down the Forth. It thus appears, especially in view of the width and depth of the sand deposits, that a powerful subglacial river flowed along the Carron trench. This interpretation immediately suggests the possibility that the trench was cut, at least in part, by the subglacial river. The absence of till between the sand and the bedrock, not only in the bottom of the trench, but on its sides, is consistent with this suggestion. Of considerable significance is the fact that the Carron trench is only a small part of a trench that extends continuously for at least 80 km across the waist of Scotland from the Clyde estuary, by Bearsden and Kirkintilloch (C. T. Clough *et al.*, 1925), through the present area, continuing thence beneath the Firth of Forth, and probably fading out as the firth widens east of the Forth bridges. Viewed as a whole, this trench bears little resemblance to features in non-mountainous areas that have been attributed to glacial erosion, but along much of its length its dimensions are comparable with those of many of the tunneldale of eastern Denmark and the Rinnentäler of northern Germany, features that are generally considered to have been formed mainly by great subglacial rivers.

It is therefore suggested that the trench extending across central Scotland is of composite origin, glacial erosion having been dominant where it is widest and deepest (especially beneath the Firth of Forth) and meltwater erosion dominant, probably at very considerable depth beneath the ice surface, where it is relatively narrow (for instance, the Carron trench).

Glacial Till

Till occurs widely beneath the post-glacial and late-glacial marine deposits. Occasionally it exceeds 30 m in thickness, the greatest thickness recorded being 56 m in a bore on the coast in the northern part of Grangemouth. Till is often absent, however, near the margin of the carse, having been removed by marine erosion, while in places it has been planated by marine action (p. 32). Beneath the Grangemouth petrochemical works, closely spaced bores, along with

evidence from the rates of penetration of test piles, reveal the occurrence of buried ice-moulded ridges and depressions approximately parallel in trend with similar visible forms in the country to the south.

In some old bores the till is described as hard and most of the tests carried out in association with recent foundation bores show that it has a high bearing capacity, evidence that suggests that most of it is lodgement till. In places, the upper part of the till is described as soft, suggesting it is ablation till or, rather, a modified form of ablation till laid down under the sea. In a few bores recorded in detail by officers of the Institute of Geological Sciences, Highland erratics are recorded in addition to local material. Apart from this, little detailed information of direct relevance to geomorphological history is available for the till.

Late-glacial Marine Sediments

As mentioned previously, late-glacial clay occurs extensively beneath the carse deposits and also occurs widely at a higher altitude landwards of the margin of the carse. It is convenient initially to consider the two areas of occurrence separately.

Beneath the carse deposits the upper surface of the late-glacial clay is an erosional surface sloping gently towards the Forth. The clay almost always rests on till or on the patchy fluvioglacial deposits that overlie the till. Since the buried surface of the till and fluvioglacial deposits varies considerably in altitude, the late-glacial clay varies considerably in thickness. It is usually more than 10 m thick, however, exceeds 20 m over a wide area in the northern part of Grangemouth, and is more than 45 m thick along the coast here.

Numerous particle size analyses of the late-glacial clay underlying the carse deposits have been carried out by various firms in connection with site investigations. These show that the clay is more accurately described as a silty clay or, less frequently, as a clayey silt. It also contains partings of fine sand that are often no more than 1–5 mm thick. The spacing of these thin sand layers varies between 5 and 30 cm according to commercial bore records, although in hand bores a spacing of 1–15 cm was measured. These rhythmites may well be varves. Ice-rafted stones are common in the clay, and boulders up to 0.6 m thick have been bored through. At some bore sites the late-glacial sediments are sandy for part of their thickness and analyses show that these slightly coarser sediments are usually silty fine sand. These silty sands almost always contain many coal fragments, but the latter are hardly ever, if at all, recorded as occurring in the laminated clay itself. The clay has a low bearing capacity and is variously described as very soft, soft or firm, the many foundation piles that enter it continuing through it to an underlying resistant stratum (usually till, occasionally fluvioglacial gravel or bedrock). Shells or shell fragments are only rarely reported in borehole logs as having been observed in the late-glacial sediments (in contrast to the numerous references to their presence in the overlying carse deposits). Bennie (1871) reported the presence of foraminifera in the clay near the Carron.

The above description may give the impression that the late-glacial sediments are of rather variable composition. It must therefore be emphasized that the description is based mainly on about 100 very detailed records in and eastward from the Grangemouth petrochemical area. Almost all the remaining records for bores in the carse, including some in the petrochemical area and including many recent foundation bores, describe the late-glacial sediments far more briefly. Often they are recorded simply as clay or silty clay.

On the other hand, the bores in the late-glacial marine sediments that lie inland of the

carse reveal a clear stratigraphical succession. This succession has been briefly described elsewhere (Sissons and Smith, 1965, pp. 152–3) and this description can now be amplified. The landward margin of the late-glacial marine sediments coincides with the Main Perth Raised Shoreline where the latter feature is present. The shoreline declines from an altitude of about 36 m at Larbert to a calculated altitude of about 29 m at the eastern margin of Figure 1. Where the shoreline is not present, the marine sediments have not been recorded in any borehole above its calculated altitude, but have been frequently reported in boreholes below this altitude. The relation between shoreline and sediments has recently been clearly shown in connection with the motorway currently under construction. About 1½ km west of the sharp bend in the River Avon, the motorway leaves the carse and climbs to the higher ground through a cutting about 10–20 m deep and 700 m long, the upper part of the cutting being in late-glacial marine sediments whose landward margin was previously interpreted as the Main Perth Raised Shoreline (Sissons and Smith, 1965, p. 151; shown in Figure 2 of that paper with the shoreline at 100.8–102.9 feet (30.7–31.4 m)). Commercial boreholes put down before excavation work began recorded firm or stiff, brown laminated or mottled clay, sandy clay or clay with sandy partings, with thin overlying sand, to a maximum altitude of 31.3 m. The bore evidence has since been amplified by the vast excavation, in which the thinning-out of the marine sediments towards the raised shoreline and their termination at that shoreline has been clearly shown. A correlation between morphology and sediments is thus established and the suggestion of F. M. Synge and N. Stephens (1966, p. 117) that sea level in this general area may have been higher than the Main Perth Raised Shoreline receives no support here.

Southward from Larbert, outwash, believed to have been laid down at or near the limit of the Perth Readvance, overlies the marine deposits. Sections in a gravel pit (NS 855 818) showing this relationship have been described (Sissons and Smith, 1965, p. 153) and the sequence may be briefly summarized as delta-bedded outwash gravel, sand with clay layers increasing in number downwards, laminated clay, till, bedrock. The original ground surface altitude here was between about 24 and 30 m. Numerous bores in the area south and south-east of the gravel pit have now become available and reveal the same succession in ground whose surface is at about 23–32 m. Farther away from the presumed ice limit (for example, eastward to Falkirk and north-eastward beyond Larbert) the gravels die out and the marine succession comprises sand merging by alternation downwards into laminated clay. Sand pits and other artificial sections show (or have shown) up to 12 m of sand with a little small gravel and occasional thin bands of laminated clay (e.g. NS 875 811). A bore record on the western side of Falkirk at NS 8746 8062 at an altitude of 23.3 m and typical of a series in the vicinity reads as follows:

	m	m
Fill	1.4	1.4
Coarse brown sand	9.4	10.8
Grey sandy silt	2.8	13.6
Brown plastic clay with silt partings	11.6	25.2
Brown sand and some small gravel	0.9	26.1
Grey boulder clay	0.2	26.3
Bedrock		

North of the River Carron a representative borehole at 26 m located at NS 8774 8334 is recorded as follows:

	m	m
Soil	0.3	0.3
Sand	4.0	4.3
Muddy sand	9.1	13.4
Plastic clay	4.6	18.0
Gravel	0.9	18.9
Boulder clay	2.7	21.6
Bedrock		

On the eastern side of Falkirk the sand becomes thinner and occurs mainly near the southern margin of the marine deposits. Northward, between this restricted sandy area and the carse, late-glacial clay is at the surface over a considerable area. A representative bore at NS 8910 8076 (natural ground surface about 15 m) proved 16.2 m of brown silty clay with sand and silt lenses, resting on 1.4 m of silty fine sand, followed by till.

Still farther east, at the deep motorway section, the surface sand, as revealed by bores and very long sections, is only 0.3–1.0 m thick. It rests on laminated clay containing fine sand laminae and thin beds of sand as well as numerous ice-rafted stones. From a feather edge at the raised shoreline the clay increases in thickness to about 14 m. The clay rests on sand that contains a few thin layers of laminated clay in its upper part.

The late-glacial marine sediments inland of the carse edge thus have a well-defined stratigraphy. The diminution in sand thickness away from the outwash deposits, coupled with the fact that the surface of the sand forms the Perth raised beaches, accords with the view that the sand was provided by meltwater discharged from glacier ice whose limit was situated near or west of Larbert. This evidence is opposed to the suggestion of Synge and Stephens (1966, p. 117) that the Perth raised beaches are 'made up of delta fragments deposited in the sea as the ice front receded westward'.

The late-glacial marine clay was referred to as 'the red clay' by H. M. Cadell (1883, 1913) and others. In commercial borehole records its colour is often given as red, red-brown, brown, chocolate brown, or rich chocolate brown, and shades of red and brown were observed in samples from many hand bores. Despite these minor colour variations the deposit is readily identifiable in hand bores and is quite different from any other deposit in the area. The clay may owe its colour to derivation from Old Red Sandstone (Cadell, 1883), an extensive outcrop of which occurs in the Forth basin west of Stirling. An additional factor may be erosion of lava outcrops (possibly by meltwater in the Campsies and associated hills). In marked contrast the deposits of the Main and Low buried beaches of the Forth valley are a distinctive light grey colour. These latter deposits appear to represent the finer material discharged into the Forth valley from glacier ice particularly during the period of ice decay that followed the Zone III Readvance (Sissons, 1966). Since the buried beaches extend up the Forth valley almost to the edge of the Highlands and abound in mica, it may be inferred that their constituents were largely derived from the Highlands. Thus the colour contrasts between the red-brown deposits and the grey ones may well reflect the sources from which they were principally derived. It is also possible, however, that climatic factors may have had some influence, for J. J. Donner (1957) has shown that in some lake deposits, where changes in source of supply of material are inapplicable, Zones I and III are represented by pink clays and Zone II by organic clay.

While the late-glacial marine clay is red or brown in colour over a large part of the area under consideration, a change occurs towards the Forth in and east of Grangemouth. A plot of

the colour of the clay as recorded in the large number of foundation bores available for this area showed a change from brown in the south, through grey-brown, to grey towards the present shoreline. While colour descriptions in commercial bore logs cannot be considered individually significant, the fact that a plot of data from a variety of sources revealed a clear pattern (apart from a very few anomalies) suggests that this pattern is meaningful. The bore evidence also indicates that the grey deposits overlie the brown (or grey-brown) ones and thicken northward. This may be interpreted to indicate that, during the time the greater part of the late-glacial clay accumulated, glacier ice extended on to or beyond the Old Red outcrops west of Stirling, but that, during the later part of the time it accumulated, glacier ice became restricted to the Highlands, so that material from Highland (and local) sources became increasingly important, ultimately becoming dominant. If the climatic interpretation of the colour of the deposits is considered significant, a not dissimilar pattern of events may be inferred, the red or brown clay implying cold climate and considerable glacier ice cover and the change to grey clay an amelioration of climate and glacier decay.

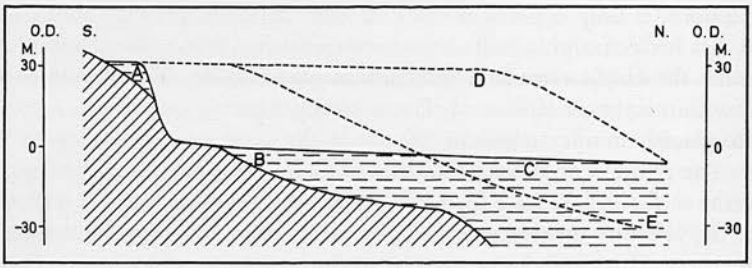


FIGURE 4—Diagrammatic section to show the relation of the higher to the lower late-glacial clay. Late-glacial clay is shown by broken horizontal shading. Fluvioglacial deposits, till and bedrock are shown by oblique shading. For explanation see text.

The colour change in the late-glacial clay is also significant in another context. Eastward from the vicinity of grid line 91 E., the higher area of late-glacial clay has been separated completely from the sub-carse area or entirely removed by later erosion. The two series of deposits have also been separated by erosion at the Hill of Airth (Fig. 6) and south of Letham Moss. West of Letham Moss, however, and also southward from the Carron valley and then eastward to about grid line 91 E., numerous boreholes show that the late-glacial clay beneath the carse is continuous with the late-glacial clay that occurs landward of the carse edge. Yet this does not fully establish the stratigraphical relationships, for the surface of the late-glacial clay beneath the carse is an erosional surface. The point is illustrated diagrammatically in Figure 4, which represents a section about 3 km long extending northward from the motorway cutting to the present coastline, and from which stratigraphical evidence that is irrelevant in the present context has been omitted. It appears from this diagram that the mass of more or less horizontally bedded late-glacial clay at A, at an altitude of about 17–30 m and sharply truncated by erosion on its northern side, is younger than the sub-carse late-glacial clay whose eroded surface at B is at present sea level a few hundred metres to the north. It might appear, therefore, that the beds at A are younger than those at a point C towards the present coastline, as would be the case if, for example, the deposits at A originally extended northward to D at a relatively high altitude. That the reverse holds true, however, is demon-

strated by the colour change from brown to grey in the sub-carse late-glacial clay. The line AE thus represents the type of limit to be envisaged for the red or brown clay.

Considering the late-glacial clay and associated deposits as a whole, it is suggested that they may be interpreted as follows. Glacier decay, before the Perth Readvance, enabled the sea to penetrate westward, meltwater activity during this decay phase being partly represented by gravels and sands at the base of the marine succession (as in the bore logs pp. 28–9). The overlying marine clay, which extends westward as far as Larbert, implies withdrawal of the ice, a subsequent readvance (the Perth Readvance) being suggested by the upward change from clay to sandy clay to sand, this sequence being capped by delta-bedded outwash gravels at and near Larbert. More than 15 m of laminated clay occur beneath the sands and clayey sands in places, and it may be inferred that near the motorway cutting, where the 14 m of laminated clay with thin beds of sand is clearly the remains of a much more extensive deposit, as much as 30 m of these materials may formerly have existed. The time interval represented by these deposits is not known. It may be observed, however, that the sequence of deposits is similar to that recorded by J. B. Simpson (1933) in the Almond valley, near Perth, where he estimated from sample measurements of varved marine clay (covered by outwash) that a time interval of about 640 years was represented by 12 m of clay. In the central Forth area (and around the head of the Firth of Tay) the marine limit is represented by the Main Perth Raised Shoreline. Thereafter, as glacier ice decayed, sea level fell relative to the land, this fall being rapid for a time (Sissons and Smith, 1965, p. 164). With continued ice decay, the deposition of red or brown clay was succeeded by the accumulation of grey clay, glacier ice by this time probably being restricted to the Highlands and relative sea level probably not being very different from that of the present. There then followed a remarkable change in marine conditions for, instead of extensive deposition, extensive erosion took place.

The Buried Gravel Layer and Associated Features

Apart from recent brief references (Sissons *et al.*, 1966; Sissons, 1967a) the major phase of erosion represented by the buried gravel layer has not been previously recognized in the story of relative sea-level changes in the British Isles. In view of this, the distribution and altitude of the gravel layer and associated features were investigated in considerable detail, the results being presented as fully as possible in Figure 5.

Extent

The gravel layer and related planation forms have been identified beneath the carse deposits over 28 km^2 of the present area and comprise the most extensive of the abandoned marine features. Borehole evidence shows that the gravel layer is well developed in some other parts of the Forth valley (for instance, Bo'ness). South-east and south of Grangemouth the inland edge of the buried feature is situated a short distance seaward of the landward margin of the carse. A similar relationship is either known to apply or probably applies as far north as Letham Moss (except in the Carron valley) and along the eastern and southern sides of Hill of Airth. West and north-west from this hill the gravel layer is limited by bedrock that rises close to but does not pierce the carse surface within the area investigated. West of Letham Moss the inland edge of the gently inclined gravel sheet swings away from the vicinity of the carse edge and cuts across the entrance of the large carse embayment that exists here, apparently a reflection of the sheltered position of this embayment (Fig. 5).

The seaward limit of the gravel layer is established in the Grangemouth docks peninsula, since the layer was found only in the southernmost bore of a large number of commercial bores put down here. Between the docks peninsula and the eastern margin of Figure 5, the gravel layer has been proved beneath part of the mudflats by commercial boring, but detailed records are not available. North-west of the docks peninsula the gravel layer has not been recorded in commercial bores towards the present coastline (the area left blank in Figure 5). Its seaward limit is probably situated not far to the east of the easternmost bores for which its altitude is given in Figure 5.

Thickness and Composition

Over almost all the area where it lies below Ordnance Datum the gravel layer is nearly always between 0.3 and 1.5 m thick, thicknesses between 0.3 and 0.9 m being most usual. At higher altitudes the layer is, on average, distinctly thinner, and in some commercial bore records it is not mentioned. Hand boring very close to some of these commercial bore sites showed that the layer is in fact present but consists merely of a few centimetres of gravel, partly mixed with the underlying red or brown late-glacial clay.

In commercial bore logs the gravel layer is variously described as gravel, sand and gravel, gravel and clay, boulder bed, boulder clay, channel, etc. In the more detailed records cobbles and boulders are quite frequently mentioned. Occasionally the layer is represented in a bore by a single boulder, one of which was bored through for 0.9 m. Some of the more detailed records mention marine shells and shell fragments mixed with the gravel. Most of the stones could have been derived from local sources, but Highland rocks such as schists are also present.

Planated Till

There is considerable evidence of till having been planated in association with the formation of the gravel layer. The largest area of planation lies south-east of Kinnaird House and extends for 1.6 km in a north–south direction with a width ranging from 0.7 to 1.4 km. The possibility that this area of planated till as shown in Figure 5 represents a misinterpretation of the superficial deposits as recorded in bore logs was rejected in view of the thicknesses and strength of the material as recorded in different bore logs from various sources. Old logs include records of 6.2, 6.4, 11.9, 4.3, 7.9 and 8.5 m of clay and stones, 6.2 m of sandy clay and stones, 10.1 m of blue till and stones, and 5.8 m of hard boulder clay and stones, while recent foundation bore records refer to 3.8, 0.9+, 1.5+, 4.9+, 4.3, 1.7+ and 2.1+ m of stiff dark brown boulder clay. In the foundation bore records the gravel layer is reported as immediately overlying the till. Eastward and southward from the planated area, bores show that the till surface slopes down to much greater depths, while within the planated area the small area of non-planation of till shown in Figure 5 indicates that the level of planation was slightly above the lowest part of the pre-existing irregular till surface.

South of the Carron the undulating surface of a buried ridge of till rises sufficiently high to have been planated in five small areas, and another small area of planated till occurs at Dalderse. Four other small areas of planated till occur near the margin of the carse south of Grangemouth. It is likely that the areas of till planation are more numerous and more extensive than shown in Figure 5. For example, stratigraphical considerations make it very probable that planated till occurs in a narrow strip near the carse margin for a considerable distance south of Grangemouth, but only a few commercial boreholes are suitably placed to demonstrate this



FIGURE 5—The buried gravel layer and associated features. 1. Planated bedrock; 2. Planated till; 3. Bedrock (beneath the carse deposits) from which till, fluvioglacial deposits and late-glacial clay have been eroded (or were not initially present); 4. Till (beneath the carse deposits) from which fluvioglacial deposits and late-glacial clay have been eroded (or were not initially present); 5. Altitude in metres of the surface of the buried gravel layer or of rockhead as recorded in hand bores (accurately levelled); 6. Altitude in metres of the surface of the buried gravel layer, of the till surface, or of rockhead as recorded in accurately levelled commercial bores; 7. As for 6, but altitudes estimated (see text); 8. Altitude in metres of rockhead as recorded in those commercial bores where it is not apparent from other map data that rock is referred to; 9. Altitude in metres of the surface of late-glacial clay where the buried gravel layer is not present or not recorded (hand bores underlined); 10. As for 9, but altitudes estimated; 11. Other relevant bores, for example, bores in the vicinity of OP that enable the area of planated rock to be accurately defined, and bores in the Grangemouth docks peninsula in which the buried gravel layer is not recorded; 12. Contours, at 1 m intervals, of the surface of the buried gravel layer (for method of construction see text); 13. Steep bluff (usually in bedrock or till) that normally rises from beneath the carse deposits and is for the most part an abandoned marine cliff; 14. Landward edge of the carse (shown only where other data permit); 15. Marked break of slope at the landward margin of the buried gravel layer, of planated till or of planated rock, being referred to in the text as the Buried Gravel Shoreline where situated at about or slightly above Ordnance Datum. *Note:* all altitudes in categories 5, 6 and 7 in areas without shading refer to the surface of the buried gravel layer where this layer overlies late-glacial clay. The eroded surface of the clay usually lies between 0.3 and 1.5 m lower, except in only a very few bores in which the gravel layer cannot be distinguished from underlying fluvioglacial gravel (for qualification regarding the area west of Letham Moss, see note 2 at end of text). Letters indicate lines of sections in Figures 6 and 7.

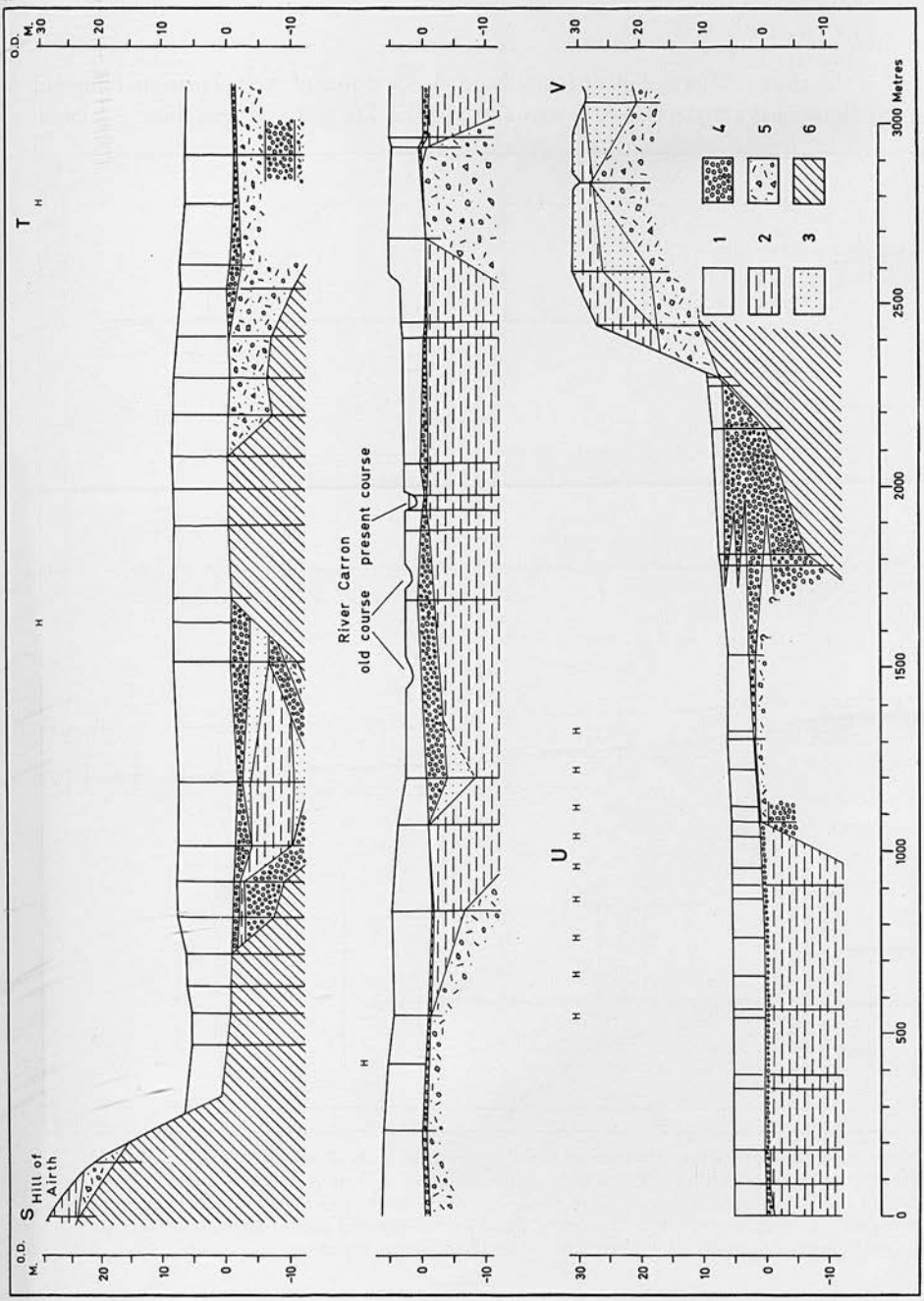


FIGURE 6—Section from Hill of Airth to the motorway cutting at Polmont (S, T, U and V shown in Figure 5), showing the relation of the buried gravel layer to planated late-glacial clay, planated till and planated bedrock. 1. Carse deposits; 2. Late-glacial clay; 3. Sand; 4. Gravel; 5. Till; 6. Bedrock. Based mainly on commercial bores; hand bores are indicated by letter H. Thick gravel between U and V occurs where the Grange burn enters the carse. The ground north of the Carron has been slightly lowered by mining subsidence.

planation. (The hand bores in this area shown in Figure 5 were not able to pass through the gravel layer.)

Planated Rock and Rock Cliffs

There is also clear evidence of planation of bedrock. One area of such planation, lying east of Kinnaird House, has a maximum east–west extent of 1.5 km and averages about 0.5 km in

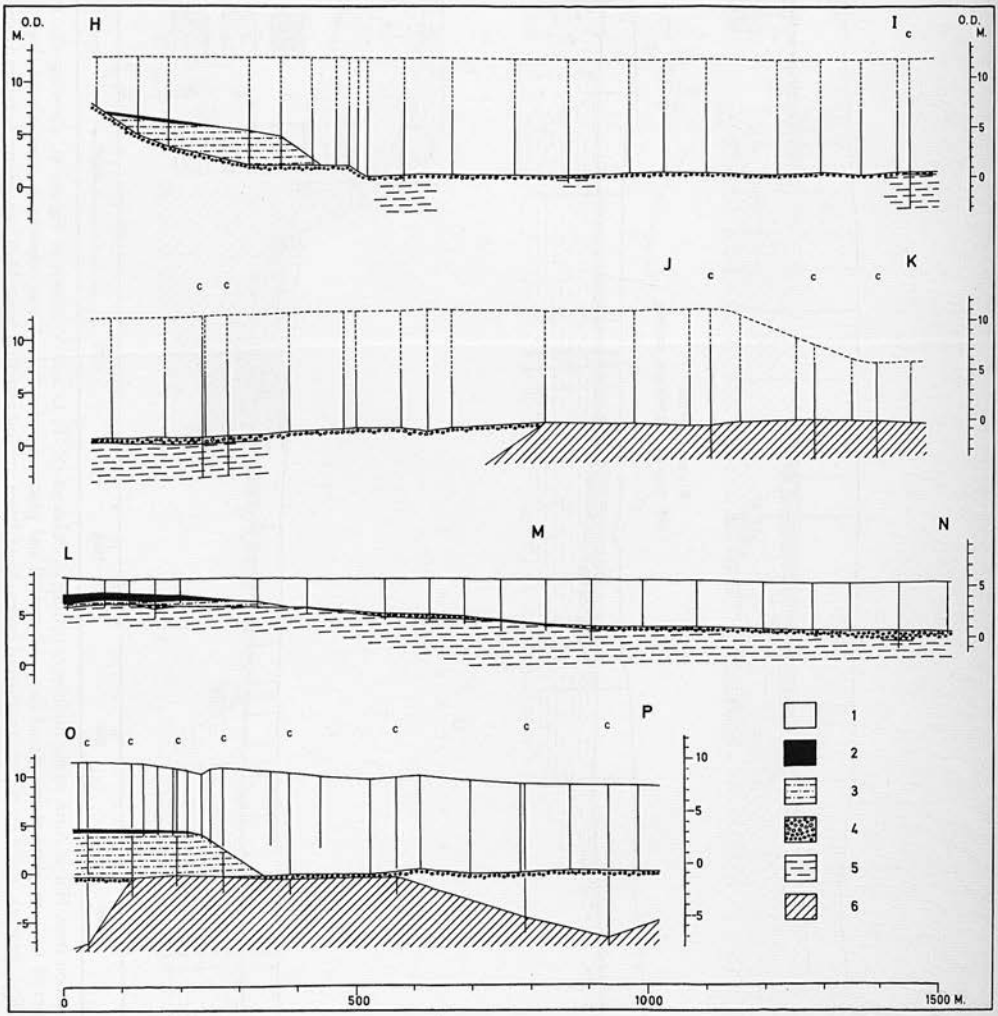


FIGURE 7—Sections showing the relation of the buried gravel layer, planated late-glacial clay and planated bedrock to buried beach deposits. Location of the sections is shown in Figure 5. 1. Carse deposits; 2. Peat; 3. Buried beach deposits; 4. Gravel; 5. Late-glacial clay; 6. Bedrock. Superficial deposits beneath the buried gravel layer in section OP are uncertain. Based mainly on hand bores; commercial bores are indicated by letter C. Section OP has been affected by slight mining subsidence.

a north–south direction. Planation is established by thirty-five bores, thirty-one of which record rockhead between altitudes of about 0 and −1.8 m, not taking into account probable

slight errors in estimating ground surface altitudes (p. 23). The limits of this platform are accurately defined by adjacent bores showing rockhead at lower altitudes, such bores to the west, north and east of the platform being marked by the symbol '×' in Figure 5. Most of the bores in this area are shallow mineral bores put down to locate rockhead, and drift details are not available, but in several bores with drift details the gravel layer is recorded resting directly on the planated rock surface. This surface is separated from the abandoned cliff to the west by a depression in the rock, but three bores suggest that a narrow platform occurs at the base of the cliff, while a bore giving rockhead at 3.7 m suggests that at least the lower part of the cliff is cut in rock.

A pronounced rock cliff is visible on the southern side of the Hill of Airth. It rises very steeply some 9 to 12 m above the carse surface and plunges 6 m below (Fig. 6). The rock platform that meets its base extends for 1.6 km in an east–west direction and over much of this distance averages 0.4 km in width. The existence of the platform is established by twenty-eight bores that encountered rockhead between altitudes of about 0.9 and −1.0 m. A buried rock platform is also demonstrated by bores in the extreme eastern part of the area shown in Figure 5 and a rock cliff lies immediately to the south. This cliff probably represents a steep rock slope resulting from glacial erosion and later modified by marine erosion.

The evidence described briefly above and shown in Figures 5, 6 and 7 demonstrates that the gravel layer and the planated rock and till surfaces are intimately related to each other. It is therefore inferred that all were produced during a single major period of erosion. An alternative interpretation, namely that the planation (especially of rock) took place before the area was last covered by glacier ice, cannot, of course, be conclusively disproved, but seems highly improbable in view of the remarkably clear stratigraphical and altitudinal correlations between the gravel layer and the planated rock and till surfaces.

Form and Altitude

Figure 5 presents all the available altitudinal evidence excluding only data from bores that are too closely spaced to be indicated. It shows the form (and composition) of the surface that would be revealed if the carse deposits, buried peat and buried beach deposits were to be removed. The altitude of the top rather than of the base of the buried gravel layer is given because (i) the former is probably more accurately recorded than the latter in commercial boreholes, (ii) in most hand bores the gravel could not be bored through, and (iii) in very limited areas fluvioglacial gravel, the base of which is irregular, has been planated and this fluvioglacial material cannot be distinguished in bore logs from the gravel layer.

The form of the buried surface is shown by contours at vertical intervals of 1 m in Figure 5. Discrepancies between these contours and the altitudes at boreholes occur partly because no correction of the bore altitudes for mining subsidence has been made, whereas the contours attempt to reconstruct the buried surface as it was before subsidence occurred. To facilitate this reconstruction some of the hand bores in subsidence areas were located on the crests of undulations in the carse surface where little or no subsidence has occurred. A second cause of discrepancies is the slight error in recording the thickness, and hence the altitude, of the different types of drift inherent in the usual commercial methods of boring. Thus the altitudes for the gravel surface as given in commercial bores, in areas where mining subsidence has not occurred, often suggest that this surface is gently undulating, although hand bores put down in such areas in lines approximately perpendicular to the edge of the carse almost always show a remarkably smooth slope declining gently towards the Forth. Contours in non-subsidence

areas were therefore drawn to accord with the hand borehole data where these differed from the commercial data. Where hand borehole data were not available the average values from commercial bores were used, account being taken of the probable relative reliability of these bores (this being related particularly to the purpose for which the bores were put down and to the firm that made them).

The buried surface rises landward from below −6 m in the northern part of Grangemouth to about +6 m in the north-eastern part of Falkirk, its gradient normally being between 1 and 3 m/km. Much the greater part of the feature is below Ordnance Datum. South-east and south of Grangemouth the smooth buried surface normally terminates sharply against more steeply sloping ground, the break occurring at or slightly above Ordnance Datum. All the planated rock areas that have been identified lie close to Ordnance Datum, while the large area of planated till is at a similar altitude. This demonstrates that, for a considerable part of the time the buried feature was being formed, the sea stood at or slightly above its present level. The shoreline that existed at the end of this period of relative stability will be referred to as the Buried Gravel Shoreline (Fig. 5).

For some distance south of the Carron valley a marked break of slope close to Ordnance Datum is not present and the gravel layer, resting on the eroded surface of the late-glacial clay (accordant in altitude with the truncated tops of mounds of till) rises landward to between 3 and 6 m above sea level. This suggests that, after the Buried Gravel Shoreline was formed, sea level rose relative to the land, an interpretation that is supported by other evidence (p. 42). Since erosion continued during the transgression, the Buried Gravel Shoreline, developed in this area in the weak late-glacial clay, was destroyed. In most other areas (with the principal exception of the area west of Letham Moss), by the time the transgression began, the sea had already removed the late-glacial clay and till and had reached bedrock, the surface of which rose inland. Consequently, as sea level rose in these latter areas, it stripped away drift from the rock but extended only a short distance farther inland. This resulted in the preservation of the Buried Gravel Shoreline in these areas and produced areas where only deposits younger than the shoreline rest on the rock surface: till, fluvioglacial deposits and late-glacial clay are missing. Such areas, where identified, are shown in Figure 5 along with small areas in which the late-glacial clay, but not the till, was completely removed.[2]

Source of Gravel

Since the gravel layer truncates late-glacial marine sediments containing ice-rafted stones and is associated with considerable planation of rock and till, it may be inferred that its components were provided mainly by this erosion. The gravel layer is thus similar to present-day beaches along many parts of the coast of the Firth of Forth farther east, where recent erosion has occurred and is still occurring in places. These beaches are littered with stones and boulders of all sizes, mixed with mud, sand and shells, and one may envisage the now-buried gravel layer as having presented a similar appearance when it was being formed. A common origin for the modern and buried features was strongly suggested by work done by Cadell (1883), for he classified in terms of rock type 336 stones from the present shore east of Bo'ness and 373 stones from the gravel layer as then artificially exposed immediately above the late-glacial clay in Bo'ness docks (2.5 km east of the eastern margin of Figure 5) and obtained closely similar results.

It is probable that a small proportion of the constituents of the gravel layer was provided by streams draining the adjacent land. Artificial excavations where the River Avon and the

Grange burn (2.5 km west of the Avon) enter the carse have exposed sheets of gravel fanning out from the valley mouths. These gravel sheets interdigitate with the carse clay and hence are of later origin than the widespread gravel layer, but it seems reasonable to infer that the streams deposited gravel at earlier times. This inference appears to be confirmed in the area west of Letham Moss where river gravel rests on the floors of buried valleys excavated when the widespread gravel layer was being formed.

Age

Since the gravel layer (i) extends into the area believed to have been ice covered during the Perth Readvance, (ii) truncates the late-glacial clay partly associated with this readvance and the subsequent period of deglaciation, and (iii) occurs at altitudes between about 30 and 43m below the Main Perth Raised Shoreline, the gravel layer is evidently younger than the Perth Readvance and its formation could not have begun until some time after the maximum of the readvance. In terms of the provisional date for the limit of the Perth Readvance (Sissons, 1967b) the formation of the gravel layer began after 13,500–13,000 BP.

To determine the relationship between the gravel layer and the buried raised beach deposits, lines of hand boreholes were put down, the resultant data being correlated with commercial bore data where available. Sections HK and OP (Fig. 7) show that the gravel layer passes beneath the buried raised beach deposits, section OP also showing that planated rock occurs beneath these deposits. In the vicinity of the Menteith moraine, three buried beaches, formed in descending order, have been identified and dated to approximately 10,300, 9500 and 8800 BP respectively (Sissons, 1966; W. W. Newey, 1966). In the present area only the two lower beaches have been identified, but since the gravel layer passes beneath both of them it follows that this layer and associated erosional features ceased to be formed at least 9500 years ago. However, the date of 9500 years relates to the base of peat immediately overlying the Main Buried Beach and it thus relates approximately to the time the beach was abandoned by a relative fall of sea level. The terminal date of formation of the gravel layer is thus older than 9500 years by at least the amount of time the buried beach deposits took to accumulate. Since these deposits represent the finer constituents of the debris discharged by glacier meltwater rivers into the sea from the Zone III ice (Menteith moraine limit) in the westernmost part of the Forth lowlands, it may well be that the main period of accumulation of these deposits commenced as the ice began to decay about 10,300 BP, in which case this is the approximate terminal date of formation of the gravel layer.

The fact that the gravel layer passes beneath the buried beach deposits does not exclude the possibility that this layer, where not covered by these deposits, was produced or modified after the buried beaches were formed. This possibility is greatly restricted by other evidence, however. In the carse deposits a very well defined shell horizon containing abundant oysters has been traced over much of the area between Airth and the eastern limit of the carse. The shell layer rises very gently inland from the present coast (or from the original coast before reclamation) for a distance of up to 2.5 km. A radiocarbon date of oyster shells from this layer at NS 9578 8013 gave an age of 6240 ±120 BP (I-1836). If this date is valid, formation or modification of the gravel layer during the last 6000 years or so is entirely excluded. This leaves only one period when this could have happened after about 10,300 BP, namely the period of relatively low sea level that reached its minimum altitude at about 8500 BP (Sissons, 1966). The lowest altitude (in non-subsidence areas) at which buried peat has been found resting on the buried raised beach deposits is 3.9 m, implying that, during the period of low sea level,

high-water mark was below this altitude. However, this period of low sea level was of very short duration and the most that can be attributed to it is limited modification of the gravel layer.

In summary, the considerable planation of rock and till and the extensive planation of late-glacial clay, along with the formation of the extensive gravel layer, must have taken place entirely (or almost entirely) in late-glacial times, beginning, apparently, later than 13,500–13,000 BP and probably ending about 10,300 BP. During a very considerable part of this time the sea in the central Forth valley stood at or slightly above its present level in relation to the land when the Buried Gravel Shoreline and associated features were being formed.

Minimum Sea Level

While it is possible that the Buried Gravel Shoreline marks the minimum level to which the sea fell between the regression from the Perth shorelines and the transgression to the upper limit of the buried raised beach deposits, it is also possible that the Buried Gravel Shoreline was preceded by a lower sea level. In favour of the latter interpretation it is suggested that the very large area covered by the gravel layer and the fact that it descends to at least −6 m, and possibly to −8.5 m, can be more easily accounted for if a transgression from a level below Ordnance Datum preceded the formation of the Buried Gravel Shoreline.

The borehole evidence available at present is insufficient to prove or disprove this interpretation and will therefore be described briefly. Beneath the easternmost part of the carse, the late-glacial clay immediately below the buried gravel layer is replaced over a belt up to 1 km wide by sands and silty sands with occasional gravel. Where the borehole evidence is most detailed, the sands are seen to occupy a broad depression in the late-glacial clay, the slopes of the depression descending regularly to about −12 to −14 m in its central part. The depression appears to begin where the Avon enters the carse and to turn eastward roughly parallel with the edge of the carse. It is tempting to suggest that the depression was cut in the late-glacial clay by the Avon at a time when sea level was lower than the lowest part of the gravel layer and that it was subsequently infilled with sand brought mainly by the Avon immediately preceding the formation of the gravel layer. On this interpretation one would expect to find a similar sand deposit associated with the Carron, but only three bores suggest the presence of such a deposit, while in many it is apparently absent. This problem could not be resolved by hand boring owing to the depths involved and its solution must await future commercial investigations.

Discussion

Certain aspects of the gravel layer and related evidence, some of them of significance outside the area considered in this paper, require further discussion.

(i) The most striking aspect of the gravel layer and associated features is that they prove that a period of extensive marine erosion occurred in an area where otherwise deposition has been almost entirely dominant. One cause of this contrast appears to have been connected with the availability or non-availability of fine debris from glacial rivers. The late-glacial clay apparently represents fine sediment discharged into the sea by meltwater rivers before, during and after the Perth Readvance. Likewise, the buried beach deposits were associated with the Zone III Readvance (Menteith moraine), and particularly with the subsequent period of

deglaciation. Although the Menteith moraine lies some 30 km up the Forth valley from the northern limit of the present area, and although the extent of the Main Buried Beach implies that glacier ice was even more distant when this beach ceased to be formed (being by then largely or entirely restricted to the Highlands), the buried raised beach deposits are clearly represented in the present area. This leads one to suspect that, when the gravel layer and associated features were being formed, glacier ice in the Highland part of the Forth drainage area had dwindled to insignificance or disappeared altogether. Since the headwaters of the Forth (which has the Teith as a major tributary) rise in the heart of the most important centre of former glacier development in the British Isles, it seems likely that, at this time, glaciers in the rest of the British Isles were also insignificant or had disappeared.

Yet lack of fine debris from glaciers does not appear sufficient alone to explain the erosional phenomena. During the last 8000 years or so, when no debris from glacial rivers has been discharged into the Forth valley, carse deposits have accumulated over tens of square kilometres of the valley, while wide mudflats now extend along the present coast at and near Grangemouth. While the evidence that the sea was transgressing on the land during part of the time the erosional features were being formed may partly help to explain this erosion, it assists little in accounting for the principal evidence of erosion, associated with the Buried Gravel Shoreline, since this appears to relate to a period of relatively stable sea level. Furthermore, that widespread deposition accompanied the next major transgression is proved by the gradual merging of peat into the overlying carse clay in numerous places. It is also noteworthy that, for a considerable distance eastward from the present area, the Firth of Forth is only 3 km wide and, being limited by rock, could never have been wider, while the open sea at the mouth of the firth is some 65 km distant. It may also be observed that winds from a westerly point, such as prevail today, would have little effect in producing the erosional features, these features being best developed in rock and till exactly where the fetch, from a direction slightly south of east, was greatest. It is possible that marine planation may have been facilitated in the cool late-glacial climate by frost action related to the semi-diurnal wetting and drying in the intertidal range, but considerable wave action would still be required. In brief, it does not appear possible in the light of present evidence to account fully for the extensive late-glacial marine erosion in this relatively sheltered locality. One therefore wonders if the climate of that time differed from the present climate not only in temperature but also in having a greater storminess and/or a much greater frequency of east winds.

(ii) The Buried Gravel Shoreline and associated evidence record a relatively low sea level before the Zone III Readvance, the formation of this shoreline being followed by a transgression. At the Menteith moraine the culmination of the transgression is represented by the High Buried Beach, which was formed while ice still stood at the moraine, and was followed by a regression of the sea. For the same reasons as were adduced in relation to the Perth Readvance (Sissons and Smith, 1965, pp. 164–6), it may therefore be suggested that the Zone III Readvance resulted in a renewed isostatic depression of the earth's crust.

(iii) Since the evidence for late-glacial marine erosion is so pronounced in the relatively sheltered mid-Forth area, one may expect erosional features produced contemporaneously to be very well developed along more exposed parts of the Scottish coast where the relations between glacio-isostatic movements of the land and eustatic movements of sea level were favourable. Such evidence should be found at increasing depths below present sea level at points increasingly farther removed from the centre of glacio-isostatic recovery than is the Grangemouth area.

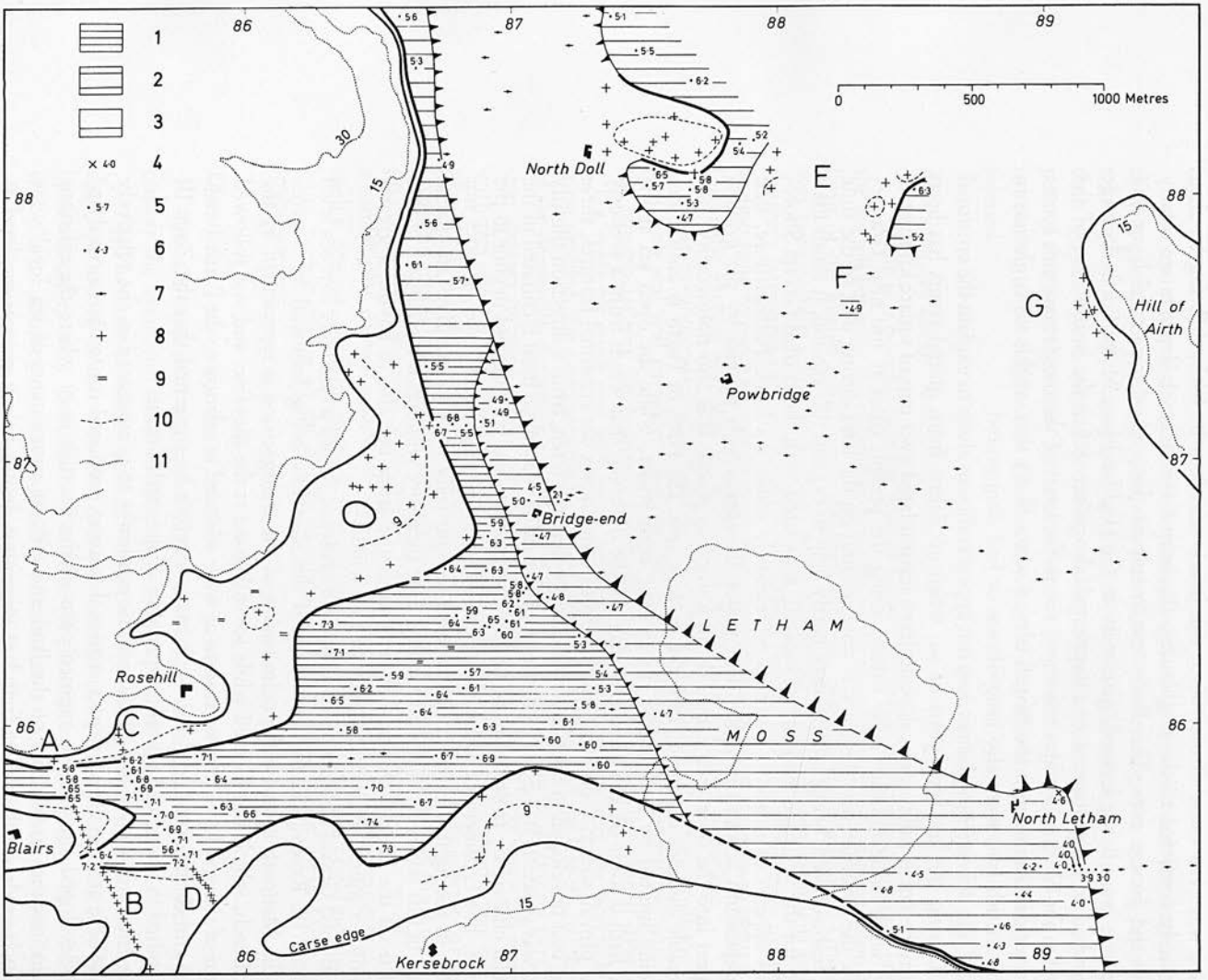


FIGURE 8—Buried raised beaches west of Hill of Airth. Based almost entirely on hand bores. 1. Main Buried Raised Beach; 2. Main and Low Buried Raised Beaches where not distinguishable from each other on present evidence; 3. Low Buried Raised Beach; 4. Estimated altitude in metres of the surface of the buried beach deposits (commercial bores); 5. Altitude in metres of the surface of the buried beach deposits as recorded in accurately levelled hand bores; 6. As for 5, but buried peat too tough to bore through, the buried beach surface altitude being estimated on the basis of the average recorded buried peat thickness in the vicinity; 7. Buried beach deposits absent in a bore carried down below their usual altitudinal range of occurrence; 8. Buried beach deposits absent because the pre-existing surface is above their upper altitudinal limit; 9. Buried beach deposits absent: the surface of late-glacial clay lies at approximately the same altitude as the surface of these deposits in the vicinity; 10. 9 m contour of the land surface (now buried) when the buried beach deposits accumulated; 11. 15 and 30 m contours of the present land surface. AB and CD indicate lines of sections comprising Figure 9. E, F and G are areas where buried beaches were not investigated.

Buried Raised Beaches

Reference has already been made to the buried raised beaches, and their characteristic light grey colour and their abundant mica content have been mentioned. These beaches are mainly restricted to the north-western part of the area investigated, their distribution here being shown in Figure 8. This map is based almost entirely on hand borehole data and shows all the principal occurrences of buried raised beach deposits in the map area, excepting only possible occurrences around and north of E, F and G, where hand bores have not been put down. Beyond the map area the buried beach deposits continue southward as a narrow strip for about 1 km. They have also been located in a narrow strip close to the margin of the carse in the north-eastern part of Falkirk (for example, Figure 7, section NL).

The buried beach deposits consist mainly of light grey micaceous silty fine sand. The uppermost part is sometimes silty clay while the basal part is usually sandier than the remainder. The deposits are thickest in the south-east of the area shown in Figure 8, where they probably average about 4 m. In the west they are often only a metre or so thick. Except along their extreme eastern margin the buried beach deposits are almost always covered with peat, showing that, except along their eastern edge, they were not eroded during the transgression associated with the accumulation of the overlying carse deposits. The maximum altitude at which the buried beach deposits have been encountered is 7.4 m and in the area around Rosehill, Kersebrock and Blairs several bores show the deposits extending up to 7.0–7.4 m (immediately north-east of Blairs the altitudes have been depressed by mining subsidence). In this area and towards Bridge-end the surface of the deposits tends to slope down gently towards the central part of the former coastal indentation they occupy and also slopes eastward, typical altitudes for the lower parts being between 5.8 and 6.4 m. Farther eastward, as around Bridge-end itself, the surface of the deposits is slightly but distinctly lower, altitudes being about 4.5–5.0 m. Beyond the thick surface peat area of Letham Moss, similar altitudes are encountered in the south-eastern part of the area shown in Figure 8, 5.1 m being the maximum recorded and the remaining points lying probably between 4.0 and 4.8 m. Two buried beach levels are thus indicated. In the more northerly parts of the area covered by Figure 8 the altitudes show the higher feature is present in places, but a distinction between it and the lower feature cannot be made on the basis of the evidence available, owing partly to the limited width of the buried beach deposits here and partly to irregular mining subsidence.

The buried beach deposits of the present area are correlated with those previously described in the Menteith moraine area (Sissons, 1966) because, owing to their distinctive colour and composition, it has been possible to trace them almost continuously from the one area to the other by hand boring. As a result of the latter investigation (as yet unpublished) it appears that the two buried beaches of the present area are to be correlated with the Main and Low Buried Beaches of the Menteith moraine area. Deposits that can be correlated with the High Buried Beach of the latter area have not been found in the present area.

South and west of North Letham and also at Bridge-end the buried beach deposits rest on the buried gravel layer, the latter being at or a little above Ordnance Datum (Fig. 7, sections HK and OP; Fig. 8). Westward from Bridge-end the gravel layer rises quite rapidly and a similar rise probably occurs elsewhere in this area in the vicinity of the line marking the Buried Gravel Shoreline in Figure 5. In the broad embayment between Bridge-end, Rosehill and Kersebrock (Fig. 8) a layer of patchy gravel normally only a few centimetres thick underlies the buried beach deposits. This gravel rests on reddish late-glacial clay whose surface over a

considerable area is between 5.5 and 7.0 m, in contrast with the extensive area east of the Buried Gravel Shoreline where the eroded surface of the late-glacial clay is at about Ordnance Datum. The gravel merges upward into the buried beach deposits which, in their lowest 0.3 m or so include medium and coarse sand, whereas the upper part of these deposits is mainly composed of the usual silty fine sand. In some bores, including those where the buried beach deposits rest directly on the late-glacial clay, the lower part of the buried beach deposits is reddish in colour, suggesting that erosion of the late-glacial clay was occurring in the vicinity as they began to accumulate. From the altitudinal data given above and in Figure 8 it will be apparent that in a considerable part of the broad embayment between Bridge-end, Rosehill and Kersebrock the buried beach deposits are quite thin, often being less than 1 m. A similar stratigraphy and buried morphology has been found beneath the carse deposits in the north-eastern part of Falkirk. As shown in section NL of Figure 7, the surface of the late-glacial clay tends to flatten out at about 5.5–6.0 m beneath the buried beach deposits, these deposits being very thin, underlain by patchy gravel, and covered by peat.

The above evidence suggests that, following the period of relatively stable sea level associated with the formation of the Buried Gravel Shoreline, the sea rose fairly rapidly relative to the land until it reached an altitude of about 6 m. Relative stability then permitted some planation of the late-glacial clay at about this altitude, this erosion partly overlapping with and being followed by the accumulation of the Main Buried Beach deposits. In terms of the sequence of events established in the Menteith moraine area, it appears that part of the period of erosion can only be correlated with the High Buried Beach, whose shoreline lies at about 12 m, giving a gradient for this shoreline of about 0.2 m/km.

Buried Valleys

In a few of the hand bores put down in the area in which the buried beach deposits are normally found, these deposits were absent and the late-glacial clay was reached at a lower altitude than usual. For example, near the centre of the embayment north of Kersebrock (Fig. 8) one bore encountered the late-glacial clay at an altitude of 4.4 m and another at 2.8 m, a layer of gravel 6 cm thick resting on the late-glacial clay in the latter instance. These and other bores suggest the existence of a narrow buried valley in the vicinity of the stream that currently drains the area, the deeper of the two bores specifically mentioned probably having passed through the old river channel deposits. This valley, now buried by the carse deposits, was produced in response to the relatively low sea level that reached its minimum level about 8500 BP.

One would expect another system of buried valleys to have been formed during the earlier period of low sea level associated with the formation of the widespread buried gravel layer. Evidence of such valleys was sought for and obtained by putting down two lines of closely spaced bores (AB and CD, Figs. 8 and 9). Section line AB was positioned very slightly east of the points where the divides between three valleys cut into the late-glacial clay disappear beneath the carse surface. The section shows the form of the buried relief, the valley floors being as much as 7.0 m below the ground surface. By the time section CD, some 300 m to the east, is reached, the divides between the buried valleys are much lower. Both sections show that the deposits of the Main Buried Beach extend along the floors of the deeper valleys, thus proving that these valleys were cut before about 9500 BP and suggesting, for the same reason as given on p. 37, that they were produced before about 10,300 BP. The gravel beneath the

buried beach deposits was presumably laid down by the streams that cut the valleys. This latter presumption may appear questionable since this gravel is (or was before mining subsidence) at a similar altitude to gravel farther east that has been interpreted as of marine origin. However, in the middle bore of the line of five bores south of Rosehill (Fig. 8) the gravel, beneath the buried beach deposits, was found between 4.5 and 3.8 m, well below its usual altitude beneath the buried beach deposits farther east in the embayment. In this line of bores the eroded surface of the late-glacial clay (from north to south) was at 7.9 m, 5.1 m, below 3.8 m, at 5.0 m, and below 4.7 m. Tracing the buried valleys farther east by hand boring was found to be impracticable.

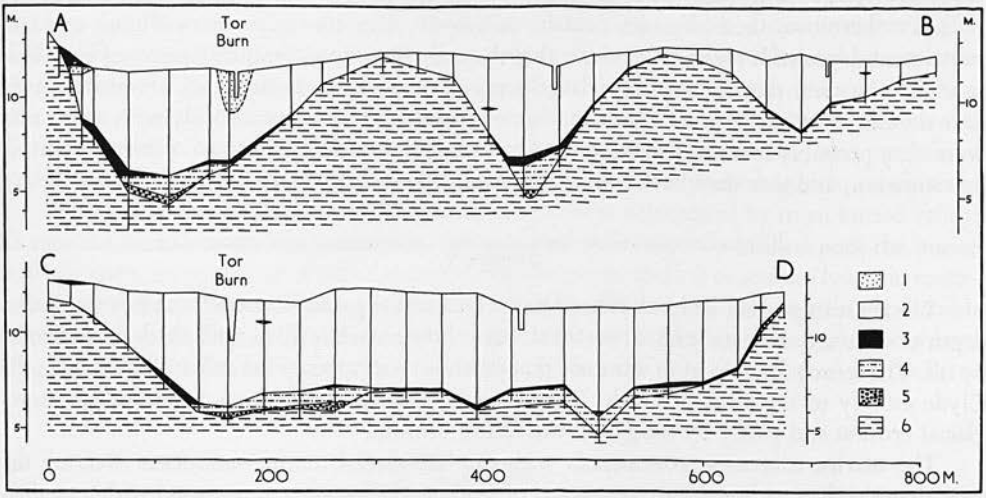


FIGURE 9—Sections across buried valleys on the lines AB and CD shown in Figure 8. Based entirely on hand bores. 1. Sand; 2. Carse deposits; 3. Peat; 4. Buried raised beach deposits; 5. Gravel; 6. Late-glacial clay. Mining subsidence not exceeding 1.2 m has affected section AB between 0 and 300 m.

Since it has been suggested above that buried valleys cut into the buried beach deposits and related to the 8500 BP low sea level occur in the embayment, they may be expected to occur also along section lines AB and CD. Such valleys appear to have been located at 400 m and 500 m in line CD (Fig. 9). The valleys have not been found elsewhere along the section lines, probably owing to their being too narrow to have been located, except fortuitously, with the bore spacing adopted. In this respect it should be noted that the valleys at 400 m and 500 m are almost certainly much narrower than depicted in Figure 9, the width shown having resulted from the method of construction of the sections which, to ensure maximum objectivity, involved joining data from adjacent bores by straight lines wherever reasonably possible.

The streams now flowing across the carse deposits in the area of the two section lines occupy artificial channels for the most part. In some instances these channels follow valleys in the carse that are a few tens of metres wide and a metre or so deep. In other instances (for example, part of the Tor burn) there is no valley at all and only the old stream channel can be detected here and there as a faint depression. Deposits attributable to the present streams consist mainly of sand and are of very limited extent: only one bore, located beside the Tor burn, encountered a considerable thickness of such deposits (section AB, Fig. 9). In great

contrast, the earlier buried valley of the Tor burn is over 300 m wide on section line AB and its floor is covered with gravel over a width of 50 m or so, while in section line CD the almost flat floor of the buried valley, veneered with gravel, is 150 m wide.

The Tor burn and other buried valleys shown in Figure 9 are the continuations of valleys cut into the late-glacial clay and sand constituting most of the surrounding higher ground. Such valleys, up to 18 m deep and several hundred metres broad, many of them streamless or drained by artificial ditches, are very common in such deposits in south-east Scotland (Sissons *et al.*, 1965). The carse deposits and the Main Buried Beach deposits extend a considerable distance along the buried floors of some of these valleys in the present area (for example, both types of deposit were found in bores some 500 m up the Tor burn valley from section line AB). Furthermore, these deposits contain or are overlain by only scanty fluvial or mass-movement deposits. It is clear, therefore, that the valleys are fossil features preserved with little modification since they were formed largely or, more probably entirely, in late-glacial times. Like the extensive buried gravel layer and related features, contemporaneously with which they were very probably formed, the valleys indicate climatic conditions that are worthy of further investigation, and that may have interesting implications for Scotland as a whole.

Summary

The most important bedrock feature in the area investigated is a trench cut to a maximum depth of over 200 m below present sea level. Part of the trench is filled with thick sand covered by till. The trench forms part of a feature that extends continuously for at least 80 km from the Clyde estuary to the Firth of Forth. It is suggested that the trench was produced partly by glacial erosion and partly by subglacial meltwater erosion.

The marine succession commences with the late-glacial marine sediments. Beneath the carse deposits these sediments are composed of laminated clay, silt and fine sand with ice-rafted stones. Inland of the carse deposits these laminated sediments are overlain by sand, the surface of which forms the Perth Raised Beaches. The sand thickens westward and, near Larbert, is overlain by outwash gravel believed to have been laid down at or near the limit of the Perth Readvance. Deglaciation following the readvance was associated with a major lowering of sea level relative to the land, the sea falling from about 29–36 m to about or below its present level. As the ice became more distant, the marine clays and silts that were being deposited changed from a red or brown colour to grey.

A major phase of marine erosion then followed. Late-glacial marine sediments, till and bedrock were planated to produce a surface sloping gently towards the Forth. This planation surface (often backed by a cliff) is the most extensive marine feature (buried or visible) in the area investigated and has been identified over 28 km^2. It is overlain by a layer of gravel that is usually less than 1.5 m thick. The surface of the gravel layer rises landward from below −6 m to about +6 m. During a considerable part of the time during which the gravel layer and associated planation features were being formed, the sea stood at or slightly above its present level, the shoreline that existed at the end of this period of relative stability being named the Buried Gravel Shoreline. The formation of this shoreline was probably preceded by a relative rise of sea level.

The major phase of erosion occurred entirely in late-glacial times, probably commencing after 13,500–13,000 BP and terminating about 10,300 BP. In this sheltered inland area, where widespread deposition has otherwise prevailed during and since deglaciation, the amount of

erosion is remarkable. The reduction of glacier ice to insignificance or total extinction, thus reducing or removing a major source of fine sediment, was doubtless a factor but is alone an incomplete explanation. One therefore wonders if the climate of the time was characterized by a greater storminess and/or a much greater frequency of east winds than the present climate.

Following the formation of the Buried Gravel Shoreline the sea rose relative to the land, probably fairly rapidly, erosion of till and the late-glacial marine sediments continuing in some areas, a thin gravel layer being accumulated on the eroded surface.When sea level had reached about 6 m, a period of relative stability occurred with some planation of late-glacial clay at about this altitude, this erosion partly overlapping with and being followed by the accumulation of the overlying Main Buried Beach deposits, which consist mainly of light grey micaceous silty fine sand. The formation of the Main Buried Shoreline, with a maximum recorded altitude of 7.4 m in the area investigated, ended about 9500 BP and was followed by a relative fall of sea level. The Low Buried Beach, of similar composition to the Main Beach, was subsequently formed, and its shoreline, with a maximum recorded altitude in the present area of 5.1 m, was abandoned about 8800 BP. Thereafter sea level fell relative to the land until about 8500 BP when a major transgression commenced and the carse deposits began to accumulate.

The period of relatively low sea level around 8500 BP is represented by small buried valleys cut into the buried raised beach deposits. Other buried valleys, partly infilled with the buried beach deposits, were very probably formed during the earlier period of low sea level characterized by extensive marine erosion. These latter buried valleys are the continuations of visible valleys, up to several hundred metres broad, cut into the exposed late-glacial marine sediments. These fossil valleys are far larger than the valleys in the carse deposits occupied by the same streams today.

Conclusion

Two general points will be made in conclusion.

(i) In the past, studies of relative sea-level changes in Scotland have been almost entirely concerned with surface form and with deposits where the latter have been exposed fortuitously in natural or artificial sections of limited extent. It is suggested that systematic investigations of the type presented in the preceding pages will have to be carried out in many other suitable parts of Scotland if the sequence of relative sea levels in the country is to be established. In this respect it should be noted that, for approximately half the time since the ice finally disappeared from the area investigated here, the sea was below the level it subsequently attained during the major post-glacial transgression. In other words, the evidence for approximately half the story of relative sea-level changes in this area is buried beneath later deposits.

(ii) Although the preceding pages have been concerned entirely with historical geomorphology, the data assembled and the correlations that have been made are often directly relevant to foundation engineering problems. Some of the data have been published separately in the context of regional planning, including, for example, a map indicating approximately the lengths of foundation piles required for large engineering structures (Sissons, 1968). From the maps and text of the present paper it will be apparent that over much of the area considered it is now possible to forecast the probable drift stratigraphy at sites where boreholes have not been put down. Such forecasts, if utilized (for instance, by putting down two or three suitably located check bores instead of the much larger number of bores normally put down during site investigations) could often save considerable sums of money.

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NOTES

[1] For brevity of expression and in accord with previous usage in Scotland, the term 'raised beach' is used throughout this paper to include 'raised estuarine deposit'. Likewise 'marine' is used to include 'estuarine'.

[2] This explanation is applicable only up to an altitude of approximately 6.0–7.4 m, this being the limit of the transgression (higher in the west than the east) that followed the formation of the Buried Gravel Shoreline. Areas above the transgression limit, and below the limit of the carse deposits, that lack till, fluvioglacial deposits and late-glacial marine clay are extremely small (Fig. 5). The removal of these earlier deposits may have been accomplished during the marine regression that followed the formation of the Main Perth Raised Shoreline and/or during the transgression that was associated with and followed by the accumulation of the carse deposits. It will be appreciated that in some small areas there may initially have been no till, fluvioglacial deposits or late-glacial clay, but that such areas cannot be distinguished from those from which these materials have been removed.

Further explanation of the data shown in Figure 5 is required for the area west of Letham Moss. In this area, landward and above the limit of the buried beach deposits, the carse deposits frequently rest on a layer of gravelly material up to 0.6 m thick. This gravelly material in places passes laterally with diminution in thickness, into the widespread gravel layer, but differs from it in that it is less resistant than the gravel layer normally is, contains bits of peat, and includes many small angular 'gritty' particles. Since the area in which this material occurs was exposed to subaerial agencies for several thousand years before being covered by carse clay, it is probable that the material was produced by weathering and mass movement. In accord with the principles on which Figure 5 was constructed, altitudes on the surface of the gravelly material as recorded in bores are shown in that Figure.

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RÉSUMÉ—*Stratigraphie des dépôts quaternaires et traits morphologiques souterrains dans la région de Grangemouth–Falkirk–Airth, Écosse Centrale.* Cette communication décrit et discute, en partant principalement de l'évidence acquise d'environ 1900 forages, les traits morphologiques souterrains et dépôts quaternaires d'une petite partie de la vallée du moyen Forth. La surface de l'assise rocheuse souterraine est tracée par une levé des courbes et l'on attribue une profonde tranchée raversant la région à l'érosion glaciaire et fluvio-glaciaire. D'épais dépôts sédimentaires de la fin de l'époque glaciaire, qui marquent le début de la succession marine, s'étendent jusqu' à la Ligne Générale de Rivage Exhaussé de Perth, à environ 29–36 m. A la suite d'une régression relativement importante de la mer, l'érosion marine à l'époque tardiglaciaire produisit une surface d'aplanissement, recouverte de galets et souvent terminée par une falaise. Au cours de l'élaboration de cette surface, il est probable que le niveau de la mer monta par rapport à celui de la terre, et qu'ensuite il s'arrêta à peu près à son niveau actuel ou légèrement au-dessus pendant une longue période pour produire la Ligne de Rivage de Galets Souterrains. L'érosion se poursuivit par endroits au fur et à mesure de l'élévation du niveau de la mer à environ 6 m, ce niveau restant ensuite relativement stable pendant un certain temps. Par la suite les plages souterraines (« Main Buried Beach et Low Buried Beach ») furent formées, avec des hauteurs relevées maximales de ligne côtière dans la région en question de 7,4 et 5,1 m et des dates de formations finales il y a 9500 et 8800 ans respectivement. La régression de la mer jusqu'il y a 8500 ans fut suivie par une transgression au cours de laquelle les dépôts alluvionnaires commencèrent à s'accumuler. On a identifié deux systèmes de vallées souterraines se rapportant aux deux périodes de recul relatif de la mer.

FIG. 1—Matériau de surface terrestre. 1. Tourbe; 2. Dépots sédimentaires pour la partie principale, mais comprenant aussi des remblais et de très petites zones de sédiments alluvionnaires fluviaux; 3. Dépôts sédimentaires (« carse ») de la principale région affectée de la Ligne Post-glaciaire de Rivage Exhaussé; 4. Sable et argile marins de la fin de l'époque glaciaire; 5. Galets déposés par les glaciers à l'époque de la réavancée de Perth; 6. Autres dépôts fluvioglaciaires et argile à blocaux; 7. Assise rocheuse

FIG. 2—Forme de surface rocheuse souterraine. Les hauteurs des ressauts rocheux sont représentées par des contours espacés de 7,6 m. Les hauteurs sont indiquées en mètres (en italiques) et en pieds. Les trous de forage qui atteignent le fond rocheux sont représentés par des points. Les forages qui n'ont pas atteint la roche sont figurés par des croix. Les affleurements rocheux sont illustrés en noir ou par un hachurage. Q et R représentent les points terminaux de la coupe paraissant à la Figure 3.

FIG. 3—Vue en coupe de la tranchée de Carron entre les points Q et R de la Figure 2. Basée sur les données de forages commerciaux. 1. Dépôts surjacents et marne de l'époque tardiglaciaire; 2. Sable et galets; 3. Argile à blocaux; 4. Sable; 5. Assise rocheuse

FIG. 4—Vue en coupe schématique montrant le rapport entre la marne tardiglaciaire supérieure et inférieure. La marne tardiglaciaire est représentée par un hachurage horizontal. Dépôts fluvioglaciaires, argile à blocaux et assise rocheuse représentés par un hachurage oblique.

FIG. 5—Couche de galets souterrains et particularités correspondantes. 1. Assise rocheuse aplanie; 2. Argile à blocaux aplanie; 3. Assiser ocheuse (sous les dépôts sédimentaires de « carse ») d'où la marne tardiglaciaire, les dépôts fluvioglaciaires et l'argile à blocaux ont été érodés (ou n'étaient pas présents à l'origine); 4. Argile à blocaux (sous les dépôts sédimentaires « carse ») d'où les dépôts fluvioglaciaires et la marne tardiglaciaire, ont été érodés (ou n'étaient pas présents à l'origine); 5. Hauteurs en mètres de la surface de galets souterrains ou hauteurs de l'assise rocheuse telles qu'elles ont été mesurées par les forages manuels (nivelés avec précision); 6. Hauteurs en mètres de la surface de galets souterrains, de la surface d'argile à blocaux ou de l'assise rocheuse, telles qu'elles ont été relevées par des forages commerciaux nivelés avec précision; 7. Comme pour 6 mais avec des hauteurs estimées (Voir le texte); 8. Hauteur en mètres de l'assise rocheuse relevée par forages commerciaux, lorsqu'il n'est pas apparent, d'autres données de la carte, que l'on se réfère à la roche; 9. Hauteur en mètres de la surface de marne tardiglaciaire où la couche de galets souterrains n'est pas présente ou n'a pas été relevée (forages manuels soulignés); 10. Comme pour 9, sauf que les hauteurs sont estimées; 11. Autres forages pertinents, c'est-à-dire par exemple les forages au voisinage de OP permettant de définir avec précision la zone de roche aplanie, forages dans la péninsule des docks de Grangemouth où la couche de galets souterrains n'est pas relevée; 12. Contours espacés de 1 m de la surface de la couche de galets souterrains (Voir le texte pour le mode de construction); 13. Escarpement raide (généralement dans l'assise rocheuse ou l'argile à blocaux) qui dépasse normalement des dépôts sédimentaires « carse » et consiste en grande partie d'une falaise laissée par la mer; 14. Rebord, côté terre, du dépôt sédimentaire « carse » (représenté seulement quand les autres données le permettent). 15. Cassure prononcée de pente au bord, côté terre, de la couche souterraine de galets, de la couche aplanie de l'argile à blocaux ou de roche aplanie, à laquelle le texte se réfère sous les termes de ligne côtiére de galets souterrains, lorsqu'elle est située à peu près à la hauteur de référence cartographique ou un peu au-dessus. Toutes les hauteurs des catégories 5, 6 et 7 ci-dessus dans les zones non hachurées se réfèrent à la surface de la couche souterraine de galets où cette couche recouvre de la marne tardiglaciaire, la surface érodée de la marne étant en général de 0,3 à 1,5 m au-dessous, sauf un très petit nombre de forages où l'on ne peut distinguer la couche de galets des galets fluvio-glaciaires sous-jacents. (Pour la réserve afférente à la région à l'ouest de Letham Moss, voir la notice 2 à la fin du texte). Les lettres indiquent les tracés des coupes des Figures 6 et 7.

FIG. 6—Coupe de la Colline d'Airth à l'autoroute encaissée de Polmont (la Fig. 5 montre S, T, U et V), le rapport entre la couche souterraine de galets et la marne tardiglaciaire, l'argile à blocaux et l'assise rocheuse aplanies. 1. Dépôts sédimentaires « carse »; 2. Marne tardiglaciaire; 3. Sable; 4. Galets; 5. Argile à blocaux; 6. Assise rocheuse. Basé principalement sur des forages commerciaux; les forages manuels sont indiqués par la lettre H. Une épaisse couche de galets entre U et V ressort à l'endroit où le ruisseau Grange entre dans le dépôt sédimentaire « carse ». Le terrain au nord de Carron est légèrement abaissé par des affaissements miniers.

FIG. 7—Coupes montrant le rapport entre la couche souterraine de galets, la marne tardiglaciaire et l'assise rocheuse aplanies, et les dépôts côtiers souterrains. La position des coupes est indiquée à la Figure 5. 1. Dépôts sédimentaires « carse »; 2. Tourbe; 3. Dépôts côtiers souterrains; 4. Galets; 5. Marne tardiglaciaire; 6. Assise rocheuse. On n'est pas certain s'il existe des dépôts superficiels sous la couche souterraine de galets à la coupe OP. Ces relevés sont basés principalement sur des forages manuels. Les forages commerciaux sont indiqués par la lettre C. La coupe OP est affectée par de légers affaissements miniers.

FIG. 8—Plages en terrasses souterraines à l'ouest de la Colline d'Airth. Fondé presque entièrement sur des forages manuels. 1. Plage en terrasse souterraine principale; 2. Plages en terrasses souterraines inférieure et principale lorsqu'on ne peut pas les distinguer l'une de l'autre avec les données disponibles; 3. Plage en terrasse souterraine inférieure; 4. Hauteur estimée en mètres de la surface des dépôts de plage souterraine (forages commerciaux); 5. Hauteurs en mètres de la surface des dépôts de plage souterraine, telles qu'elles ont été relevées par des forages manuels précis; 6. Comme pour 5, sauf que la tourbe souterraine est trop dure pour son forage, la hauteur de la surface de la plage souterraine étant estimée en se basant sur l'épaisseur moyenne de la couche souterraine de tourbe enregistrée dans le voisinage; 7. Les dépôts côtiers souterrains sont absents des forages effectués à plus grandes profondeurs que leur plage de niveaux normale; 8. Les dépôts côtiers sont absents du fait que la surface préexistante dépasse leur limite de niveaux supérieure; 9. Les dépots côtiers souterrains sont absents et la surface de la marne tardiglaciaire est à peu près à la même hauteur que la surface de ces dépôts dans le voisinage; 10. Contour espacé de 9 m de la surface terrestre (maintenant enterrée) quand les dépôts côtiers souterrains s'accumulèrent; 11. Contours espacés de 15 et 30 m de la surface terrestre actuelle. AB et CD sont des lignes de coupes comprenant la Figure 9. E, F et G sont des zones où l'on n'a pas examiné les dépôts côtiers enterrés.

FIG. 9—Vues en coupe de vallées souterraines selon les lignes AB et CD de la Figure 8. Relevés uniquement fondés sur des forages manuels. 1. Sable; 2. Dépôts sédimentaires « carse »; 3. Tourbe; 4. Dépôts côtiers de plaque en terrasse souterraine; 5. Galets; 6. Marne tardiglaciaire. Un affaissement minier ne dépassant pas une profondeur de 1,2 m a affecté la coupe AB entre les niveaux 0 et 300 m.

ZUSAMMENFASSUNG—*Drift-Stratigraphie und verdeckte morphologische Charakteristika im Raum Grangemouth–Falkirk-Airth (Mittelschottland)*. Die Quartärablagerungen und verdeckten morphologischen Charakteristika eines kleinen Gebiets im mittleren Forthtal werden an Hand der durch etwa 1900 Bohrlöcher gewonnenen Befunde beschrieben und erörtert. Die verdeckte Grundgestein-Oberfläche wird auf der Höhenlinienkarte dargestellt, und ein tiefer Graben, der das Gebiet durchzieht, wird auf Schmelzwasser und Eiserosion zurückgeführt. Dicke spätglaziale Sedimente, mit denen die Meeres-Schichtenfolge beginnt, erstrecken sich bis zur Haupt Perth-Küstenlinie bei etwa 29–36 m. Im Anschluss an starkes relatives Abfallen der Meereshöhe trat in spätglazialer Zeit erhebliche Meereserosion auf, so dass eine Einebnungsfläche entstand, die mit einer Kiesschicht bedeckt war und häufig von einer Klippe gestützt wurde. Im Verlauf der Entstehung dieser Fläche stieg das Meer wahrscheinlich im Verhältnis zum Land und blieb dann anschliessend geraume Zeit etwa in jetziger oder etwas grösserer Höhe, so dass die verdeckte Kies-Küstenlinie entstand. Die Erosion ging stellenweise weiter, während die Meereshöhe zunächst auf etwa sechs Meter anstieg und dann eine Zeitlang relativ gleich blieb. Danach bildete sich der verdeckte Haupt- und Niederstrand, wobei die grössten festgestellten Küstenlinien-Höhen in dem betreffenden Gebiet 7,4 und 5,1 m betragen, während das Ende der Entstehungszeit bei etwa 9500 BP beziehungsweise 8800 BP liegt. Der Regression des Meeres bis etwa 8500 BP folgte eine Transgression, während der sich die Aufschwemmungen anzusammeln begannen. Zwei verdeckte Talformationen entsprechend den beiden Perioden relativ niedriger Meereshöhe wurden identifiziert.

ABB. 1—Bodenoberflächenmaterie. 1. Torf; 2. Hauptsächlich Aufschwemmungen ('carse'), aber auch aufgetragener Boden und sehr kleine Flächen mit Flussablagerungen; 3. Aufschwemmungen ('carse') im Hauptauftretensbereich des Hauptteils des nachglazialen erhöhten Strandes; 4. Spätglazialer Meeres-Sand und -Lehm; 5. Kiessand vom Wiedervordringen des Perth; 6. Andere fluvioglaziale Ablagerungen und Grundmoräne; 7. Grundgestein

ABB. 2—Form verdeckter Gesteinsoberfläche. Höhen der Gesteinsspitze dargestellt durch Höhenlinien im Abstand von 7,6 m. Höhenwerte sind in Meter (kursiv) und Fuss angegeben. Bohrungen bis zur Gesteinsspitze sind durch Punkte dargestellt. Bohrungen, die nicht bis zur Gesteinsspitze gelangten, sind durch Kreuze kenntlich gemacht. Gesteins-Ausbiss ist schwarz oder schraffiert dargestellt. Q und R sind die Endpunkte des in Abbildung 3 dargestellten Abschnitts.

ABB. 3—Abschnitt quer zum Carron-Graben zwischen Punkt Q und R der Abbildung 2. Bohrwerte von einer Industriefirma. 1. Spätglazialer Lehm und übergelagerte Ablagerungen; 2. Sand und Kies; 3. Grundmoräne; 4. Sand; 5. Grundgestein

ABB. 4—Abschnitts-Graphik, zeigt das Verhältnis zwischen höherem und niedrigerem spätglazialem Lehm. Spätglazialer Lehm dargestellt durch gestrichelte horizontale Schraffierung. Fluvioglaziale Ablagerungen, Grundmoräne und Grundgestein sind durch Schrägschraffierung kenntlich gemacht.

ABB. 5—Die verdeckte Kiesschicht und damit zusammenhängende Charakteristika. 1. Eingeebnetes Grundgestein; 2. Eingeebnete Grundmoräne; 3. Grundgestein (unter den Aufschwemmungen), aus dem Grundmoräne, fluvioglaziale Ablagerungen und spätglazialer Lehm ausgewaschen worden sind (bzw. sie waren von Anfang an nicht vorhanden); 4. Grundmoräne (unter den Aufschwemmungen 'carse'), aus der fluvioglaziale Ablagerungen und spätglazialer Lehm ausgewaschen worden sind (bzw. sie waren von Anfang an nicht vorhanden); 5. Höhe (in m) der Oberfläche der verdeckten Kiesschicht oder Höhe der Gesteinsspitze: gemessen durch Handbohrungen (genau nivelliert); 6. Höhe (in m) der Oberfläche der verdeckten Kiesschicht, der Grundmoränen-Oberfläche oder der Gesteinsspitze: gemessen in genau nivellierten Industriebohrlöchern; 7. Wie 6, ausser dass die Höhenwerte geschätzt sind (siehe Text); 8. Höhe (in m) der Gesteinsspitze, gemessen in Industriebohrungen, wo aus anderen kartographischen Unterlagen nicht hervorgeht, dass von Gestein die Rede ist; 9. Höhe (in m) der Oberfläche des spätglazialen Lehms, wo verdeckte Kiesschicht nicht vorhanden bzw. nicht festgestellt (Handbohrungen unterstrichen); 10. Wie 9, ausser dass Höhenwerte geschätzt sind; 11. Andere wichtige Bohrungen, z.B. Bohrungen in der Umgebung von OP, durch die die Fläche des eingeebneten Gesteins genau bestimmt werden kann und Bohrungen in der Grangemouth–Hafen–Halbinsel, wo keine Unterlagen über die verdeckte Kiesschicht vorliegen; 12. Höhenlinien (im Abstand von einem Meter) der Oberfläche der verdeckten Kiesschicht (Baumethode: siehe Text); 13. Steilufer (gewöhnlich in Grundgestein oder Grundmoräne), das normalerweise von unterhalb der Aufschwemmungen aufsteigt; zum grössten Teil eine zurückgelassene Meeresklippe; 14. Landwärts gelegener Rand der Aufschwemmung 'carse' (nur dort gezeigt, wo es mithilfe anderer Unterlagen möglich ist); 15. Deutlicher Bruch der Senke am landwärts gelegenen Rand der verdeckten Kiesschicht, der eingeebneten Grundmoräne oder des eingeebneten Gesteins; im Text als Verdeckt-Kies-Küstenlinie bezeichnet, und zwar etwa in mittlerer Höhe über Normalnull oder geringfügig darüber. Anmerkung: Alle Höhenwerte unter 5, 6 und 7 in unschraffierten Flächen beziehen sich auf die Oberfläche der verdeckten Kiesschicht, wo diese Schicht spätglazialen Lehm überlagert; die ausgewaschene Oberfläche des Lehms ist in der Regel zwischen 0,3 und 1,5 m niedriger, ausgenommen einige wenige Bohrungen, in denen die Kiesschicht nicht von dem darunter liegenden fluvioglazialen Kies unterschieden werden kann (Einschränkung in Bezug auf das Gebiet westlich von Letham Moss: siehe Anmerkung 2 am Ende des Textes). Die Buchstaben bezeichnen die Abschnittslinien in Abb. 6 und 7.

ABB. 6—Abschnitt von 'Hill of Airth' zum Autobahn-Durchstich bei Polmont (S, T, U und V in Abb. 5): zeigt das Verhältnis der verdeckten Kiesschicht zum eingeebneten spätglazialen Lehm, zur eingeebneten Grundmoräne und zum eingeebneten Grundgestein. 1. Aufschwemmungen (carse); 2. Spätglazialer Lehm; 3. Sand; 4. Kies; 5. Grundmoräne; 6. Grundgestein. Hauptsächlich Industrie-Bohrungen. Handbohrungen sind durch den Buchstaben H gekennzeichnet. Dicker Kies zwischen U und V tritt auf, wo der Bach Grange in die Aufschwemmung eintritt. Boden nördlich von Carron durch Bergbau-Bodensenkung leicht abgesunken.

ABB. 7—Abschnitte zeigen Verhältnis von verdeckter Kiesschicht, eingeebnetem spätglazialem Lehm und eingeebnetem Grundgestein zu verdeckten Strandablagerungen. Lage der Abschnitte geht aus Abbildung 5 hervor. 1. Aufschwemmungen (carse); 2. Torf; 3. Verdeckte Strandablagerungen; 4. Kies; 5. Spätglazialer Lehm; 6. Grundgestein. Oberflächliche Ablagerungen unterhalb der verdeckten Kiesschicht in Abschnitt OP ungewiss. Hauptsächlich Handbohrungen. Industriebohrungen durch Buchstaben C gekennzeichnet. Abschnitt OP durch leichte Bergbau-Bodensenkung beeinflusst.

ABB. 8—Verdeckte erhöhte Strände westlich von 'Hill of Airth'. Fast ausschliesslich Handbohrungen. 1. Hauptteil des verdeckten erhöhten Strands; 2. Haupt- und niedriger Teil des verdeckten erhöhten Strands, wo die beiden aufgrund der vorhandenen Unterlagen nicht unterscheidbar sind; 3. Niedriger Teil des verdeckten erhöhten Strands; 4. Geschätzte Höhenwerte (in m) der Oberfläche der verdeckten Strandablagerungen (Industrie-Bohrungen); 5. Höhe (in m) der Oberfläche der verdeckten Strandablagerungen: gemessen in genau nivellierten Handbohrlöchern; 6. Wie 5, aber verdeckter Torf zu hart zum Bohren; Höhe der verdeckten Strandoberfläche wurde geschätzt auf der Grundlage der gemessenen Durchschnittswerte für Torfdicke in der Umgebung; 7. Verdeckte Strandablagerungen nicht vorhanden in Bohrlöchern, die tiefer vorangetrieben wurden als die normale Tiefe, in der diese Ablagerungen auftreten; 8. Verdeckte Strandablagerungen nicht vorhanden, weil vorher vorhandene Oberfläche oberhalb ihrer oberen Höhengrenze; 9. Verdeckte Strandablagerungen nicht vorhanden und Oberfläche des spätglazialen Lehms etwa in gleicher Höhe wie Oberfläche dieser Ablagerungen in der Umgebung; 10. 9 m-Höhenlinie der Landoberfläche (jetzt verdeckt), als sich Strandablagerungen ansammelten; 11. Höhenlinien 15 und

30 m der jetzigen Landoberfläche. AB und CD kennzeichnen Linien der Abschnitte in Abb. 9. E, F und G sind Gebiete, in denen verdeckter Strand nicht untersucht wurde.

ABB. 9—Abschnitte quer zu verdeckten Tälern auf Linien AB und CD in Abb. 8. Ausschliesslich Handbohrungen. 1. Sand; 2. Aufschwemmungen (carse); 3. Torf; 4. Verdeckte Ablagerungen auf erhöhtem Strand; 5. Kies; 6. Spätglazialer Lehm. Bergbau-Bodensenkung bis 1,2 m im Abschnitt AB zwischen 0 und 300 m.

Isobases for the Main Perth Raised Shoreline in South-East Scotland as determined by Trend-Surface Analysis

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THE MOST conspicuous Late-glacial raised shoreline in the vicinity of the western part of the Firth of Forth is the Main Perth Shoreline (J. B. Sissons and D. E. Smith, 1965). It therefore seems reasonable to infer that this shoreline will be prominently developed in some other parts of south-east Scotland. The purpose of this paper is to show how morphological evidence and trend-surface analysis support this inference and also to consider the pattern of isostatic warping in south-east Scotland as revealed by the latter.

All altitudes used in the trend-surface analysis were expressed to one-tenth of a foot, the results being subsequently metricized. They were obtained by accurate levelling based on Ordnance Survey bench marks and refer approximately to the former shoreline. Locations of the levelled points were expressed to the nearest 10 m in terms of the National Grid, modified to give consecutive numbers. Columbia University's IBM 7094 computer was used to compute the surfaces. Three programs were employed: E. H. T. Whitten's (1963) Fortran II program, G. P. H. Styan and P. C. Trenholme's (1967) Fortran IV program, and a program in Fortran IV written by P. Kahn to produce maps of the surfaces computed.

The Western Areas

Between Burntisland and Kincardine on the northern side of the Forth and between South Queensferry and Plean on the southern side, the Main Perth Raised Beach is frequently a prominent feature, many of the individual remnants exceeding 1 km in length. Westward the beach merges into outwash deposits near Kincardine, Larbert and Plean, and these deposits are believed to have been formed at or near the limit of the Perth Readvance. Two lower raised beaches formed during the period of ice decay that followed the readvance have also been identified. The raised beaches have been described and discussed in detail elsewhere (Sissons and Smith, 1965) and the gradient of the Main Perth Shoreline calculated separately for the north and south sides of the Forth by projecting the shoreline altitudes into vertical

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planes aligned in various directions and applying linear regression analysis (see also Sissons, 1967a). No attempt was then made to consider both sides of the Forth in one calculation. A linear trend surface for both sets of data has now been computed, the results being given in Table I and Figure 1. The gradient of 0.44 m/km towards E.16°S. obtained for this surface

TABLE I

Statistical Data for Linear Surfaces illustrated in Figure 1

Area	A	B	C	D	E
Number of height values	208	170	74	18	44
Percentage sum of squares accounted for by surface	98.46	98.26	93.34	91.25	99.06
Direction of slope	E.16°S.	E.12°S.	E.14°S.	E.71°S.	E.34°S.
Slope, m/km	0.44	0.51	0.35	1.25	0.42

may be compared with the gradients previously obtained by linear regression analysis, for example, for the north side, 0.41 m/km when the measured points were projected into a vertical plane aligned W.16°N.—E.16°S. and, for the south side, 0.46 m/km in a plane aligned W.12°N.—E.12°S.

Along the coast of the western part of the Firth of Tay and in the adjacent lower Tay and Earn valleys, there occurs an extremely well-defined series of Late-glacial raised beaches. As in the Forth valley, outwash spreads descend to the level of some of the beaches at their western ends and merge with them. The highest shoreline in this series has a gradient not greatly dissimilar from that of the highest shoreline in the western part of the Forth valley. A gradient of 0.49 m/km in a vertical plane of projection aligned W.15°N.—E.15°S. has been calculated by linear regression analysis for ninety-eight points on the northern side of the Earn-Tay, and a gradient of 0.56 m/km in the same plane has been obtained for seventy-seven points on the southern side. These similarities of morphology and gradient suggest that the highest shoreline in the Earn-Tay area is equivalent to the highest shoreline (that is, the Main Perth) in the western part of the Forth area. This appears to be confirmed by linear trend-surface analysis. The computed surface for the Earn-Tay area, sloping towards E.12°S. at 0.51 m/km, does not differ greatly from that for the western Forth area. The difference in gradient of the respective linear surfaces of 0.07 m/km is not attributable to difficulties in field measurement, for the raised shoreline is a very clear feature and has been measured at a large number of points along both flanks of both the Forth and Earn-Tay valleys. The difference may reflect the different conditions of Late-glacial sedimentation in the two estuaries as influenced by such factors as rate of sediment supply, estuarine configuration and exposure. It may in part also reflect the respective positions of the two areas in relation to the elliptical dome of isostatic uplift (p. 42).

The Eastern Areas

For considerable distances eastward from the two areas considered so far, the Main Perth Shoreline is either missing or cannot be measured accurately. This results partly from the presence of built-up areas, partly from the occurrence of steep rock slopes descending to the present shoreline, and partly from subsequent erosion enhanced by greater exposure to

wave attack. It is, therefore, not possible to trace the Main Perth Shoreline from the western areas into the eastern part of the Forth-Tay region, and it thus becomes necessary to try to determine which, if any, of the raised shoreline remnants in the latter region (all identifiable shorelines have been mapped and levelled) are to be equated with the Main Perth Shoreline.

The identification of this shoreline in the eastern areas is facilitated by the existence of upper and lower limits imposed by earlier and later shorelines, as is evident from the altitude-distance diagram published elsewhere (Sissons *et al.*, 1966; Sissons, 1967b). A lower limit to the altitude range in which a visible (as opposed to a buried) Main Perth Shoreline could be identified is imposed by the highest Post-glacial shoreline, for which representative altitudes are 5.8–6.1 m near Dunbar, 9–10 m in Edinburgh, 7.6–7.9 m near St. Andrews and 8.5–8.8 m in Dundee. An upper limit is imposed by the more steeply sloping shorelines formed earlier than the Main Perth. Thus in eastern Fife the lowest of six earlier shorelines identified by R. A. Cullingford and Smith (1966) descends from about 25 m at Largo and about 22 m near St. Andrews to about 14 m at Fife Ness. The upper limit for the Main Perth Shoreline is further restricted in these eastern areas by the altitudes of kettle holes (for example, 17.4 m north of St. Andrews, 16.8 m near Largo and 13.4 m near Dunbar), unless one assumes that the melting of the glacier ice that produced these holes was unduly protracted.

Within this altitudinal zone on the south side of the Firth of Forth there is a well-marked raised beach along considerable parts of the coastal zone in the Cockenzie to Aberlady area, the longest individual remnant extending for 2 km. The former shoreline, measured at seventy-four points, declines gently in altitude from 11.9 m inland of Cockenzie to 9.1 m about 1.6 km east of Aberlady. The altitudes obtained for these two end locations by continuing the surface computed for area A in Figure 1 are 11.0 and 8.2 m respectively, while the average discrepancy between the seventy-four points in area C and the linear surface extrapolated from area A is +0.9 m.

On the opposite side of the Firth of Forth from Largo to immediately east of Elie, short stretches of raised shoreline between altitudes of 9 and 15 m have been measured. The eighteen altitudes obtained include eight on the second lowest of the rock platforms at Kincraig Point. Some of the remaining ten shoreline altitudes obtained in the area are probably influenced by blown sand, which is locally very abundant.

On the eastern coast of Fife, in the altitudinal zone between the series of Late-glacial shorelines referred to above and the highest Post-glacial shoreline, steep coastal slopes towards the south restrict the area in which the Main Perth Shoreline can occur as a measurable feature. North of the Eden estuary, however, there is a well-marked raised beach, the shoreline of which, as measured at thirty-three points over a distance of 5 km is between 15.5 and 17.0 m. In the higher raised beach deposits that occur inland of this shoreline, several kettle holes occur, the lowest point on the rim of the largest one having been levelled by J. I. Chisholm (1966) as 17.4 m. The altitude of the well-marked raised shoreline in the immediate vicinity is 15.5–16.4 m. As Chisholm argues (pp. 171–2), the occurrence of kettle holes in the higher raised-beach deposits implies that glacier ice still occupied these holes when these deposits accumulated. The kettle holes have a greater significance, however, for they do not occur below the 15.5–17.0 m shoreline. This sudden disappearance of kettles is difficult to explain if the latter shoreline is interpreted simply as a member of a series of successively lower shorelines formed during the deglaciation of the local area. On the other hand, if the 15.5–17.0 m feature is interpreted as the Main Perth Shoreline and marks the limit of a much later transgression, the absence of kettles below the shoreline is readily explicable.

Farther along the Fife coast, south-east of St. Andrews, only two measurable features that can be attributed to the Main Perth Shoreline have been identified. Each forms a small but clear terrace standing distinctly above the upper limit of the Post-glacial raised beaches, the shoreline of one having been levelled at five points and of the other at six.

In Figure 1, linear surfaces calculated separately for each of the three eastern areas are shown. The number of measured points for two of these areas is small, while all three suffer from the disadvantage, in relation to trend-surface analysis, of comprising a single length of coastline (in contrast to the much better control provided by the double lines of altitudes in each of areas A and B). Furthermore, it seems probable that, owing to the small size of these areas, local factors affecting beach formation may have tended to mask the regional trend, as

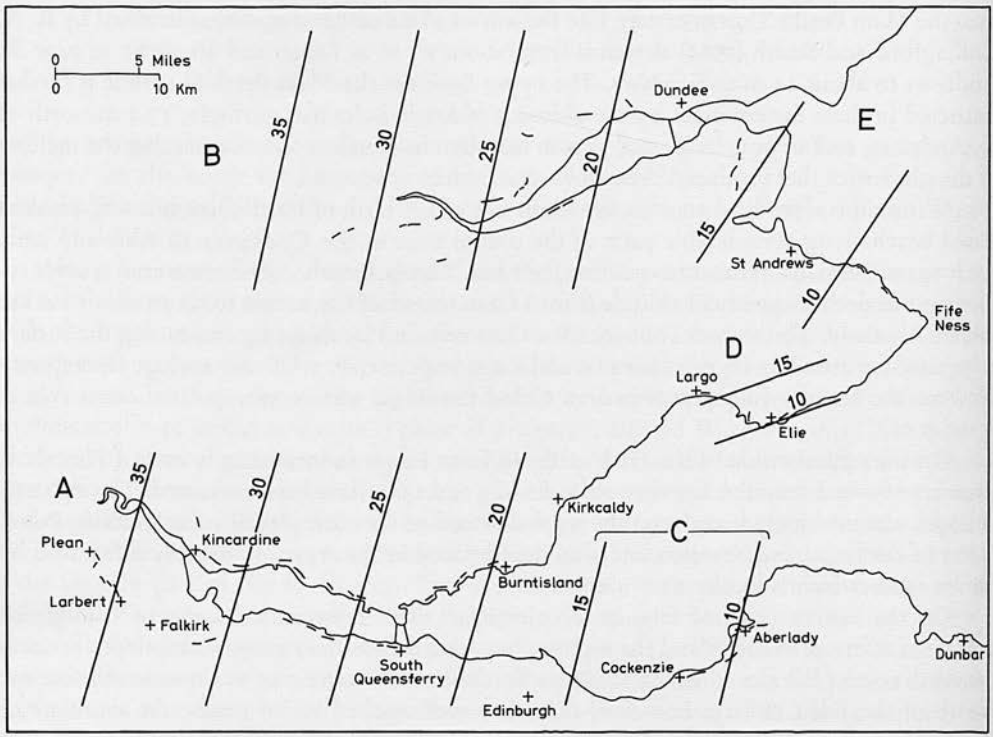


FIGURE 1—Linear trend-surfaces for the Main Perth Raised Shoreline in areas A to E (see text). Altitudes in metres. Locations where the shoreline has been measured are indicated.

was found even in area A when short sections of shoreline were projected into a vertical plane and compared through linear regression analyses (Sissons and Smith, 1965, p. 161). As is evident from Figure 1, the linear surface calculated for area D is markedly anomalous, a result probably related particularly to the additional complications introduced by blown sand. On the other hand, the surfaces for C and E do not appear unduly divergent from those for the two western areas in view of the limitations indicated above. The directions of slope for C and E differ by about 2 and 22 degrees respectively from that for the two western areas, and the computed gradients differ from those for the western areas by 0.09 m/km in each case. Trend-surface analysis thus does not appear to contradict the conclusion reached above on

morphological grounds that the shorelines referred to in areas C and E are part of the Main Perth Shoreline.

The Area as a Whole

As a step towards trend-surface analysis of all the data on the Main Perth Shoreline for the Forth-Tay region, linear surfaces were first computed for area G, comprising the Earn-Tay

TABLE II

Statistical Data for Linear Surfaces illustrated in Figure 2

Area	F	G
Number of height values	300	214
Percentage sum of squares accounted for by surface	99.59	98.82
Direction of slope	E.13°S.	E.7°S.
Slope, m/km	0.42	0.43

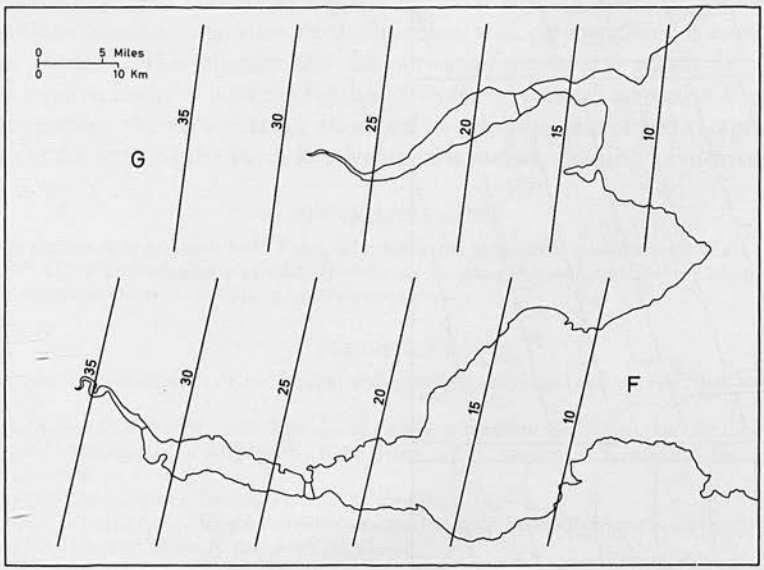


FIGURE 2—Linear trend-surfaces for the Main Perth Raised Shoreline in areas F (Forth) and G (Earn-Tay with eastern Fife). Altitudes in metres.

area with eastern Fife (areas B and E), and for area F, comprising the areas in the Forth valley (A, C and D). Area D was included in the latter computation since it was felt that its omission would introduce subjectivity. The results of the computations are shown in Table II and Figure 2. Although the surfaces depicted in Figure 2 represent a second stage in the generalization of the data, the isobases vary little from those in Figure 1 at the actual locations where the shoreline has been measured in the field (except in area D).

The surface computed for area G is slightly steeper than that for area F (0.43 as opposed to 0.42 m/km) and also slopes rather more towards the east. While these results may in part reflect limitations in the field evidence in the eastern areas, they are in accord with what is known of the form of isostatic uplift in Scotland. The dome of isostatic uplift has long been assumed to be approximately elliptical with its major axis lying roughly north and south (W. B. Wright, 1937). Areas about the minor axis of the ellipse should therefore have the steepest shoreline gradients, sloping, in eastern Scotland, towards the east. With increasing distance from the minor axis, shoreline gradients should diminish and, in eastern Scotland,

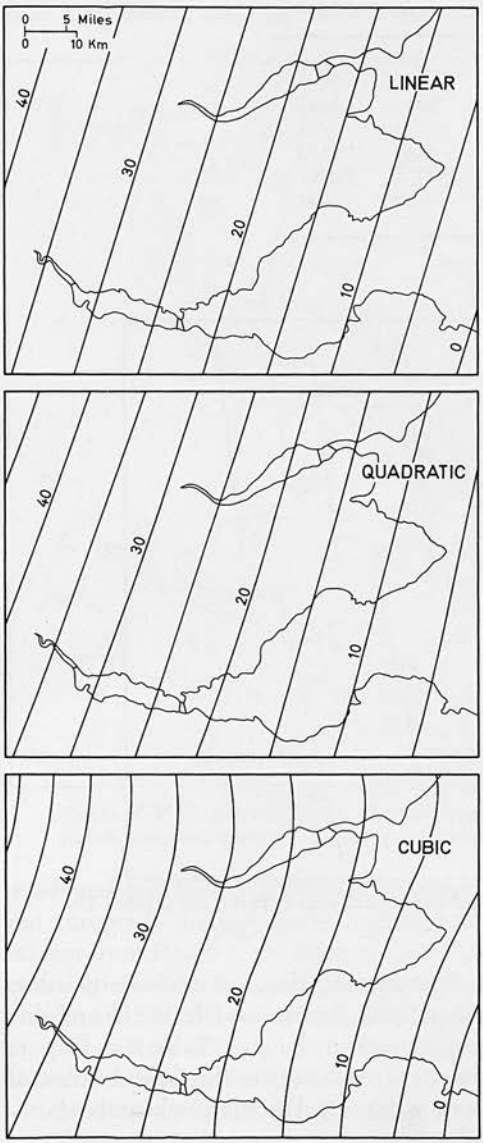


TABLE III

Statistical Data for Surfaces illustrated in Figure 3

Type of surface	Linear	Quadratic	Cubic
Number of height values	514	514	514
Percentage sum of squares accounted for by surface	99.26	99.41	99.55
Direction of slope	E.17°S.	—	—
Slope, m/km	0.43	—	—

FIGURE 3—Linear, quadratic and cubic trend-surfaces for the Main Perth Raised Shoreline in south-east Scotland based on 514 measured points. Altitudes in metres.

the direction of slope should have a diminishing eastward component. In terms of these considerations, the linear trend-surfaces depicted in Figure 2 are consistent with the Forth-Tay area being located a short distance south of the minor axis of the ellipse.

A third stage in the generalization of the data is presented in Figure 3 and Table III. The linear surface is, of course, unable to reveal the variations in direction and amount of slope just discussed. It appears significant, however, that its direction of slope (E.17°S.) is very similar to the direction of slope (E.18°S.) calculated for an older raised shoreline in eastern Fife by combining the results of linear regression analyses (Cullingford and Smith, 1966, p. 45). The quadratic and cubic surfaces indicate an increase in shoreline gradient in the west in accord with the results of linear regression analyses for area A (Sissons and Smith, 1965, p. 161). Comparison of the maps in Figure 3 shows that, at places where the altitude of the shoreline has been measured in the field, the altitudes of the computed surfaces vary only slightly from one another. All the surfaces account for a very high proportion of the sums of squares and all have a very high confidence level (more than 99.95 per cent), although it should be noted that these high values are partly a result of the unavoidably irregular areal distribution of the points measured in the field.

Conclusions

Trend-surface analysis appears to support the view that the shoreline data used in this paper refer to one shoreline, the Main Perth Shoreline, with the exception of some of the data in area D of Figure 1. This suggests that the outwash deposits into which the raised beach merges at its western limits in both the Forth and Tay areas were all formed at about the same time, thus supporting the view that the associated ice-margins, interpreted as representing the limit or almost the limit of the Perth Readvance, are, in fact, essentially synchronous.

ACKNOWLEDGEMENTS

The authors express their gratitude to P. Kahn, who wrote the program to produce maps of the surfaces; to G. P. H. Styan and P. C. Trenholme, who gave valuable advice on the use of their program; and to Columbia University, for financial assistance and the use of facilities at its computer centre.

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Fig. 1—Modèle topographique de la ligne de rivage exhaussé de Perth, dans les zones de A à E (Voir le texte). Les altitudes sont indiquées en mètres. Les points de levée de la ligne de rivage sont repérés.

Fig. 2—Modéle topographique de la ligne de rivage exhaussé de Perth, dans les zones F (Forth) et G (Earn-Tay avec l'est de Fife). Les altitudes sont indiquées en mètres.

FIG. 3—Modèles topographiques linéaires, quadratiques et cubiques de la ligne de rivage exhaussé de Perth, dans le sud-est de l'Ecosse, basés sur 514 points de triangulation. Altitudes en mètres

ABB. 1—Lineare Neigungsfläche der Main Perth Raised Shoreline in den Gebieten A bis E (siehe Text). Höhen in Metern. Orte, an denen die Küstenlinie gemessen wurde, sind angegeben.
ABB. 2—Lineare Neigungsfläche der Main Perth Raised Shoreline in den Gebieten F (Forth) und G (Earn-Tay mit östlichem Fife). Höhen in Metern
ABB. 3—Lineare, quadratische und kubische Neigungsflächen der Main Perth Raised Shoreline in Südostschottland nach 514 vermessenen Punkten. Höhen in Metern

Drift stratigraphy and buried morphology beneath the Forth at Rosyth

BY

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Synopsis

During work under contract to the Institute of Geological Sciences, over 500 old bore records have become available for a small area at Rosyth, the majority relating to the area below low water mark. They show that till and patchy fluvioglacial deposits are covered by fine late-glacial marine sediments whose maximal recorded thickness exceeds 60 m. In the western parts of the area a glacial readvance or halt stage is indicated particularly by ' boulder clay ' interbedded with the late-glacial marine sediments. Sea-level when glacier ice last extended into the Rosyth area was about 35 m above present. Later in late-glacial times the sea fell below its present level and marine sediments deposited earlier were extensively planated, the coarser (ice-rafted) constituents of these sediments and the coarser constituents of the interbedded ' boulder clay ' (where present) being left as a layer of stones and boulders sloping gently seawards. This coarse stratum now lies beneath a layer of silt and shells. The drift sequence in the Rosyth area can be correlated with that at Bo'ness and Grangemouth, enabling the gradient of the shoreline limiting the main period of late-glacial marine erosion to be provisionally calculated as between 0·20 and 0·25 m/km.

Introduction

While working on a feasibility study on data-banking of drift borehole records under contract to the Institute of Geological Sciences the authors had access to more than 500 borehole logs for a small area at Rosyth. Some 300 of these boreholes were put down below low water mark and reveal a very clear picture of the drift stratigraphy beneath this part of the Firth of Forth. The purpose of this paper is to describe and interpret this stratigraphy and to show that the evidence from the Rosyth area can be correlated with evidence already available from areas further west in the Forth valley.

The area for which the most detailed evidence is available lies on the northern side of the Firth of Forth between 1 and 4 km WNW. of the northern end of the Forth Road Bridge. It extends southwards into the Forth to a distance of between 1·2 and 1·6 km below H.W.M. The borehole investigations of this area were carried out during a period of several years before World War I. Since that time the erection of sea walls and other structures, along with extensive dredging, have greatly modified the area. Thus the former high water mark is hundreds of metres north of the present artificial coastline and the significant and widespread buried gravel layer that will be discussed below has been removed over a large area.

Scott. J. Geol. 6 (3), 272–284, 1970.

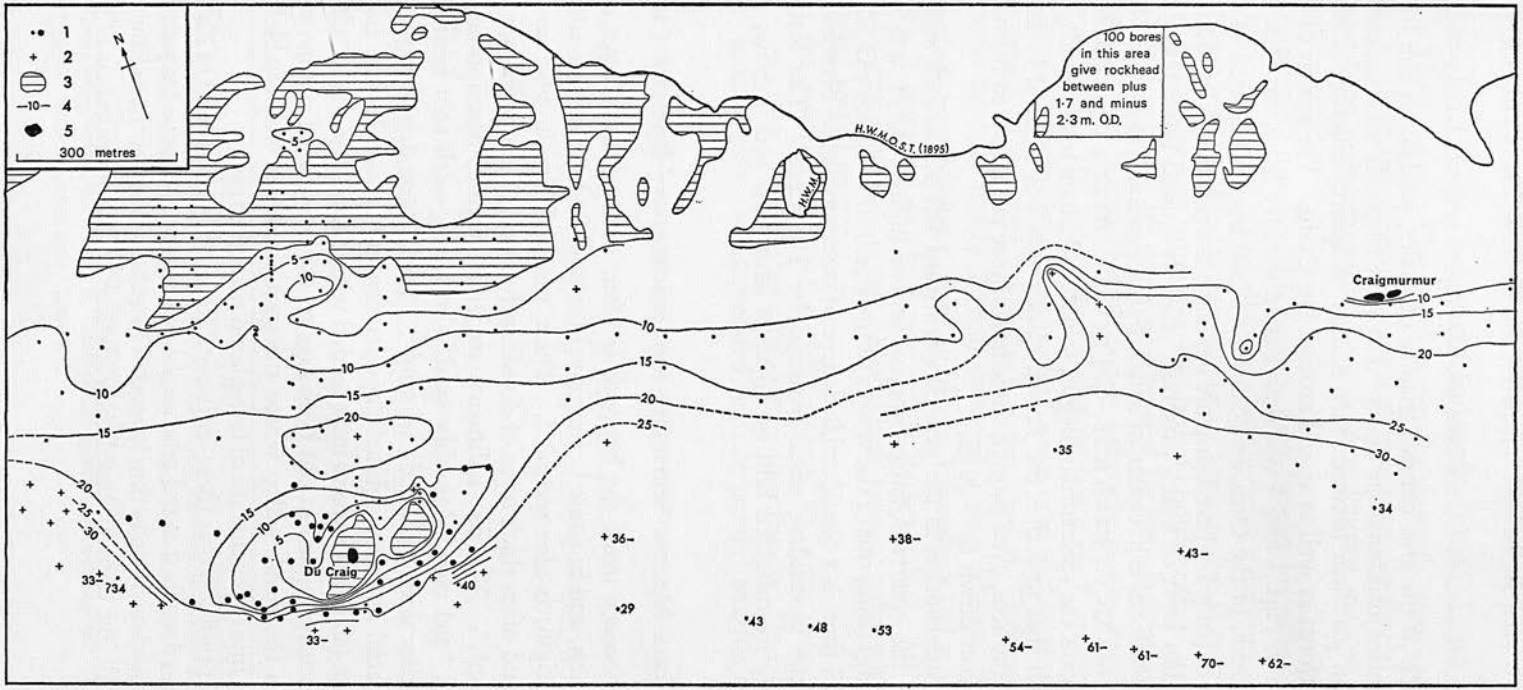


FIG. 1. Rockhead. 1, Bores to rockhead, larger dots indicating bores in which rock is recorded as ' whinstone '. 2, Selected bores not reaching rockhead. 3, Rock platform between tide marks. 4, Rockhead contours at 5 m intervals. 5, Small rock islands rising above H.W.M. Altitudes in metres below O.D.

Bedrock Relief

The bedrock of the area consists mainly of shales, sandstones and limestones of the ' Carboniferous Limestone Series ' that were formerly well exposed on the marine abrasion platform between tide marks. A quartz-dolerite sill also crops out on the platform as well as at and around Du Craig. The extent of the platform as shown in Figs. 1 and 4 was determined initially from the rock drawing on the 1895 edition of the Ordnance Survey 1:2500 map. The area of the platform has been extended where borehole evidence is available to include bedrock covered by a thin layer of beach material. Closely-spaced bores reveal that the surface of the platform is irregular in detail, but in general it slopes gently southwards to L.W.M.O.S.T., which is at $-2{\cdot}1$ m O.D. At or a short distance south of low water mark the rock surface begins to descend much more rapidly towards the deep trench that underlies the Forth. Rock was not reached in the south-eastern part of the area, even though three bores were put down to below -60 m and a fourth went down to -70 m.

The main interruption in the general southward fall of the rock surface corresponds with the quartz-dolerite outcrop whose highest point was formerly a rock (Du Craig) rising out of the waters of the Forth to 7 m above O.D. As can be seen in Fig. 1 there is a considerable degree of correspondence between igneous bedrock and the upstanding mass crowned by Du Craig. The feature thus resembles many ice-moulded hills on adjacent land areas and it shows a similar alignment.

Late-Glacial Marine Sediments and associated ' Boulder Clay '

Below low water mark the bedrock is often covered by a layer of till of variable thickness, and in some bores gravel (or sand and gravel), presumably of fluvioglacial origin, is also recorded. These materials are in turn covered by a widespread and often thick layer of dominantly fine sediments that are often referred to as ' soft '. The fine sediments are given various descriptions, such as ' puddle clay ', ' red clay ', ' muddy sand ', ' red clay with thin beds of sand ' and ' puddle clay with thin beds of sand '. Since these deposits immediately succeed the glacial and fluvioglacial deposits almost throughout the area below low water mark (and are in places interbedded with them), and since such descriptions are frequently applied in old bore logs to the late-glacial marine sediments farther west in the Forth valley where the stratigraphy is well established, it is clear that these fine sediments are of late-glacial marine origin.

The records indicate that these sediments are mainly composed of clay, which is almost always described as red when a colour is given. Sand is frequently interbedded with the clay, often as thin layers. The subdivisions of the sediments in the bore logs are almost entirely based on variations in the proportions of sand and clay.

The late-glacial marine sediments are not identifiable above an altitude of between −3 and −4 m. They thicken rapidly seawards as rockhead falls away, and in the south of the area investigated they are usually several tens of metres thick. One bore passed through 60 m of these deposits without reaching their base.

In a considerable number of bore logs ' boulder clay ' is recorded interbedded with (or, occasionally, overlying) the late-glacial marine sediments. The ' boulder-clay ' occurs in beds recorded as varying from 15 cm to several metres in thickness. In many of these bore logs more than one bed of ' boulder clay ' is mentioned; in two instances five layers are reported separated by late-glacial marine sediments. The ' boulder clay ' is usually described as soft, sometimes as medium, and occasionally as hard. The most detailed description of such material is available from the section exposed when a test cylinder was sunk in the extreme west of the area investigated. One layer, 1·4 m thick, is described as ' Dark purple clay, containing stones up to 9 inches (23 cm) diameter ', thus differing in colour from the red clay above and below. A second layer, 1·2 m thick and at a lower level in the section, is described as ' Stiff sandy clay, darker colour, with stones up to 12 inches (30 cm) diameter, nearly all of which were angular with flat polished faces showing glacial scratches '.

When the anomalous bores (i.e. those referred to in the previous paragraph) are plotted on a map a remarkably clear distribution appears. In Fig. 2 every bore in which ' boulder clay ' is interbedded with or (rarely) overlies the late-glacial

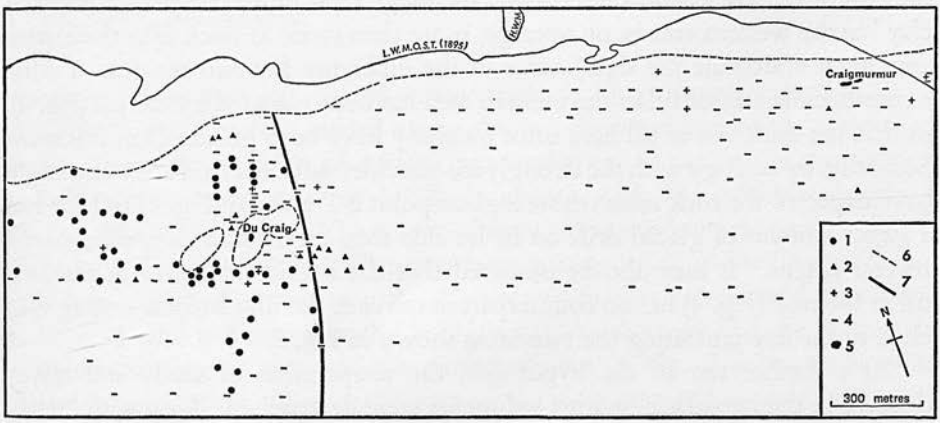


FIG. 2. Deposits associated with the late-glacial marine sediments. 1, Bores in which ' boulder clay ' is recorded as interbedded with (or, occasionally, overlying) late-glacial marine sediments. 2, Late-glacial marine sediments overlying ' boulder clay ' which in turn overlies ' sand ' or ' sand and gravel '. 3, ' Sandy clay and stones ', ' clay with stones ', etc. interbedded with late-glacial marine sediments. 4, Other bores in which late-glacial marine sediments were encountered. 5, Small rock islands rising above H.W.M. 6, Limit of late-glacial marine deposits. 7, Line separating ' western area ' from ' eastern area ' (see text).

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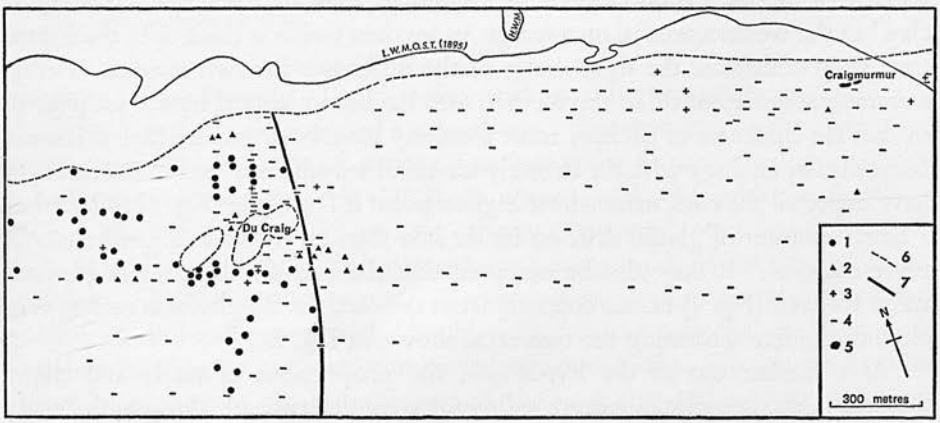


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marine sediments is plotted, excepting only where the term ' boulder clay ' has clearly been used to describe the buried gravel layer (p. 278). All the anomalous bores lie in the western part of the area and a line has been drawn on Fig. 2 to separate this ' western area ' from the 'eastern area '. In the western area 46 per cent of the bore logs record the anomalous ' boulder clay ' (56 out of 122), whereas in the eastern area it is not referred to at all in 116 bore logs.

Since glacial till is not always called ' boulder clay ' in bore logs, but is sometimes given descriptions such as ' clay with stones ', ' sandy clay and stones ', etc., bores in which such deposits are recorded in association with the late-glacial marine sediments are also plotted in Fig. 2. In addition bores in which ' boulder clay ' is recorded as resting on ' sand ' or ' sand and gravel ' are indicated. Of the 29 bores in these two categories, 22 lie in the western area or a short distance east of the line separating the two areas. It is suggested that the only explanation that can account for the distributions shown in Fig. 2 is deposition of the ' boulder clay ' from glacier ice at a time when the margin of grounded ice was approximately stationary or oscillating in the vicinity of the thick line marked in Fig. 2, or during a readvance to the vicinity of the line.

If this concept is valid one might expect glacial deposits to be of greater thickness in the western area than in the eastern. To test this expectation the thickness of all material recorded as ' boulder clay ' in all bores reaching rockhead and situated below L.W.M.O.S.T. was determined separately for the western and eastern areas as defined in Fig. 2. In the former area 99 bores reached rockhead, the mean thickness of ' boulder clay ' being 3·1 m. The comparable figures for the eastern area are 93 bores and 1·4 m. Although these figures show that ' boulder clay ' in the western area is, on average, more than twice as thick as in the eastern area, they understate the significance of the difference for two reasons. Firstly, a considerable area of till in the western area has been planated by the sea (Fig. 4), so that the thickness of till here must formerly have been greater than it is now. Secondly, by analogy with the strongly ice-moulded adjacent land areas one might have expected the rock mass whose highest point is Du Craig (Fig. 1) to have had a larger amount of glacial drift on its lee side than on its stoss side, whereas the reverse is true. It may also be observed that the considerable area of planated till in the west (Fig. 4) has no counterpart in the east, the till platform ending very close to the line separating the two areas shown in Fig. 2.

As a further test of the hypothesis, the proportions of sandy and clayey deposits in the late-glacial marine sediments were determined, along with ' sand ' or ' sand and gravel ' that is probably of fluvioglacial origin, the assumption being that, if a halt in ice decay or a readvance occurred in the western area, there should be a higher proportion of fluvioglacial material and of sandy marine sediments in this area as compared with the eastern area owing to the longer time during which glacial rivers would have discharged. For each bore in which fluvioglacial and late-glacial marine deposits together exceeded 6 m in thickness, the thickness

of sand (including ' clayey sand ', etc.) and gravel as a percentage of the total thickness of late-glacial marine and fluvioglacial deposits was calculated. When these percentages were plotted on a map, high values (30–100 per cent) were found to be common in the west (especially south and west of Du Craig) and low values in the east, almost all bore logs in a large area in the north-east recording neither sand nor gravel.

In a few bores in the west, however, no sand or gravel are recorded, while closely-spaced bores in the west often show considerable variations in the per-

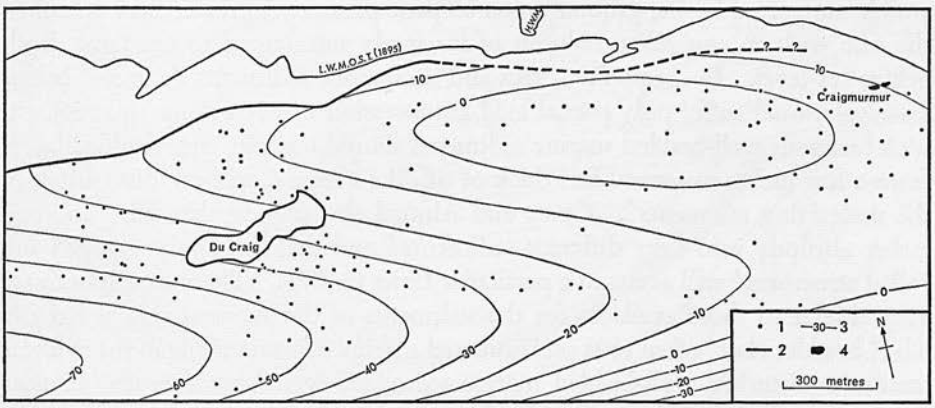


FIG. 3. Cubic trend–surface of thickness of sand (including ' clayey sand ', etc.) and gravel as a percentage of total thickness of late-glacial marine and fluvioglacial deposits. 1, Bores used in analysis. 2, Limit of area to which data apply. 3, Isolines at 10 per cent intervals. 4, Small rock islands rising above H.W.M.

centages of these deposits. Trend-surface analysis, using the program of O'Leary *et al.* (1966), was therefore carried out to obtain a clearer picture of the percentage distribution. The linear trend–surface (accounting for 43 per cent of the sums of squares) slopes down towards NNE. in accord with the evidence described above. The cubic surface (52 per cent of sums of squares), reproduced in Fig. 3, shows in addition a belt of high percentages passing south of Du Craig and declining in value from west to east. Considered in conjunction with the interbedding of sand and gravel with ' boulder clay ' in the west and the absence of such interbedding in the east, the pattern shown in Fig. 3 is consistent with glacier ice occupying the western part of the map area and rivers discharging eastwards from it, especially south of Du Craig.

Immediately south of the Firth of Forth, in and near the grounds of Hopetoun House, in a belt extending from 4 km SW. of Du Craig to 3 km south of it, there occurs an area of moundy drift and meltwater channels followed eastwards by a kilometre-long outwash plain that in turn merges eastwards into a faint raised shoreline that has been traced for a kilometre and levelled at 13 points, giving

altitudes between 33·5 and 35·0 m O.D. (10 points between 34·2 and 35·0 m). Apart from indicating that the sea was about 35 m above its present level when glacier ice last occupied this part of the Forth valley, this morphological assemblage, suggesting the possibility of a halt in ice decay or a readvance, occurs in the location expected from the evidence at Rosyth.

The accumulation of the complex deposits of the western area (15 to 20 layers are distinguished in some bores) appears to be most satisfactorily explained in terms of ideas put forward by Carey and Ahmad (1961). They have shown that in the grounded shelf zone of a wet-base glacier the proportion of the glacier's weight supported by the ground-water seepage pressure is greater than would be the case with an equivalent column of ice freely submerged to the same depth below sea-level. Consequently ' tills and meltwater sediments deposited below sea-level would suffer only partial load compression or even none ' (p. 884). In such locations well-bedded marine sediments should contain ' interbedded layers from a few inches to several feet thick of till-like material without disturbance of the underlying sediments '. Carey and Ahmad also suggest that tills ' will pass rather abruptly into very different sediments ' and that ' strongly dragged and rolled structures ' will occur in a particular facies (p. 871). These descriptions are very similar to those available for the sediments of the western area at Rosyth. The ' boulder clay ' often rests on laminated marine deposits while in the cylinder section undisturbed well-bedded marine sediments were seen beneath ' boulder clay '. As mentioned above, the ' boulder clay ' is usually described as soft, while descriptions such as ' puddle clay with stones ' and ' very soft boulder clay similar to soft red clay but containing stone ' suggest material transitional between glacial till and marine clay. Rapid lateral changes in the deposits are indicated by the marked variations in the thickness and type of sediment recorded in closely-spaced bores (e.g. sections AB and CD, Fig. 5). The number of ' boulder clay ' layers actually recorded in different bores varies from none to five, but in some bores the ' boulder clay ' layers were so thin and closely spaced that condensed descriptions such as ' puddle clay with thin beds of boulder clay ' were given. While the strongly dragged and rolled structures described by Carey and Ahmad cannot be identified from bore logs, descriptions such as ' medium boulder clay mixed with sandy clay ' and ' boulder clay mixed with puddle clay ' are suggestive. It is therefore inferred that the deposits of the western area were laid down beneath a grounded glacier shelf zone, the eastern edge of the grounded ice corresponding approximately with the thick line shown in Fig. 2.

The Buried Gravel Layer

In the Rosyth area the late-glacial marine deposits cannot be confidently identified in bore records above an altitude of between −3 and −4 m. It is inferred that they have been separated from similar deposits on the adjacent land

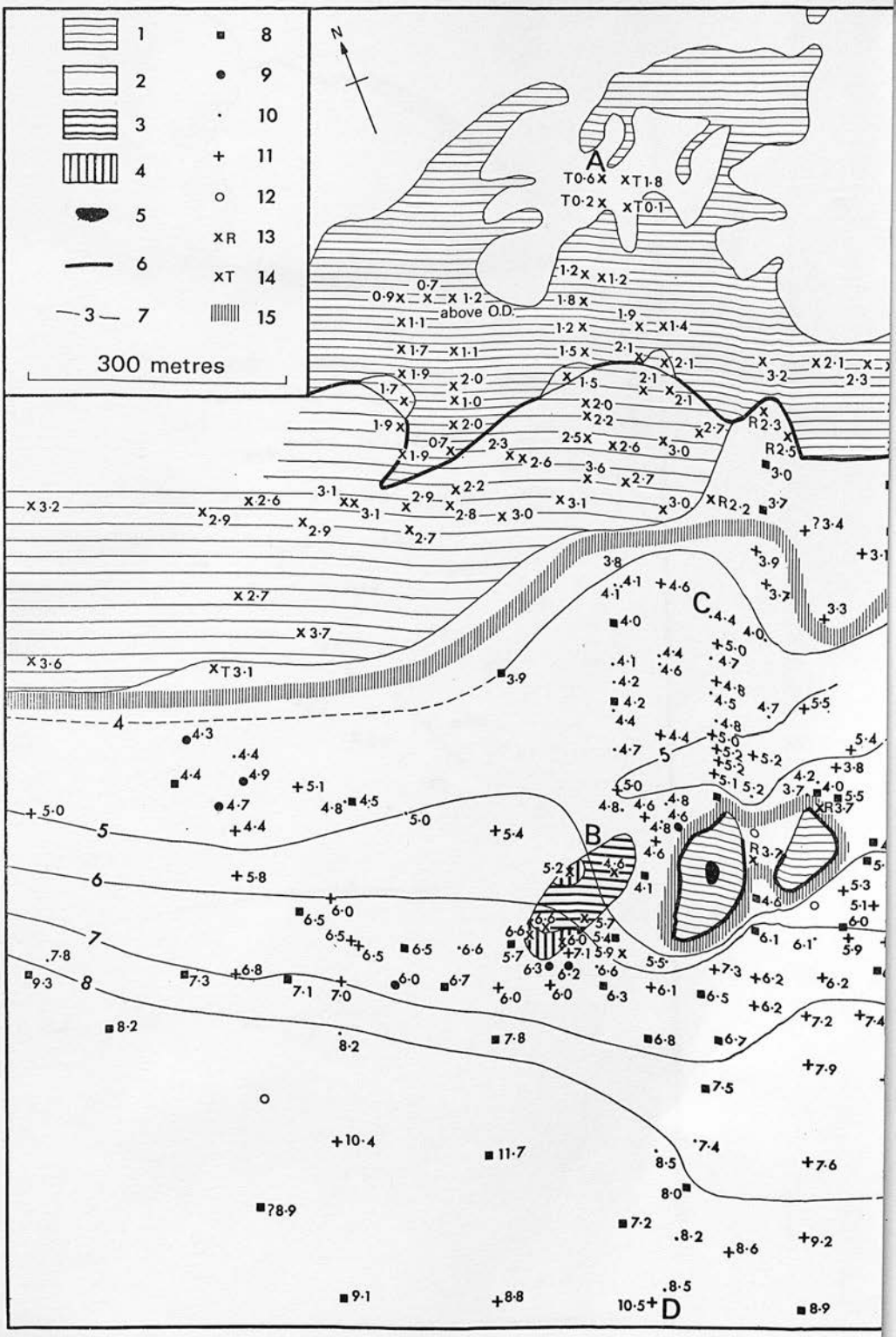
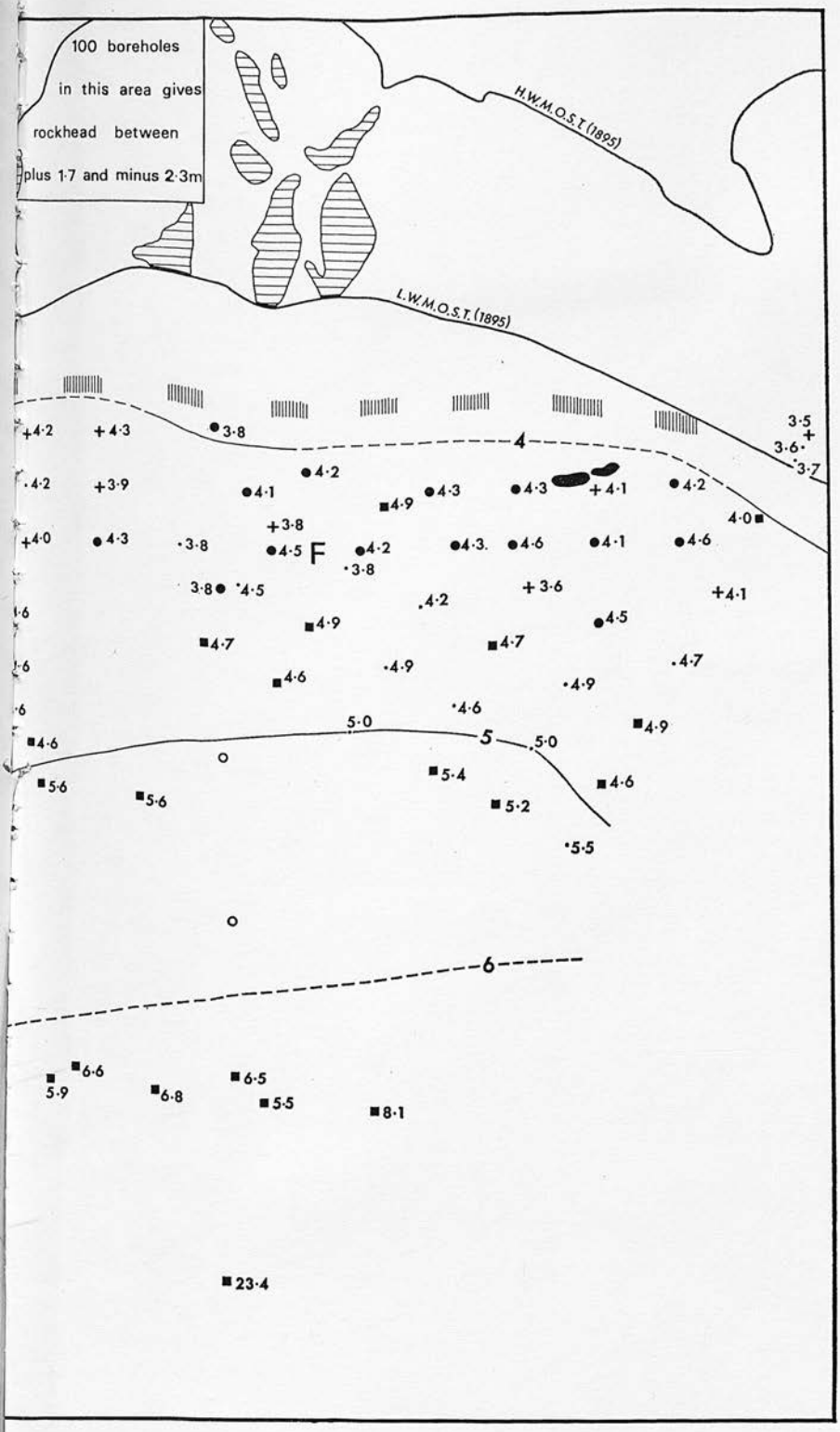
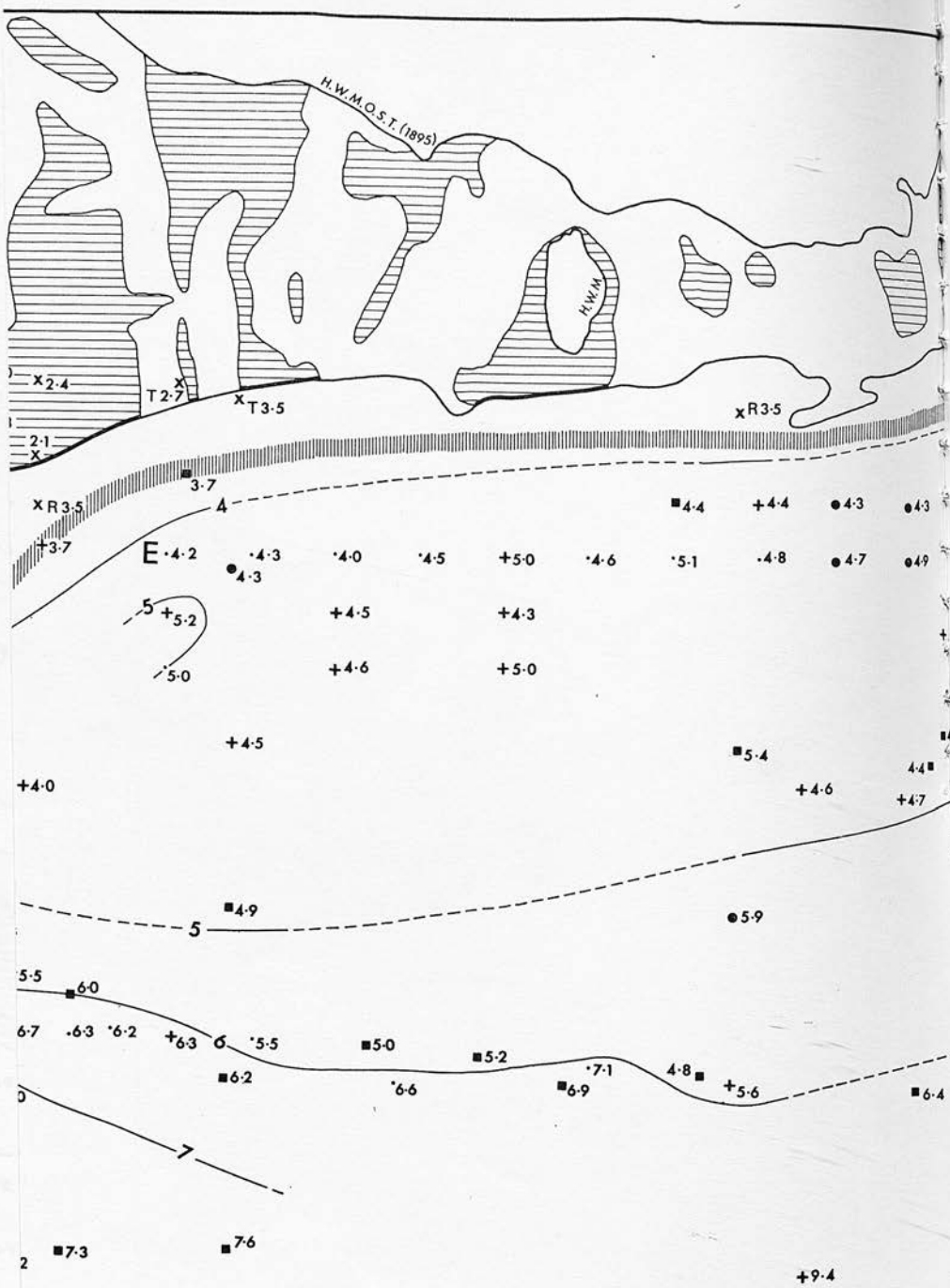


FIG. 4. The buried gravel layer. 1, Rock platform between tide marks. 2, Till platform, almost entirely below L.W.M. 3, Rock planation associated with buried gravel layer. 4, Till planation associated with buried gravel layer. 5, Small rock islands rising above H.W.M. 6, L.W.M. on rock. 7, Contours in metres below O.D. on surface of buried gravel layer (and on planated surface of late-glacial marine deposits where this layer is absent). 8, Bores in which buried gravel



in which buried gravel layer not recorded; figures are altitudes of top of late-glacial marine its. 12, Bores in which interpretation uncertain. 13 and 14, Bores in which rock and till diately overlain by ' silt and shells ' layer or by thin modern beach deposits. 15, Inferred on of Buried Gravel Shoreline.



layer or upper part of it is recorded as ' sand and gravel ', ' gravel ' or ' shingle '. 9, Bores in which buried gravel layer or upper part of it is recorded as ' boulder ' or ' boulders '. 10, Bores in which buried gravel layer or upper part of it is recorded as ' clay and stones ', ' boulder clay ', etc. For 8, 9 and 10 figures give altitudes of surface of buried gravel layer in metres below O.D. 11,

largely or (more probably) entirely by marine erosion, witnessed by the rock platform of considerable extent situated between present tide marks (Fig. 4). In the west evidence of marine planation continues below L.W.M. (−2·1 m), for the bore records reveal a considerable area of planated till lying mainly between −2·1 and −3·1 m.

In addition to this evidence of planation there is very clear evidence in the bore records of much more extensive marine planation at a lower level. Evidence of this planation occurs in some 250 bore records, only 5 records from the area in which planation occurred failing to identify it. Almost every bore log in this area refers to a surface layer of ' silt ' or ' silt and shells ', which is usually between 1 and 4 m thick and in general increases in thickness southwards. Beneath these soft deposits and resting on the soft late-glacial marine sediments the majority of bore logs refer to a completely different material. In 77 bores this material (or the upper part of it) is described as ' gravel ', ' sand and gravel ', ' shingle ', etc.; in a further 62 bores terms such as ' clay and stones ', ' sandy clay and stones ' and (much less frequently) ' boulder clay ' are used; and in an additional 29 bores the layer or its upper part is represented by a single ' boulder ' or (occasionally) by ' boulders '.

The composition of the coarse layer is most fully described in the record of the deposits encountered in the test cylinder section. The top 0·8 m is described significantly, as ' Stones, some smooth, some angular, with sand and mud resembling sea beaches now in neighbourhood '. This rests on stiff and compact sandy clay ' with worn rounded stones ', this in turn being succeeded by soft clayey sand and stones, occasionally 15 to 20 cm in diameter, followed by ' pure sand with lumps of stiff brown clay and occasional stones '. The total thickness here is 2·7 m.

The coarse layer varies in thickness and within the general area of its occurrence is not recorded at all in 81 bores. However, in these bores the junction of the surface layer of silt and shells with the underlying late-glacial marine sediments occurs at altitudes closely comparable with the surface of the coarse layer in adjacent bores (Fig. 4). The coarse layer tends to thin out towards its northern limit and is often less than a metre thick here, while bores in which the layer of silt and shells rests directly on the late-glacial marine sediments are most numerous here. The coarse layer is thickest in the south and south-east, where it attains 2 to 4 m (and occasionally more) in a considerable proportion of bores, and it is here that it is most frequently described as ' shingle ', ' gravel ' or ' sand and gravel '.

In Fig. 4 the altitude of the top of the coarse layer and, where it is not recorded, the altitude of the top of the late-glacial marine sediments, is given as recorded in some 250 bore logs. In drawing the one-metre isopleths shown in this figure, account was taken of the fact that altitudes of drift deposits in commercial bore logs are often slightly inaccurate (comparison of adjacent bore records suggests

errors up to about half a metre in the present area). The isopleths were therefore drawn as smooth lines related, where possible, to groups of bores rather than to individual bores. Such minor discrepancies are not significant in the general pattern, however, for it is clear from Fig. 4 that the surface of the coarse layer (and of the late-glacial deposits where this is not present) slopes down gently from about or slightly above −4 m in the north to about −9 m in the south, the principal exception being the slight southward rise on the northern side of Du Craig and adjacent rock outcrops. The form of this surface is thus entirely in accord with a marine erosion interpretation.

Over almost the whole area of its occurrence the coarse layer truncates the late-glacial marine sediments (along with associated ' boulder clay ' in the west). In a small area west of Du Craig, however, the surface ' silt and shells ' layer rests directly on bedrock in three bores, and in three more it rests on thin ' boulder clay ' (interpreted as glacial till), at altitudes between −4·6 and −6·6 m. These altitudes accord with those of the surface of the coarse layer (or of the top of the late-glacial marine deposits where the coarse layer is absent) in adjacent bores (Fig. 4) and strongly suggest that here a small area of bedrock and till has been planated by the sea.

Four anomalous bore records occur in the southern part of the area shown in Fig. 4. In a bore in the south-east the surface silt layer continued down to −23·4 m before a thin gravel layer was met with. In two bores south-west of Du Craig the silt layer extended down to −10·4 m and −11·7 m, while a third bore in the same area was terminated while still in the silt layer at −15·6 m. All four of these bores are in locations where the present sea bed is locally lower and the anomalies are very probably associated with present or former deep channels beneath the Forth. At the present time the Forth deepens rapidly to −15 m or more a short distance south of the map area.

In Fig. 4 the position of the shoreline at the time the coarse layer (or, more probably, the higher landward part of it) was produced is suggested, based on the following evidence. Over a considerable area in the central part of the site the surface of the coarse layer rises very gently northwards, the distance between the −6 and −4 m contours being over 600 m. Between 75 and 175 m north of the northernmost bores in which the coarse layer is recorded old Ordnance Survey maps show rock extending down to L.W.M.O.S.T. Since the latter is at −2·1 m and the surface of the coarse deposits in the bores mentioned is between −3·7 m and −4·6 m, it is evident that the very gently inclined surface of the coarse layer ends in the intervening area. It thus appears reasonable to infer that the altitude of the shoreline (defined as the landward edge of the ' flat ') is at or slightly above −4 m, probably in the range −3·7 to −4·0 m. In the western part of the map area the coarse layer again rises gently northwards to about or slightly above −4 m. A short distance farther north occurs the area of planated till whose surface normally lies between −2·1 and −3·1 m. It thus appears that here also the

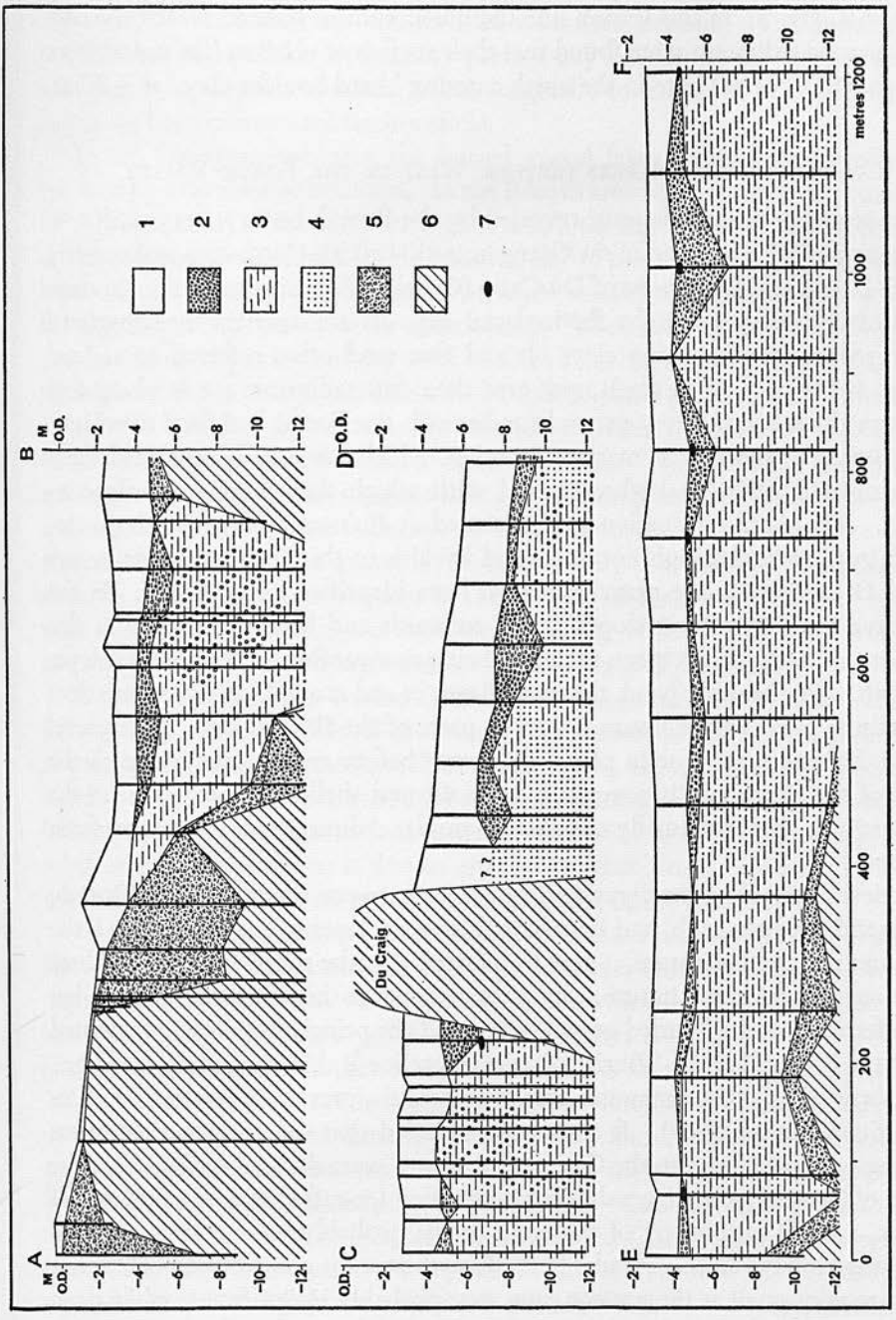


FIG. 5. Sections between A and B, C and D, and E and F in Fig. 4. 1, Silt and shells. 2, Gravel. 3, Clay. 4, Sand. 5, Till. 6, Bedrock. 7, Boulders. The sections are drawn to give both interpretation and, so far as possible, the original borehole data. For example, where the late-glacial marine sediments are recorded as 'red clay', 'puddle clay', etc. they are shown by symbol 3, where they are described as 'sandy clay', etc. symbols 3 and 4 are used, where they are inter-bedded with 'boulder clay' 2 and 3 are used, etc.

shoreline is at or slightly above −4 m. A similar inference may be made from section AB, Fig. 5: in this section line the northernmost bore in which the late-glacial marine sediments were found met their surface at −3·8 m (the coarse layer being absent), the next bore to the north entering 'hard boulder clay' at −2·7 m.

Comparison with Areas farther West in the Forth Valley

The sequence of drift deposits revealed by the Rosyth bores is very similar to that occurring at Bo'ness and in the Grangemouth–Falkirk–Airth area, respectively some 9–10 and 13–22 km west of Du Craig (Cadell 1883; Sissons 1969). In these areas bedrock, till and patchy fluvioglacial deposits are overlain by late-glacial marine sediments comprising clay, silt and fine sand often referred to as 'red clay' in old bores. As in the Rosyth area these fine sediments are in places tens of metres thick beneath low ground (or beneath the Forth) but they thin landwards owing to subsequent marine planation, which has usually separated them from similar deposits on higher ground with which they were formerly continuous. The marine planation is represented at Bo'ness by a gravel layer described by Cadell (although not attributed by him to this agency), while in and around Grangemouth the gravel layer has been identified over 28 km^2. In the latter area the gravel layer slopes gently seawards and becomes thicker in this direction, in both these respects therefore being comparable with the coarse layer at Rosyth. On the other hand, the gravel layer in and around Grangemouth does not attain as great a thickness as it does in parts of the Rosyth area. The gravel layer at Bo'ness occurs (or in places did occur before reclamation) beneath the waters of the Forth and is covered by soft silt and shells, as at Rosyth; in the Grangemouth area it is usually covered by similar sediments now raised to form carse lands.

In view of the close stratigraphical similarity between the three areas (Rosyth, Bo'ness and Grangemouth) and since in each area the layer of coarse deposits is the most extensive marine feature, it may be inferred that the coarse layer of the three areas is one and the same stratum. Since this stratum in the Grangemouth area has been referred to as the 'buried gravel layer' and the principal shoreline associated with it has been called the 'Buried Gravel Shoreline', these terms are used here. It was shown in the Grangemouth area that these features were produced in late-glacial times (Sissons 1969). It was also concluded that the coarser constituents of the buried gravel layer in the Grangemouth area were derived mainly from the erosion of late-glacial marine sediments containing ice-rafted stones, of till and of bedrock, a small proportion of the constituents probably having been supplied by streams from the adjacent land. In the Rosyth area the contribution of streams, which are very small at the present time, was probably insignificant, while there is only limited evidence of erosion of till and bedrock. It therefore appears that the coarser constituents of the buried gravel layer at Rosyth were derived mainly

from ice-rafted material present in the late-glacial marine deposits and, in the west, also from the ' boulder clay ' interbedded with these deposits. The buried gravel layer is thus comparable with many present-day beaches in the vicinity in consisting of a lag deposit of coarse debris originally contained in finer deposits that have now been transported farther afield.

In the Grangemouth area the buried gravel layer is only occasionally represented by a boulder or boulders. In the Rosyth area boulders are more common. In 19 (11 per cent) of the 168 bore logs in which the buried gravel layer is recorded the layer is represented by a single boulder, in 7 (4 per cent) more a single boulder is recorded as forming the top part of the layer, while in 3 more bores the whole or the upper part of the layer is described as ' boulders ', such records together amounting to 17 per cent of the total. Most of the boulders are recorded as bored through for 0·3 m, but this is likely to be their smallest (approximate) dimension. In addition boulders are occasionally mentioned within the gravel layer, including one bored through for 1·5 m and another for 1·3 m. The considerable number of references to boulders thus suggests that a significant part of the buried gravel layer is composed of very coarse material. It is also noticeable that boulders are usually recorded at the top of the gravel layer, or as wholly constituting it, and that these occurrences are far more frequent in the vicinity of the Buried Gravel Shoreline than elsewhere. This suggests shoreward pushing of the boulders by winter sea ice, a suggestion in accord with the late-glacial origin of the feature.

The formation of the Buried Gravel Shoreline in the Grangemouth area was followed by a rise of sea-level that resulted in further planation at a higher level in parts of the area. In the Rosyth area this period of planation cannot be recognised: to what extent it contributed to the formation of the platform that occurs between present tide marks (and extends somewhat lower in the west), and to what extent this platform is due to erosion at other times, especially by the sea while standing at about its present level in recent times, is unknown.

In the Grangemouth area the Buried Gravel Shoreline is situated at or slightly above Ordnance Datum and the buried gravel layer has been identified down to slightly below − 6 m. In the Rosyth area, however, the shoreline appears to be located at about − 3·7 m to − 4·0 m and the gravel layer slopes down to − 9 m (and perhaps descends further to the south of the area investigated by boring). While more evidence is needed to determine the shoreline gradient confidently, it nevertheless appears from these figures that it declines eastwards at between 0·20 and 0·25 m/km. Since isobases so far determined for late-glacial shorelines in SE. Scotland trend approximately N.18°E.–S.18°W. (Cullingford and Smith 1966; Smith *et al.* 1969) the true gradient of the Buried Gravel Shoreline is probably greater than that measured in a west–east direction. The difference is likely to be slight, however, and the true gradient is probably within the range given. This gradient is in accord with the evidence of successively decreasing gradients for successively younger shorelines in the Forth valley. For example the (older)

Main Perth Shoreline has a gradient of about 0·43 m/km and the (younger) Main Postglacial shoreline slopes at about 0.80 m/km.

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TECHNICAL NOTES

Geomorphology and foundation conditions around Grangemouth

JOHN BRIAN SISSONS

SUMMARY

Study of the geomorphic history of the Grangemouth area, based particularly on borehole data, has revealed a series of relative sea-level changes. Each major change in sea-level resulted in a particular set of drift deposits and, in one instance, in widespread marine planation. Through an understanding of the geomorphic history of the area some 2000 drift borehole records have been correlated and the areal and altitudinal limits of various deposits established. Much of this information can be portrayed on simple maps of value in site investigations and in local and regional planning. It is suggested that many badly planned site investigations involving expensive boring operations could be avoided not only in this area but in the country as a whole if such maps were available. This problem could be dealt with by the establishment in a centre such as the Institute of Geological Sciences of a computer data bank of drift borehole records linked with an advisory service staffed by suitably qualified specialists.

1. Morphology, drift deposits and relative sea-level changes

GRANGEMOUTH is situated on almost flat ground composed of post-Glacial mud flats raised slightly above sea-level and known in Scotland as 'carse'. The extent of the carse around Grangemouth is shown in Fig. 1; it lies mostly below about 25 ft (8 m) O.D., except in the north-west, where it reaches a maximal altitude of 42 ft (12·8 m) O.D. Landwards from the carse the ground often rises steeply through an interval of 50 ft (15 m) or more. In many places another stretch of flattish ground occurs whose surface lies between about 80 and 120 ft (24 to 37 m) O.D. This higher area constitutes late Glacial raised beaches and is composed essentially of sand merging downwards into laminated clay.

The laminated late glacial clay also occurs extensively beneath the carse muds of the lower ground and in places exceeds 100 ft (30 m) in thickness. The laminated clay rests on patchy fluvio-glacial deposits and these in turn on glacial till, which rests on bedrock. Bedrock is rarely exposed at the surface on low ground: its main outcrops are in the bluff that separates the carse from the late Glacial raised beaches.

The main features of the drift succession and of the related morphology are shown diagrammatically in Fig. 2. Late Glacial marine deposition is shown by the laminated clays, capped by sand, which extend up to an altitude of 120 ft (37 m), the maximal altitude of the highest (sloping) raised shoreline. The period of high sea-level was followed by marked emergence, and the sea then stood for a considerable period in late Glacial

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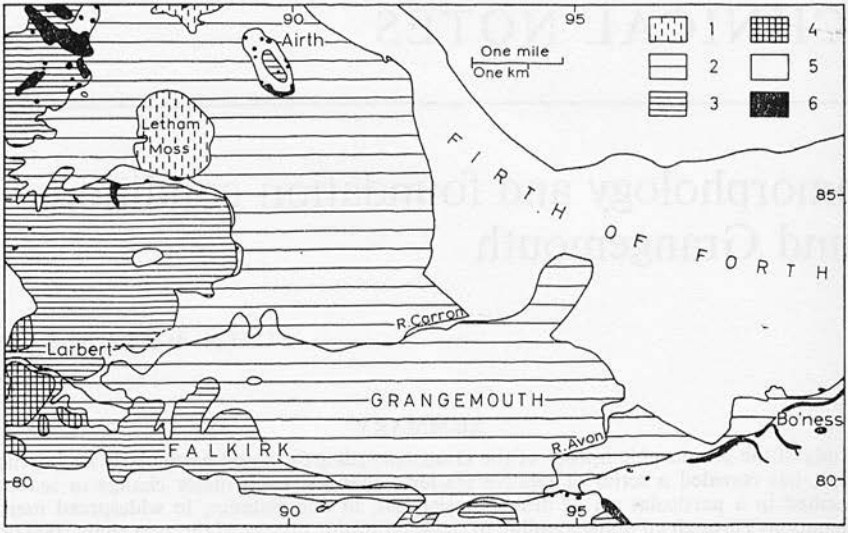


FIG. 1. Surface drift deposits of the Grangemouth area. 1, Peat. 2, Carse clay. 3, Late Glacial marine deposits, mainly sand and laminated clay. 4, Outwash gravel. 5, Till, occasional sand and gravel. 6, Bedrock.

times at about its present level. During this period extensive marine planation occurred: the late Glacial sediments were truncated over a large area, and, locally, rock and till were also truncated. In and around Grangemouth the total area subject to planation at this time was about 11 sq mile (28 km^2). In Fig. 2 the planation is represented by a buried gravel layer (the old beach) resting on late Glacial marine sediments, till and rock, and by the bluff that backs the carse. Rock outcrops are common in the bluff owing to removal of the former drift cover by the late Glacial sea.

A rise of sea-level followed and the period of erosion was succeeded by one of deposition, a consequence of the Zone III re-advance and subsequent glacier decay. The deposits

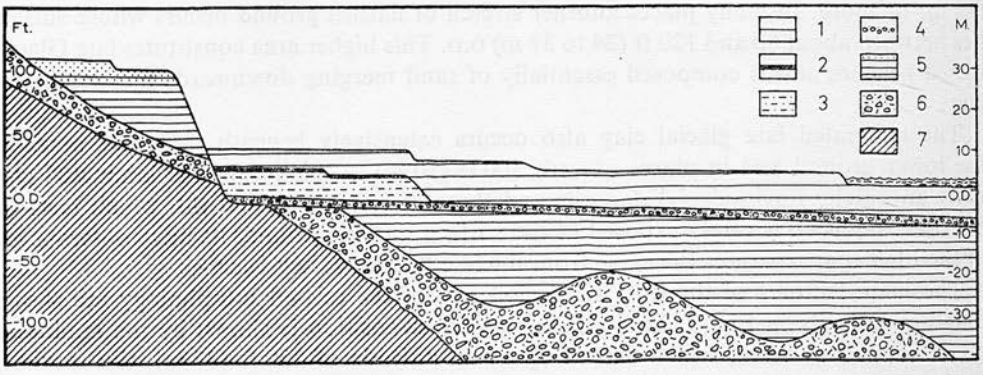


FIG. 2. Diagrammatic section of drift deposits and related morphological features in the Grangemouth area. 1, Carse clay. 2, Peat. 3, Buried beach deposits. 4, Buried gravel layer. 5, Late Glacial marine deposits. 6, Till. 7, Bedrock.

consist of micaceous grey silt and fine sand and, in this area, do not exceed 24 ft (7·3 m) O.D., the limit of the transgression. When the sea withdrew from the associated raised beach, peat began to accumulate. Meanwhile another beach of rather finer composition was formed at a slightly lower level and, on this beach also, with subsequent lowering of sea-level, peat began to accumulate (Fig. 2).

There then followed a transgression that resulted in the widespread deposition of the carse clay, up to a maximal altitude of 42 ft (12·8 m) O.D. Consequently the peat-covered beaches were buried beneath the carse clay, so that one may now refer to 'buried beaches' covered by 'buried peat'. Finally, the sea fell in stages to its present level, widespread deposition of carse clay meanwhile continuing. Similar material still continues to accumulate and is seen as the broad mud flats exposed at low tide at Grangemouth (cf. Kenney 1964).

2. Foundation conditions

The carse in and around Grangemouth is composed mainly of silt and clay and has a high water content. The surface 3 ft (1 m) or so has dried out and weathered to form a firm 'crust' below which is very soft mud. The deposits have been drilled through in about 1500 boreholes and, especially in Grangemouth, hundreds of samples have been laboratory analysed and tested in the field by various site-investigation firms. The result of these numerous tests is that in all instances known to the writer the recommended safe bearing pressure has been given as between 0·25 and 0·60 tons/sq ft (27 and 64 kN/m^2). It is also recommended that flexible joints be placed between adjoining buildings of different heights, that foundation loads should be spread, that the crust should not be pierced by foundations, and so on. Since the carse is a morphological unit, whose areal and altitudinal limits are precisely known, it is possible to apply these recommendations to a large area around Grangemouth. For light structures it is therefore no longer necessary to put down considerable numbers of boreholes and carry out expensive laboratory tests. Instead, at most, only one or two boreholes and related tests need normally be made for a site to check that conditions are in accord with those described.

The late Glacial laminated clays lying landwards of the carse also appear to have a considerable degree of uniformity as regards bearing capacity. Where exposed at the surface or covered by a surface layer of sand, recommended safe bearing pressures are given as between 0·75 and 1·25 tons/sq ft (80 and 134 kN/m^2) in reports seen by the writer. More information is needed about these deposits, but it should be possible in the future to make generalizations about them.

Peat can give rise to construction problems owing to its high compressibility, and it is therefore useful to know its distribution. Surface peat is obvious, but the distribution of buried peat is not. When bores recording the buried peat that underlies the carse clay in parts of the Grangemouth area are plotted on a map, however, the peat is seen to occur towards the landward margin of the carse. This distribution becomes meaningful only when it is recognized that the buried peat rests on the buried beaches and on the sub-carse slope between the upper limit of these beaches and the edge of the carse. These relationships are evident in Fig. 3, which shows the north-western part of the area covered by Fig. 1. The relationship between buried beaches and buried peat is not established only in a small area in the north-eastern part of Fig. 3 where the full

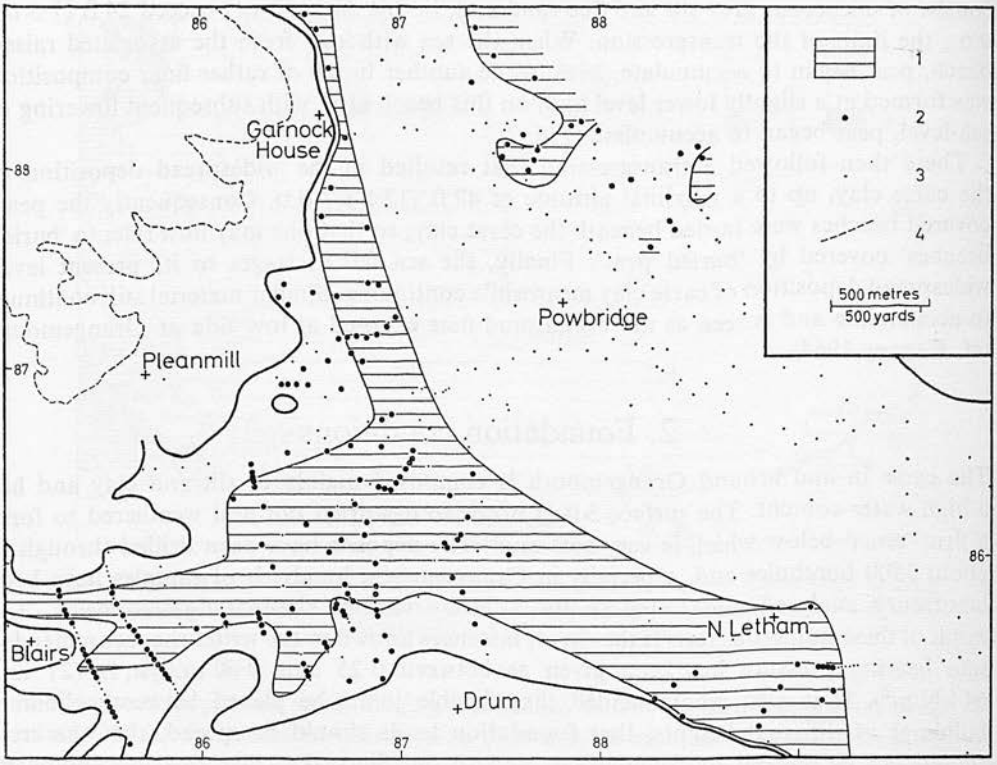


FIG. 3. Buried beaches and buried peat in the north-western part of the area shown in Fig. 1. (1), Extent of buried beaches, where known. (2), Bores in which buried peat recorded. (3), Bores with drift details in which buried peat not recorded. (4), 100 ft (30 m) contour.

extent of the buried beaches is not known. The peat is locally absent within the area of the buried beaches, but it has nowhere been found between their outer margin and the Forth. One may also add that, over most of the area in which buried beach deposits are shown in Fig. 3, two features can be identified. On the lower buried beach the peat layer is up to 2 ft (0·6 m) thick and is more highly compressed than on the higher beach; on the latter beach and between its shoreline and the carse shoreline the buried peat is thin but often very soft.

When the buried gravel layer (Fig. 2) was being formed the streams that now descend to the carse from the surrounding higher ground cut down their valleys in response to the relatively low sea-level. These valleys were later partly infilled by the buried beach deposits, on which peat then accumulated. Later they were further infilled by the carse clay. The infilling is most extensive in the Carron valley on the north side of Falkirk (Fig. 1). Since the time of the highest post-Glacial sea-level the River Carron has cut into its valley fill, the remnant parts of which now form an almost horizontal terrace composed mainly of carse clay resting on the buried peat, which in turn rests on the buried beach deposits or on river gravel. As can be seen in Fig. 1, small valleys tributary to the Carron were also partly infilled with carse clay at their lower ends, while similar infilling occurred in valleys in the north-western part of the area shown in Fig. 1. In

these small valleys peat accumulation was favoured, the buried peat in one valley having been shown by bores to be several feet thick and in one place attaining a thickness of 8 ft (2·4 m) and entirely replacing the carse clay. These small valleys thus provide unfavourable ground for building. The reason for this is obvious once the geomorphic history is known and the limits of the unfavourable ground can often be accurately delimited using this knowledge; without this knowledge, however, interpolation between boreholes is much more likely to be incorrect.

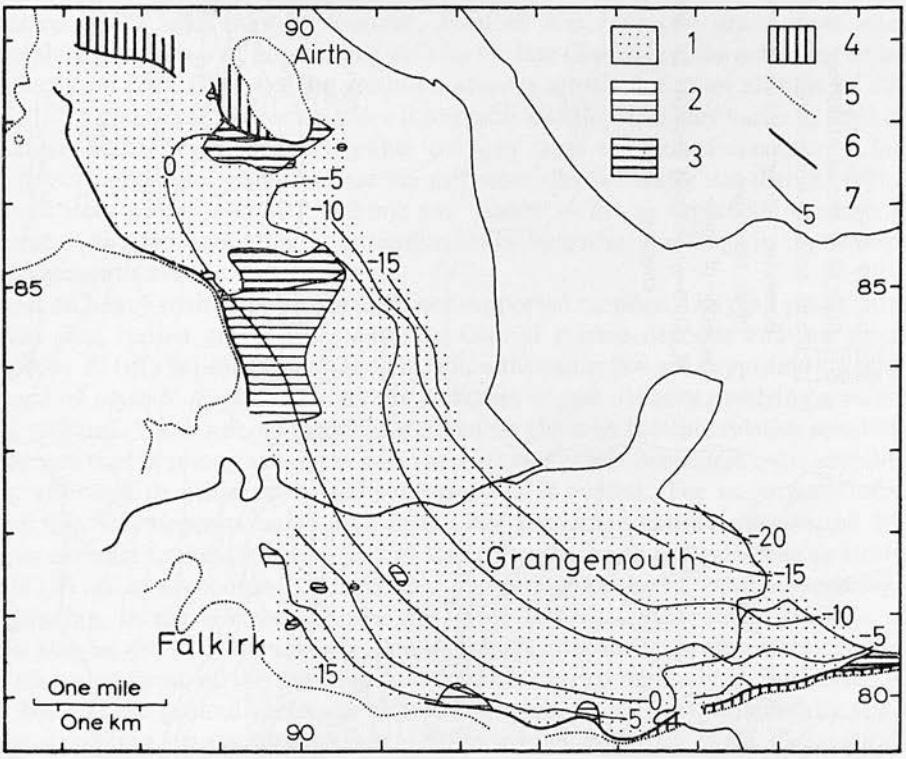


FIG. 4. Altitude and extent of the buried gravel layer. (1), Buried gravel layer resting on planated late Glacial marine deposits. (2), Till planated in association with buried gravel layer. (3), Rock planated in association with buried gravel layer. (4), Carse clay less than 40 ft (12 m) thick and normally overlying rock. (5), Edge of carse. (6), Well-defined limit of marine planation. (7), Contours on surface of buried gravel layer (or on planated bedrock or till) at 5 ft (1·5 m) intervals.

Beneath the greater part of the carse clay in and around Grangemouth, the buried gravel layer produced by marine erosion in late Glacial times occurs. The altitude of the gravel layer is accurately known from some 900 bore logs, which show that it slopes down gently towards the Forth. A detailed map of the buried gravel layer showing its relation to other geomorphic and stratigraphical evidence has been published elsewhere (Sissons, 1969). In Fig. 4 the altitude of the surface of the gravel layer is shown by 5 ft (1·5 m) contours, the borehole data being omitted. When the altitude of the ground surface is known the thickness of the carse clay can be determined accurately over almost the whole of the Grangemouth area. This is particularly important where the gravel

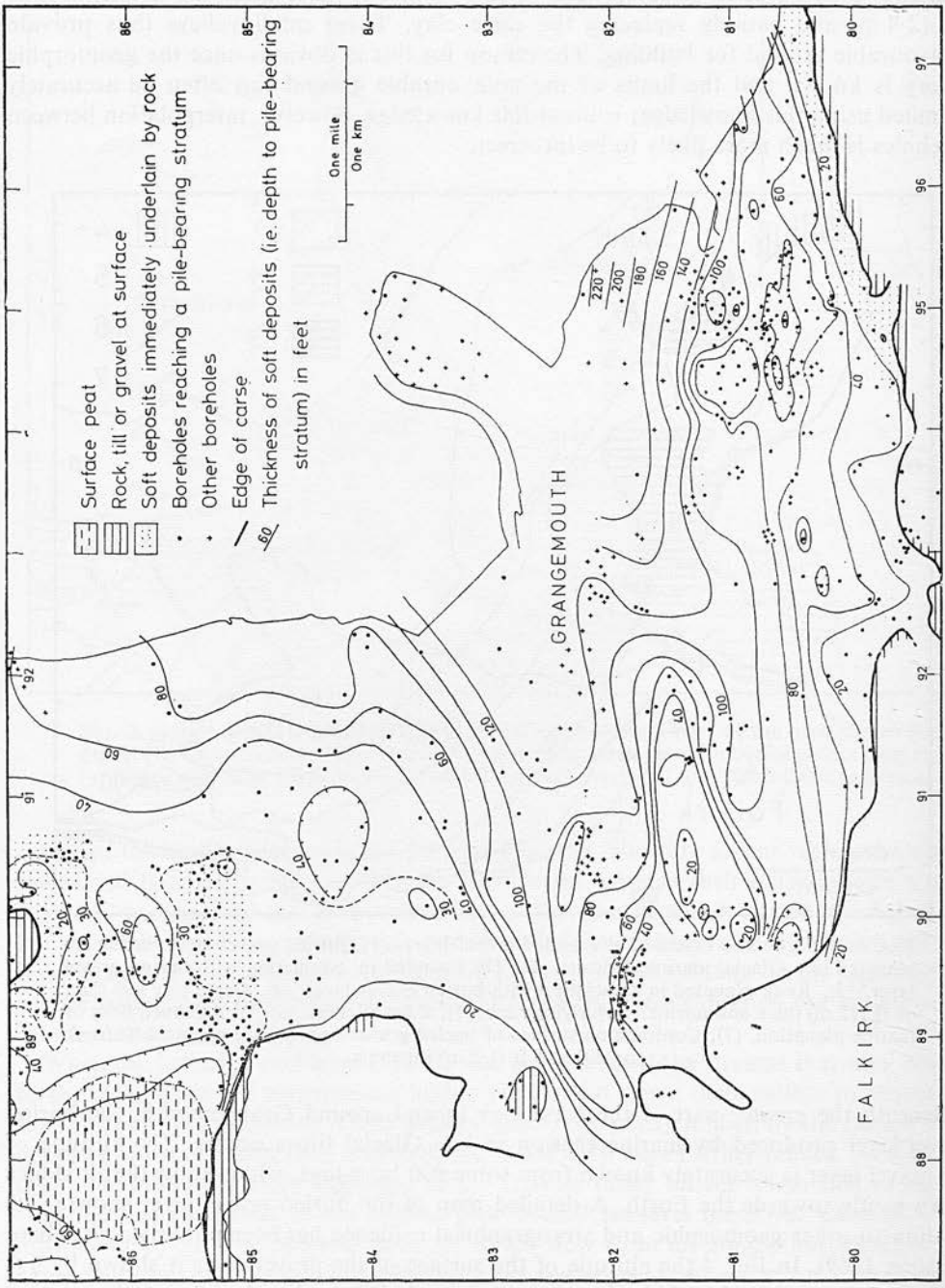


FIG. 5. Depth to a pile-bearing stratum in the carse of the Grangemouth area (20 ft = 6·1 m).

layer rests on or is replaced by planated till or planated rock. Two such areas, of considerable extent, occur between Grangemouth and Airth (Fig. 4), each being proved by a large number of bores. In the more northerly, near Airth, the carse clay rests on a wave-cut rock platform that extends for a mile (1·6 km) in a W–E direction. In the more southerly, rock and strong till are truncated over an area 1·3 miles (2·1 km) long and up to a mile (1·6 km) wide. In both areas the rock and till have been planated by the late Glacial sea within a few feet of O.D. Since the ground surface is at an altitude of 20 to 25 ft (6 to 8 m), there is only this thickness of soft deposits, and beneath is good material on which piles may be founded. West of Airth are two other areas where, owing mainly to removal of pre-existing drift by the late Glacial sea, the carse clay usually rests directly on rock (Fig. 4). The ground surface is mostly flat at an altitude of 35 to 40 ft (11 to 12 m) but the bedrock surface is irregular and the carse clay varies in thickness up to about 40 ft (12 m). Taken together the four areas mentioned constitute a large belt of flat or almost flat ground where the soft carse clay is usually less than 25 ft (8 m) thick (and does not exceed 40 ft (12 m)) and which, so far as superficial deposits are concerned, is the most suitable for extensive heavy industrial buildings in the whole of the Grangemouth area.

Almost all heavy structures on the carse are supported on piles. The piles pass through the carse clay, buried gravel layer and late Glacial marine deposits and are usually taken about 10 ft (3 m) into the underlying till, although a few are supported by fluvioglacial gravel or rock. Figure 5 shows the thickness of soft deposits overlying a suitable bearing stratum. The borehole locations marked on the map indicate relative reliability: it can be seen that in some parts of the area the thickness of soft deposits is quite accurately known, although in other parts more information is needed. The importance of the areas of thin soft deposits mentioned in the last paragraph can be appreciated from Fig. 5; in contrast to these areas, in part of the region the depth to a pile-bearing stratum is 100 ft (30 m) or even over 200 ft (60 m). Figure 5 could be of value in regional or local planning, in the selection of sites for large factories, high blocks of flats, etc. It could also be useful in the investigation of specific sites for large structures. The map shows that over most of the area the soft deposits change regularly in thickness. This occurs because the ground surface is almost flat, while the soft deposits usually rest on a gently undulating ice-moulded surface of till, or on planated rock or till. Consequently, if the possibility of erecting a structure that could need piles were being considered, the most economical site investigation procedure in most of the area covered by Fig. 5 would be to put down one or two bores on the proposed site: if the results accorded approximately with the isopleths shown in Fig. 5a detailed site investigation would not normally be needed.

3. Significance of geomorphological studies in relation to site investigation

The preceding section has sought to show how the interpretation of drift borehole data in terms of geomorphic history may convert a large volume of detailed borehole information into simple patterns that are readily portrayed on maps that could be of value in reducing costs of site investigations. Failure to make use of the information that already exists has resulted in considerable sums of money being wasted in site investigations in the Grangemouth area. Two examples may serve to illustrate this.

In one part of the area it was decided to investigate the possibility of erecting some large structures. On either side of the selected site and close to it bore logs already available showed soft drift deposits to be 100 ft (30 m) thick. Despite this, and despite the fact that the site was quite small, ten bores were commissioned to go down to a suitable bearing stratum for piles. All ten bores gave almost exactly the same detailed record which, in summary, meant that piles about 100 ft long would be needed for the proposed structures. The project was therefore abandoned, £2000 having been spent on the site investigation. A single bore in the centre of the site would have been sufficient to confirm the borehole evidence previously available and about £1800 could thus have been saved.

The other example relates to the proposed erection of large structures on a site near one end of which an old bore had previously revealed 200 ft (60 m) of soft deposits and at the other end of which another old bore had previously proved the existence of 300 ft (90 m) of soft deposits. To find a suitable bearing stratum 25 bores were commissioned to go down to a maximal depth of 130 ft (40 m). Not surprisingly the bores encountered only soft deposits. The site investigation cost £7000, a sum that was completely wasted.

The money wasted in these two examples was a very small sum in comparison with the cost of the proposed engineering structures themselves. This does not mean, however, that it can be ignored: an alternative point for consideration is the cost in the United Kingdom as a whole (and indeed in the rest of the world) of site investigations that are partly or wholly misconceived. Such investigations include those where bore records that already exist are ignored or not known; those where the significance of existing bore records is not sufficiently appreciated, leading to additional excavation and foundation costs; those where more bores are put down than is necessary for the purpose; and those where bores are made on an arbitrary grid pattern when, had geological and geomorphic evidence been taken into account, a more reliable picture of the drift stratigraphy might have been obtained, possibly with fewer bores. When one bears in mind that the rate at which site-investigation bores are being put down has increased greatly in the last ten years or so, and that in some localities bores are so numerous and close together as to be almost unplottable on the 1 : 10 560 scale, one suspects that the sum wasted annually in the United Kingdom on unnecessary or badly located bores may run into hundreds of thousands of pounds.

If the vast amount of drift borehole information that is being obtained is to be made use of, it must be systematically collected and stored in a way that makes it readily accessible to all potential users. The organization best fitted to carry out this task in the United Kingdom is the Institute of Geological Sciences, which already has a large number of borehole records. To enable the Institute to carry out this task effectively there appear to be three principal requirements.

Firstly, a computer data bank for drift borehole records needs to be established. A pilot study for such a project has already been made by D. W. Rhind and the writer under contract to the Institute of Geological Sciences. Programs have been prepared for outputting data in various forms, such as printing out full bore logs in an area specified by National Grid references, printing out bores in which a particular deposit is recorded (e.g. sand and gravel, peat), and producing distribution and contour maps of rockhead or drift thickness in a specified area[1].

[1] Rhind, D. W. & Sissons, J. B. Data banking of drift borehole records for the Edinburgh area. (Report to the Institute of Geological Sciences, 1969.)

Secondly, the records of drift borehole data in the Institute of Geological Sciences need to be brought up to date and kept up to date. As an illustration of their imcompleteness one may cite the Grangemough area where the writer, for certain 1 : 10 560 sheets, has up to four times as many borehole records as appear in the Institute's files. In the area of the Edinburgh 1-in Geology sheet, which was used for the data bank pilot study, over 6000 bore logs have been collected in the course of a year to bring the Institute's total for the sheet to about 14 000. In giving these figures the writer is not implying any criticism of the Institute: the Institute does not have sufficient staff available to cope with the volume of data that exists. Consequently much valuable borehole information has already been permanently lost, one cause of this being that many firms 'lose' or destroy bore records within a year or two of the completion of an engineering structure.

Thirdly, it is suggested that in the compilation and interpretation of such data, it is necessary to employ specialists in the study of landforms and drift deposits.

It is recognized that the cost of establishing and maintaining such a data bank on a national scale would be considerable but the cost of not doing so is likely to be far greater.

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The last glaciers in the Lochnagar area

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Abstract

Moraines in Glen Callater, Glen Muick and the four corries of Lochnagar are used to delimit a glacial readvance. The freshness of the forms and their relation to large solifluction lobes imply that the readvance is the Loch Lomond or Zone III readvance. The very different type of interpretation recently given by Sugden for the adjacent Cairngorm area is discussed and rejected.

Introduction

The ground draining to the Dee around Lochnagar contains abundant glacial features that have been relatively neglected in the literature. The basic work is the account by Barrow and Craig (1912), while Charlesworth (1955) has briefly considered the deglaciation of the area. The former referred to small glaciers lingering in the corries of Lochnagar after the valley glaciers had disappeared, while the latter tentatively mapped several retreat stages. It is suggested in the present paper that there is evidence for one readvance of ice in the corries and in certain valleys and that this readvance occurred during Zone III.

The area is formed of granite except west of the Callater Burn, where metamorphic rocks crop out. Most of the higher ground lies



between 600 and 1000 m, although Lochnagar attains 1154 m. By far the most pronounced valleys are glens Muick and Callater, each along part of its length being a well developed glacial trough with truncated spurs, hanging tributary valleys and precipitous rock slopes. The results of glacial erosion are also very evident by Lochnagar, where four corries face directions between N. and NE.

Evidence for the last glaciers

Glen Callater The head of this glen consists of two short branches each cut some 250-300 m into the surrounding plateau. One branch ends in the vertical cliffs of Corrie Kander while the trough end of the main valley, bordered by cliffs, rises steeply to the plateau. That glaciers flowed down the glen from these sheltered N.- to NE.-facing valley heads at a late stage in the last glaciation is indicated by the abundant fresh morainic mounds that cover the lower slopes and, in places, the floor of Glen Callater for 7 km. From Loch Callater up-valley fluted moraines are well developed. These ridges, presumably indicative of the final direction of ice movement at a particular location, are aligned parallel with the main valley or trend slightly down-slope towards its axis (Fig. 1). Some fluted moraines occur almost at the top of the steeply sloping north-facing trough head of Glen Callater, suggesting that ice was also supplied from the shallow plateau valley that leads into the trough. Nearby on the eastern side of the trough fresh moraines climb the trough side and extend onto the floor of another shallow valley in the plateau, while moraines also occur in a depression in the plateau east of Corrie Kander. Thus the Callater glacier was partly nourished from three sources on the plateau.

Sections in the mounds in Glen Callater normally show unsorted morainic debris, boulders being common both in and on the mounds. A few sections in the mounds reveal water-deposited sediments. For brevity and because the internal composition of only a small proportion of the mounds can be readily determined, the term "moraine" is used throughout this paper. It is noticeable, however, that where tributary streams plunge down the slopes of Glen Callater (and of Glen Muick) mounds are often especially large and extend higher up the valley sides, suggesting that in such locations a considerable amount of debris has been carried into the ice by these streams.

North of Loch Callater fresh morainic mounds extend as far as 163863, where they end abruptly. The last feature is a 5 m high mound that rises sharply from the valley floor, just north of the bridge on the private road. Nearly 2 km within the moraine limit terraced outwash begins. Initially the surface of the outwash is about 4 m above the normal level of the present river but by the time it fades out, where the glen narrows a short distance north of the moraine limit, it has fallen to about 2 m above river level.

The positions of the lateral margins of the former glacier cannot normally be determined from the limits of morainic mounds, for these usually die out up-slope as the gradient of the valley side increases. In one locality, however, a lateral moraine marks the position of the glacier margin. This moraine occurs in the hanging valley of the Allt a'Chlaiginn situated north of a truncated spur whose upper part is a rock face (Fig. 1). The moraine begins on the flank of the tributary valley, where the latter is cut off by the main

valley, immediately north of the rock face. Here it is about 3-5 m high and consists of a jumbled mass of granite blocks, some several metres long, presumably derived mainly from the adjacent truncated spur. The moraine descends north-eastwards along the side of the tributary valley and then turns NW. as a subdued rise with many large blocks. North of the Allt a'Chlaiginn the line of blocks continues, although their frequency continues to diminish, until it is lost on the steep slope of the next truncated spur to the north. The total length of the feature is about 1 km.

At its southern end the lateral moraine is at an altitude of about 700 m. Since the valley floor at the glacier limit $4\frac{3}{4}$ km to the NW. is at an altitude of 440 m, the average gradient of the ice surface over this distance was 55 m/km. Opposite the southern end of the lateral moraine the ice over the valley axis was approximately 200 m thick.

Western corries of Lochnagar The two western corries of Lochnagar are virtually one, for the arête between them has been almost destroyed. Large granite boulders, many several metres long, have been carried from the corries in great quantities by glacier ice. In and close to the corries the boulders cover the ground completely in places and they remain plentiful along the valley leading NW. from the corries for a total distance of more than 2 km. The outward movement of ice from the more easterly corrie is revealed by lines of boulders clearly visible on vertical aerial photographs but often difficult to identify on the ground. Farther out from the corries low morainic mounds occur. Mounds come to an end where the ice-transported

boulders suddenly diminish in abundance and together appear to record a readvance of ice whose limits are shown on Fig. 1.

Eastern corries of Lochnagar These comprise a major corrie more than a kilometre across and a poorly developed corrie to the northwest of it. Abundant boulders have been carried out by glacier ice from the major corrie and many of them are heaped into moraines lacking any visible fine material. The outermost moraines show that the glacier extended nearly 2 km from the corrie backwall. Around the corrie lake the ground is almost covered with granite blocks, some of them arranged as end moraines that form a series of curves, one within the other, and define successive margins of the glacier as it shrank into the corrie. Boulders arranged in slightly irregular lines along the steep NW-facing slope of the corrie are probably lateral moraines disturbed by minor solifluction.

In the adjacent minor corrie the only clear moraine is 700 m long and formed of boulders. For most of its length it is a lateral feature, but at its northern end it curves round to define a glacier terminus. At its southern end it terminates against a moraine of the glacier that occupied the major corrie, implying that the two corrie glaciers were in contact with each other for a short distance (Fig. 1).

Glen Muick The upper part of Glen Muick is a rather shallow open feature with a floor at about 800-900 m. A rock step that slopes down to Dubh Loch (640 m) coincides with a complete change in the valley, which becomes a steep-sided glacial trough, the southern wall at one point being a vertical rock face 270 m high. From Dubh Loch the

valley floor falls rapidly to Loch Muick at 400 m, the valley as a whole meanwhile becoming less spectacular although remaining a deeply incised glacial trough.

The valley floor above the rock step is littered with boulders and low morainic mounds, while a few small mounds occur on the face of the step. Numerous small morainic mounds are present on both sides of Dubh Loch, but around the exit of the loch there is a massive accumulation of hummocky drift over 30 m high. Farther down the valley the fresh morainic topography continues to Loch Muick, the mounds being especially large and numerous where the ancestors of the present streams carried debris into the ice occupying the valley. Fluted moraines pointing down-valley occur SW. of the head of Loch Muick, but by far the largest feature of this type is an almost straight ridge 600 m long and up to 10 m high situated by Loch Buidhe.

The limit of the valley glacier responsible for the moraines in Glen Muick can be established in various places. On the northern side of Loch Muick moraines commence about $1\frac{1}{2}$ km from the head of the loch (283830). They have a sharp upper limit that rises up-valley. Seen from a distance they collectively give the impression of a layer of debris added to the steep valley side. Between Loch Muick and Dubh Loch the higher slopes of the northern valley side are very steep and morainic mounds are absent but, despite the gradient, drift is plastered on them. Several gulleys show that the drift has a definable upper limit, which increases markedly in altitude up-valley and continues the glacier limit recorded by the moraines alongside Loch Muick.

The limit becomes very clear again around Loch Buidhe. This loch is situated on the floor of a hanging valley 200 m above the bottom of Glen Muick, yet it is surrounded by fresh moraines including the 600 m-long feature referred to above. On the relatively gentle slope of this valley side a lateral boulder moraine extends for about 500 m. Up-valley its line is continued exactly by a linear accumulation of boulders that is the lowest of a series of large solifluction features that mantles the adjacent hill side. This coincidence of features seems unlikely to be fortuitous, especially as no large solifluction features occur below the line, and it appears that the glacier truncated the mass of solifluctated boulders. A short distance farther up the hanging valley the line is continued by a broad mound that blocks the valley floor (apart from a narrow stream cut). The limit of the valley glacier is then lost on rock outcrops but is probably recorded thereafter for about a kilometre by the highest of the lateral boulder moraines situated high up on the steep side of Glen Muick.

South of Loch Muick the glacier limit is indicated at a few points on the steep valley side by a sudden increase in drift thickness revealed by gulleys. The limit is clearly defined for $\frac{3}{4}$ km behind and up-valley from Corrie Chash by a boulder moraine that includes blocks up to 5 m long and follows the edge of the plateau just above the steep valley side. The boulders were presumably derived from the rock faces farther up-valley and the feature, which declines markedly in altitude down-valley, shows that the glacier here extended slightly beyond the confines of the valley onto the plateau edge.

The surface of Loch Muick is at an altitude of 400 m, while the bottom of the loch at the glacier terminus is at about 330 m. By Loch Buidhe, $3\frac{3}{4}$ km from the terminus, the glacier margin attained an altitude of 770 m. Over this distance the glacier surface therefore had an average gradient of 100 m/km measured to the surface of Loch Muick. This steep gradient presumably in part relates to the rapid descent of the valley floor and to the glacier having ended in a deep loch.

Date of the readvance

The existence in two valleys and four corries of fresh morainic features that end abruptly down-valley or are limited by clear lateral and terminal moraines indicates valley and corrie glacier readvance. Outside the readvance limit there is in various localities evidence of quite different glaciological conditions. For example, only 1 km NE. of the terminal moraines of the eastern corrie glaciers there are extensive fluvioglacial deposits related to ice-sheet decay, while $1\frac{1}{2}$ km SE. of these corrie glacier moraines a flight of some 30 melt-water channels similarly relates to ice-sheet conditions.

No direct evidence of the date of the readvance is available. However, since a Zone III readvance (the Loch Lomond Readvance) has been proved by radiocarbon dating at four localities in Scotland (Loch Lomond, Lake of Menteith, Loch Creran and eastern Mull) (Sissons, 1967a; Peacock, 1971; Gray and Brooks, in the press), and since this readvance is characterised by fresh hummocky moraines and is the only known readvance in Scotland frequently associated with clear terminal and

lateral moraines, it is difficult to avoid the conclusion that the readvance identified in the Lochnagar area is of Zone III age. It is also relevant that fluted moraines have so far been identified in Scotland only in locations known or believed to have been covered by Zone III ice.

Further indirect evidence is provided by periglacial features. Much of the higher ground of the Lochnagar area is covered by large solifluction lobes formed of granite boulders, some of the lobes having frontal slopes 5 m or more high. Galloway (1961) concluded that these massive lobes are fossil features. These large lobes are nowhere found within the readvance limit but they do occur just outside it. For example, they are present immediately outside the outer lateral moraine on the spur between the two eastern corries. Near Loch Buidhe large solifluction lobes are extensively developed on the hill slope above the lateral moraine, the lowest extending to within a few metres of the moraine. As mentioned above, the glacier appears to have truncated a mass of soliflucted boulders. Below the glacier limit there are no solifluction lobes: instead fresh hummocky moraines are present no more than 30 m vertically below the lowest of the lobes that occur outside the moraine. It is also significant that outside the readvance limits of the Lochnagar area major solifluction lobes are found in various places down to altitudes of between 700 and 800 m, yet such features are nowhere found within the ice limits shown in Fig. 1 even though these include considerable areas of much higher ground. Thus even in the corries at altitudes of 900-1000 m major lobes are absent within the mapped ice limits despite the steep slopes

and the abundance of boulders: only minor solifluction forms occur and some of the lateral boulder moraines can be followed across steeply sloping ground.

These relationships carry two implications. Firstly, they imply that the corrie and valley glaciers were contemporaneous. Secondly, they imply that no major solifluction lobes have been produced since the moraines were formed. Since the last period of severe cold occurred in Zone III, the simplest explanation is that the lobes were formed (or further developed) at this time, thus in turn implying that the glaciers are of Zone III age.

Meltwater channels and periglacial activity

Most of the 30 or so meltwater channels, mentioned above as occurring on the eastern slope of Lochnagar, are almost obliterated by solifluction. The lowest channels of the series, below about 700-750 m altitude, are very clear features however. Considerable solifluction down to about 700-750 m since ice-sheet decay is thus indicated. Yet at much higher altitudes there are four meltwater channels that are sharp clear features. Two of these occur on the Lochnagar ridge, commencing at 960 and 1000 m respectively, and the other two (one south of the limit of Fig. 1) occur on the southern watershed of Glen Callater beginning at 890 and 870 m respectively. All four channels are cut in rock, are between 5 and 10 m deep, and occur in the bottom of cols. The channels are not parts of complex channel systems such as commonly occur in cols but each is isolated, suggesting simple escape of water through each col from decaying ice.

The altitudes of the channels and their freshness suggest formation in association with the glaciers of inferred Zone III age. The two in the Glen Callater watershed lead southwards from shallow plateau valleys in which fresh moraines also suggest that such glaciers existed. It seems reasonable to infer that ice nourishing the Glen Muick and Glen Callater glaciers filled valley heads up to the level of the col channels, and the glacier margins have been drawn accordingly in Fig. 1. In view of the altitude and position of the four channels it seems likely that they were formed at a late stage in the decay of the valley glaciers.

Sugden's alternative interpretation

Sugden (1970) has given for the adjacent western Cairngorms a very different interpretation of the last glacial events from that presented above. Sugden concludes that in the western Cairngorms there was no Zone III readvance of valley glaciers; instead he presents "the concept of an ice sheet downwasting in situ with only a minor fluctuation to interrupt its overall decline" (1970, p. 213). He then criticises interpretations of the type presented above. Excluding criticisms requiring discussion of detailed evidence in the Cairngorms, Sugden's main points are as follows.

(i) "This sequence of events /i.e. Sugden's/ is far from being unusual and has many close parallels in Scandinavia" (p.213).

(ii) "In the Cairngorms the 'hummocky moraine' consists of fluvioglacial landforms associated with widespread stagnant ice. In

a mountainous area, stagnation on such a scale is more likely to reflect the downwasting of an ice-sheet ....." (p. 213).

(iii) Certain "high and impressive corries /in the Cairngorms/ have supported no more than minor glaciers" since the Spey ice-sheet downwasted (p. 214).

These criticisms may be answered as follows.

(i) Analogies with Scandinavia are not appropriate since the climate of much of Scandinavia is (and presumably was) quite different from that of Scotland, the mountains are much higher, and the Scandinavian ice-sheet was very much larger than the British ice-sheet and hence took longer to decay. It is more logical to make analogies with western Scotland and to argue that since a Zone III readvance of valley glaciers occurred there it should have occurred in the east also.

(ii) This argument is an expression of opinion. It may be pointed out that fluvioglacial evidence of glacier stagnation is extensive in association with the Menteith moraine complex of proved Zone III age. Abundant fluvioglacial features of presumed Zone III age indicate widespread glacier stagnation in Glen Roy and Glen Spean (Sissons, 1967b, pp. 119-122).

(iii) This criticism carries with it the implication that, since large corries have had only small glaciers, then valleys such as those of the Cairngorms cannot be expected to have contained valley glaciers since ice-sheet decay. Yet the evidence from the Lochnagar area shows

that small corrie glaciers and large valley glaciers existed contemporaneously. The authors suggest that this is not a problem and that the apparent difficulty arises from over-emphasising the importance of corries as glacier sources. Glaciers may develop on plateaux and also in valleys lacking corrie form: there is no corrie in the ground draining to Glen Muick. It may be pointed out that snow swept by wind from the rather small summit area of the Lochnagar ridge would be distributed among four corries: on the other hand snow blown from the high ground bordering Glen Muick, or Glen Callater, would, in each case, contribute to the nourishment of a single glacier. Alternatively, one may consider the catchment area for direct snowfall within the respective situations. Thus the floor of the large eastern corrie of Lochnagar is at 800 m; the total area of the corrie and adjacent ground within the corrie drainage basin above 800 m is 1.4 sq km. On the other hand, if one considers only upper Glen Muick (west from Loch Buidhe) the area of ground above 800 m is 8.7 sq km.* It is suggested that in view of such considerations it would be surprising if a glacier had not existed in Glen Muick contemporaneously with the Lochnagar corrie glaciers. The authors find it difficult to believe that the situation in the adjacent Cairngorms was significantly different.

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* Clearly such crude figures can be refined and qualified. For example, it is arguable that the SW.-facing valley slopes should be excluded from the Glen Muick calculation since they cannot be fairly compared with the NE.-facing slopes of the Lochnagar corrie. On the other hand, avalanching of snow from the SW.-facing slopes of Glen Muick would doubtless have helped to nourish the glacier below.

In the Lochnagar area the distribution of the corrie and valley glaciers implies renewed glaciation from upland sources and the steep gradients of the valley glaciers accord with this view. One would anticipate a similar situation in the Cairngorms. It is therefore relevant to note that in at least some valleys in the Cairngorms there are very clear limits to the hummocky moraines. For example, in the eastern Cairngorms a mass of fresh hummocky moraines comes to an abrupt end on the floor of the Gairn valley south of Loch Builg, after which, for several kilometres down-valley, accumulations of mounds of any sort are completely absent. In the western Cairngorms, significantly a few hundred metres beyond the limit of the area covered by Sugden's very detailed morphological map (Fig. 1, 1970), abundant hummocky moraines (and associated meltwater channels), whose upper limit declines down-valley, terminate abruptly and are succeeded by terraced outwash that extends for a kilometre along the valley floor. Outside this moraine limit the hill slopes completely lack similar features. Such evidence causes the authors to question Sugden's statement that "the Cairngorm equivalent of the Loch Lomond Zone III readvance is problematical" (1970, p. 215).

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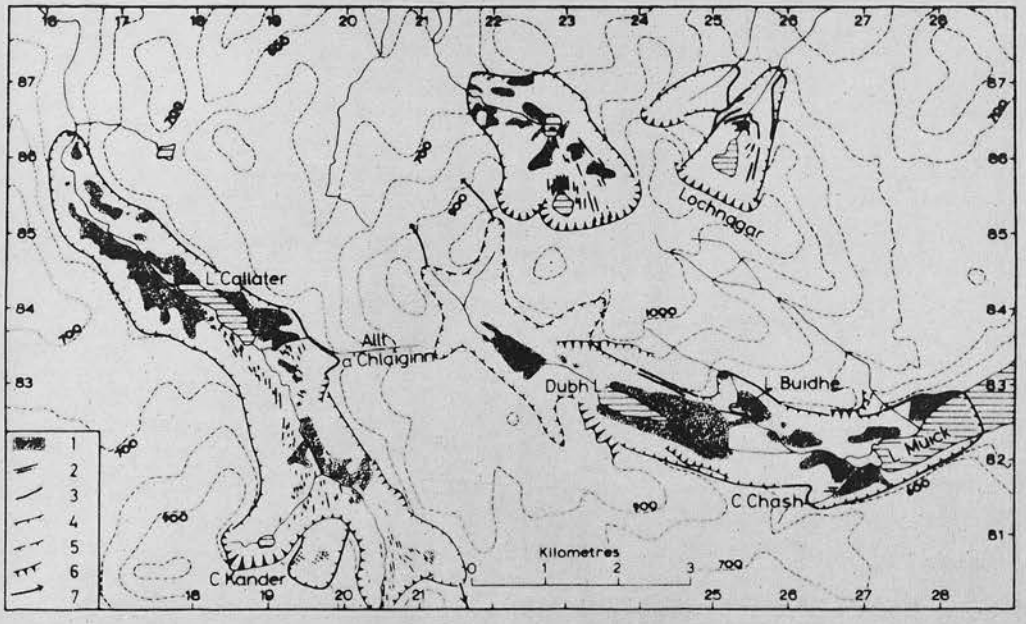
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Caption

Fig. 1. The last glaciers in the Lochnagar area. 1. Hummocky moraines. 2. Fluted moraines. 3. Lateral and terminal moraines. 4. Probable readvance limits. 5. Possible readvance limits. 6. Precipitous rock slopes. 7. Col meltwater channels. Contour interval 100 m.


L Callater
Allt a'Chlaiginn
Dubh L
Lochnagar
L Buidhe
L Muick
C Chash
C Kander
Kilometres
0 1 2 3
1
2
3
4
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6
7

The geomorphology of central Edinburgh

J. B. Sissons

With the rapid increase in borehole investigations for construction sites during the last ten or fifteen years a great deal of information has been (and is being) obtained concerning the composition and thickness of superficial deposits and the form of the buried bedrock surface. It often happens that very detailed information exists for particular sites in an area but that the sites themselves are too widely separated from each other for meaningful interpolation between them to be attempted. In the central areas of cities, however, owing to redevelopment schemes, the construction of modern urban highways, and the erection or extension of office blocks, large stores, educational institutions, etc., very detailed borehole information is likely to be available from numerous closely spaced sites. Such information has been largely neglected by geomorphologists yet it demands their attention, for not only may it assist in the interpretation of landform evolution in city centres but its correlation and plotting on large-scale maps can be of practical value in relation to future site investigations.

This paper is concerned with an approximately rectangular area in central Edinburgh measuring about 2.0 km by 2.8 km. Records for 1400 boreholes and trial pits[1] were collected for this area. Almost all these records are from site investigations made during the last 15 years, the great majority having been made during the latter half of this period.

Borehole and other information

Altitudinal data For the construction of a rockhead map altitudinal information is required. Fortunately the ground surface of the great majority of boreholes and trial pits in the area is recorded on the logs. Where altitudes were not given they were sometimes determined from detailed site plans giving levelled heights, but more often from the height information on Ordnance Survey 1:1250 plans coupled with knowledge of ground configuration. For a few bores where the ground surface changes rapidly in altitude and Ordnance Survey altitudes were inadequate, bore altitudes were determined by surveying aneroid on very short traverses in suitable weather conditions.

Plotting of data In the site investigation reports most of the bores and trial pits were located on large site plans, usually at scales between 1:96 and 1:500, although some were marked on 1:1250 plans and a few on 1:2500 plans. Those in the last category were plotted on Ordnance Survey 1:2500 plans, all the remainder being plotted on Ordnance Survey 1:1250 plans. Each bore was then given a 10 figure National Grid reference (i.e. to the nearest metre), it thereafter being a simple (though lengthy) task to plot any type of information from the records on squared paper. For this purpose it was found most convenient to use standard graph paper with squares of one-tenth of an inch, making an inch equivalent to 100 m on the ground. Figs 1 to 3 were all drawn at this scale, which is equivalent to just over 16 inches to the mile or 1:3937.

Sources of inaccuracy The site investigation reports contain occasional minor errors such as individual bed thicknesses that do not total to the full depth recorded as penetrated by a bore, incorrect ground surface altitudes, and incorrect scales on site plans. Such errors can often be rectified, although inevitably there must be some minor errors that cannot be corrected or even detected. Major errors are unlikely to escape notice, especially when the data are plotted. One major error at two sites is incorrect recording of rockhead in certain bores, an error caused by failing to penetrate the assumed bedrock (actually a boulder) for the recommended minimal $1\frac{1}{2}$ m.

In some bore logs it is difficult to decide where bedrock begins. On shales there is often a transition downwards from typical glacial till, through debris containing increasing amounts of shale, into badly broken shale and finally into firmer shale. This transition may occur over a metre or may sometimes take several metres. Rockhead was interpreted from the bore logs as the level below which rock apparently predominated over drift. Thus descriptions such as "broken sandstone", "weathered shale", and "sandstone with clay-filled joints" were interpreted as referring to bedrock whereas terms such as "clay with abundant shale fragments", "clay and igneous boulders", and "sandy clay with abundant subangular sandstone fragments" were considered to apply to drift. Since (i) the use of these and other descriptions cannot be consistent between different boring firms, (ii) there occur descriptions such as "clay with abundant shale fragments (weathered

shale)", and (iii) drilling methods may not allow stratum thicknesses to be accurately determined, rockhead cannot be regarded as precisely definable. It would perhaps be realistic to consider the rockhead altitudes from which Fig. 1 was constructed as accurate to within $\pm$ 1 m.

One source of error that has to be guarded against and that is peculiar to bore logs from built-up areas relates to the fact that many bores are put down from basement level. Such bores usually commence 1 to 4 m below the general level of the surrounding ground, although at one site in central Edinburgh they start as much as 10 m below. When a larger area is being investigated most bores will be at normal ground level but where houses (for example) have basement entrances it is often found that the borers have taken advantage of the artificially low ground for one or two bores. Basement bores clearly give a misleading indication of drift thickness in their vicinity. Consequently, in constructing Figs 2 and 3, drift thicknesses recorded in such bores were adjusted in relation to local ground level. Where old basements were excavated in bedrock bores put down from them will also give a misleading indication of the altitude of rockhead. Such a bore merely shows that rockhead is not below a certain altitude.

It must be pointed out that maps of drift thickness, drift composition and rockhead altitude in a built-up area, especially one of considerable antiquity like part of central Edinburgh, cannot be accurate in detail however abundant the borehole data. For example, beneath every road there is a complex of pipes and cables, for the

laying of each one of which a trench has been dug. The trenches are now filled with debris ("fill" or "made ground" in bore logs) but between the trenches the original rock or drift cover often remains. These and similar complications, related to old foundations, old wells, long-abandoned quarries, etc., mean that drift conditions are excessively complex in detail. For this reason areas of made ground are not shown in Fig. 3 and the key to the map is intentionally imprecise[2]. These complications affect Figs 1 and 2 only where the natural drift cover is thin.

In Fig. 2 many of the rather intricate lines indicating drift thickness are not supported by any evidence shown on the map. This is because, in addition to the borehole data, an infinite number of values for drift thickness can be obtained by subtracting mapped rockhead altitudes from present ground surface altitudes (the rockhead contours (Fig. 1) being drawn on the assumption that ice-moulded features in relatively weak sedimentary rocks such as typify central Edinburgh possess smooth forms, an assumption generally supported by the borehole evidence). By this means the great thickness of made ground (with some underlying natural deposits) in the Mound, linking the Royal Mile with Princes Street, is clearly brought out, despite the lack of borehole data. In constructing Fig. 2 account has also had to be taken of the fact that in some places the ground drops vertically for as much as 5 to 10 m through man's activities, as a result of which drift thickness may change abruptly. For example, this causes the abrupt changes along a straight line southwest from the Castle Rock.

Other information In addition to the borehole data, information from various other sources was used in constructing Figs 1 to 3. These include:

(i) Relevant Geological Survey six-inch sheets, which show areas where rock is known or presumed to crop out, as well as certain points at which rock has been exposed in temporary excavations at various times in the past. The rock outcrop information had to be used with caution, however, since in some areas boreholes have proved thick drift where the geological sheets show rock at the ground surface.

(ii) Published papers in which drift deposits or temporary rock exposures in central Edinburgh are described. The temporary rock exposures, where they could be accurately located, along with those marked on Geological Survey sheets, are distinguished separately on Figs 1 to 3. While not giving accurate information on the altitude of rockhead, they usually imply that it is near the ground surface.

(iii) Ordnance Survey 1:1250 and 1:2500 plans, which accurately depict most of the areas where bare rock is exposed.

(iv) Old plans of Edinburgh, which include information on old lake sites, abandoned quarries, etc.

(v) The geological section of the Scotland Street Tunnel, held in Edinburgh Public Library.

(vi) Sections in building sites, trenches, etc., examined by the writer.

Glacial erosion

Central Edinburgh is dominated by landforms produced by glacial erosion. Deposition in glacial and postglacial times has partly obscured some of the features but the borehole information enables these buried features to be mapped.

Little need be said about the crag-and-tail of the Castle Rock and Royal Mile, for it has long been known that the resistant volcanic plug of the Castle Rock has partly protected the sedimentary rocks on its lee side from glacial erosion, so that they now form a tail at least 1.4 km long whose crest declines constantly towards east-north-east. Deposits of till on the tail are thin and patchy and, in its natural state, the tail can be regarded as almost entirely solid rock. A considerable amount of debris deposited by man in various places, along with buildings, obscures the form of the tail, which, as far eastwards as the Bridges, is a steeper feature in rock than is often apparent today.

Calton Hill forms a much less impressive crag-and-tail, probably because of the considerable extent of the lava outcrop of which it is composed, coupled with the faults that control the alignment of its southern and north-western sides. On the other hand, Salisbury Crags and the long slope below them sweep round in a great curve that reveals the influence of ice-moulding, the hill as a whole pointing towards the direction from which the ice approached.

Around the western end of each of the three hills local overdeepening by ice has occurred. A considerable part of the curving

depression that sweeps round the base of Salisbury Crags and the steep slope below is visible and is followed by Queen's Drive, although north-eastwards the feature passes beneath the deposits of the former Holyrood Loch. The southern part of the trench is partly controlled by geology, for the St Leonard's Sill (corresponding with the 75-85 m ridge crest in Fig. 1) borders it on the west and south. On the south-western side of this sill there is further evidence of local overdeepening by the ice, boreholes indicating the presence of a small closed rock basin associated with deflection of the basal ice layers by the resistant sill.

The movement of basal ice north-eastwards along the north-western face of Calton Hill is indicated by striae marked on the Geological Survey six-inch sheet and by the trench aligned SW-NE shown in Fig. 1. The existence of this trench is established by many boreholes, although its north-eastern limit cannot be determined owing to an absence of bores. The axis of the trench, which descends to only about 15 m above sea-level at its lowest known point, lies beneath a pronounced surface hollow that is situated at the foot of the very steep north-western face of Calton Hill. It is possible that the interruption of the buried trench represented by the "bulge" in the 25 and 30 m rockhead contours is due to a mass of rock having slipped down into it from the adjacent hill side.

Calton Hill ends in a very steep rock slope on its southern side. This abrupt descent continues beneath the superficial deposits: thus a bore 10 m to the south of the base of the visible rock face

showed 14 m of drift while one 28 m from the rock face showed 29.6 m of drift. The latter bore proved rockhead at only 21 m above sea-level and marks the lowest known point of the closed rock basin that occurs south of Calton Hill. This basin indicates localised overdeepening of the trench that runs along the northern side of the Castle crag-and-tail by basal ice deflected into it by the resistant mass of Calton Hill.

Ice deflection is even more strikingly demonstrated at the base of the Castle Rock. The vertical and near-vertical sides of the volcanic plug that forms the Rock continue to descend rapidly beneath the drift deposits that surround its base. A bore 16 m from the base of the visible rock face revealed 12 m of drift while one 33 m from the rock face showed 29 m of drift. The latter bore, along with an adjacent one, proved rockhead at about 21 to 22 m above sea-level, more than 40 m below rockhead on the top of the George Street rise to the north and about 110 m below the summit of the Castle Rock some 200 m to the south-east. The pronounced trench revealed by these bores occurs beneath the surface depression that is occupied by Princes Street Gardens and Waverley Station. Since only a few bores have been put down in this depression the greatest depth of the buried trench is unknown and the rockhead contours shown in Fig. 1 merely indicate the probable general form of the trench.

Southwards from the bores specifically mentioned above, however, the form of the buried trench is well defined by a considerable number of bores. These prove that the trench sweeps around the base of the

Castle Rock in a great arc, the rock floor on the south-western side of the Rock being at an altitude of about 40 m. Here the trench lies beneath a surface depression followed by King's Stables Road. This surface depression continues east-northeastwards for almost the full length of the Royal Mile tail. The bores along this part of the depression, although not sufficiently numerous to allow a reconstruction of the detailed form of the buried trench, nevertheless strongly suggest that it is nowhere as pronounced as at the base of the Castle Rock.

Another area where glacial over-deepening has occurred is beneath and around the flat ground now occupied by the Meadows. As can be seen in Fig. 1, the 70 m rockhead contour just fails to delimit a closed depression in rock. The rock surface is lowest beneath the eastern part of the Meadows, where a depression contour for 65 m is shown, the lowest recorded rockhead altitude being 61 m. The lake that once existed at the Meadows thus owed its existence to the presence of a rock basin eroded by glacier ice.

The full depth of rock removed by glacial erosion in central Edinburgh cannot be calculated, for no identifiable fragment of the preglacial relief survives. One can, however, estimate the minimal amount of glacial erosion at various locations. For example, if, as suggested above, Salisbury Crags and the steep slope below owe their form to glacial moulding, then it is reasonable to infer that in preglacial times the Crags were a simple cuesta with the scarp face aligned parallel with the rock strike (i.e. nearly north-south). If

one makes the minimal assumptions that this scarp was situated no farther west than the present westernmost point of the Crags and did not exceed in altitude their present highest point, then, where now beneath the site of the former Holyrood Loch the rock surface is at only about 20 m, there would have been situated a cuesta attaining about 175 m. Thus one may argue that a thickness of over 150 m of rock has been removed by the ice at this location.

Since the sedimentary rocks on the lee side of the Castle Rock, where they have been relatively protected from glacial erosion, attain an altitude of 125 m, there appears to be no reason for not assuming that in preglacial times these rocks reached at least this altitude on all sides of the Castle Rock. Hence, on the northwestern side of the Rock, where rockhead is now at only about 20 m O.D. at one location, it may be inferred that at least 105 m of rock have been removed by glacial erosion. The full amount of erosion may be much greater than this, of course, for the Castle Rock itself may have been greatly lowered by the ice and the sedimentary rocks may have attained a mugh higher altitude in preglacial times than their present highest point.

Arguing along the same lines, one may consider the area lying northwest, west and southewst of the Castle Rock beyond the basal trench. Over much of this area the rock surface is gently sloping or more or less level (Fig. 1), and it may appear that there is no reason for postulating much glacial erosion here. Yet this area of sedimentary rocks lies mostly between 55 and 70 m, while similar

rocks were protected by the Castle volcanic plug reach 125 m. Hence over at least part of this area it seems likely that the minimal amount of glacial erosion is 55 to 70 m.

The George Street and Lauriston rises appear to have no direct geological explanation. Both are formed of sedimentary rocks unprotected by an igneous mass. The George Street rise is approximately in line with the summit of Calton Hill, while the Lauriston rise is almost in line with the apex of Salisbury Crags. This suggests that, in addition to local deflection of the basal ice around their stoss ends, these hill masses may have influenced the ice flow and hence glacial erosion a considerable distance upstream. In this context it may also be noted that the Meadows rock basin is aligned in quite a different direction from the Castle crag-and-tail and adjacent features, for it trends south of east. If this trend is continued it passes just south of the large resistant mass of Arthur's Seat, which rises to an altitude of 250 m a short distance outside the area represented in Fig. 1. Since there appears to be no local geological variations that could account for the divergent trend of the Meadows basin, it seems likely that this trend is due to the basal ice movement having been influenced by the mass of Arthur's Seat.

The evidence described above clearly indicates that the pattern of basal ice movement in central Edinburgh was greatly influenced by resistant rock outcrops. Variations in basal ice flow associated with irregular rock surfaces, especially as revealed by striae, have been studied by various workers (e.g. Demorest, 1938; Virkkala, 1951, 1960; Gjessing, 1965), although the writer does not know of any work in

which features on the scale of the deep curving trench around the base of the Castle Rock have been reported. It would seem that basal ice flowed down into this trench and along it to curve round the Rock, ice movement at the upstream end of the Rock being perpendicular to the general movement towards a direction slightly north of east that prevailed at a higher level in the ice-sheet. That local ice movement perpendicular to the general flow can occur on a very small scale was clearly demonstrated by Demorest (1938).

Glacial till

The superficial deposits in central Edinburgh vary greatly in thickness, ranging from zero to 30 m. They include till, sand and gravel, lake silts and clays, peat, solifluction deposits and made ground.

Over most of the area till is no more than a metre or two thick and is almost or completely lacking over much of the higher ground. It exceeds 8 m in small areas on the southern side of the deeper part of the Meadows rock basin and on the southern flank of the Lauriston rise, and exceeds 10 m locally in the St Leonard's Sill rock basin. The thickest till in the whole area is at the eastern end of the George Street rise, where it is in places interbedded with and/or overlain by sand and gravel. The greatest recorded thickness of a single layer of till here is 12 m but in one bore where till occurs as three layers separated by sand and gravel its total thickness is nearly 21 m.

In relation to ice movement this thick till occurs on the lee side of the George Street rise. However, the till does not continue into the trench that flanks the north-western side of Calton Hill. This trench and the trenches associated with the Castle Rock and tail, as well as the depression that curves round Salisbury Crags, contain very little till and many bores record none at all. This strongly suggests that these features were at least partly excavated by the last ice-sheet.

Fluvioglacial deposits

A belt of sand and gravel with a maximal known thickness of 9 m and a length of several hundred metres commences west of the Castle Rock and passes to the south of it. The sand and gravel is contained within the deep trench, covering its southern slope, and also lies on its floor beneath the soft deposits shown in Fig. 3. Since no significant stream has existed in this area in postglacial times the sand and gravel must have been laid down by a meltwater river. This is also indicated by the distribution of the sand and gravel, which is banked up against the north- and northeast-facing side of the trench, suggesting that glacier ice existed in the trench while much or all of the sand and gravel was being deposited. This suggestion is supported by the absence of a comparable belt of sand and gravel in the much deeper part of the trench that curves round the northern side of the Castle Rock. (For example, two bores in the trench northwest of the Castle Rock passed through 28 and 29 m of superficial deposits and the only deposits found that may be of

fluvial origin were 1-1½ m of "sand with traces of clayey peat and fine gravel" immediately above rockhead.)

The belt of sand and gravel may well have been deposited by a meltwater stream that cut an otherwise anomalous valley that runs SW-NE towards the Castle Rock. This valley is now almost obliterated but a small part of it can still be seen behind the Caledonian Hotel. It is shown by a form of hachuring on Robert Kirkwood's 1817 plan, where its southern side is marked as continuous with the southern side of the depression that is situated south of the Castle Rock (thus suggesting a relation between the valley and the sand and gravel belt).

Eastwards from the sand and gravel belt the meltwaters probably followed the Grassmarket-Cowgate trench. Near the eastern end of this trench closely-spaced boreholes prove a buried valley cut 15 m into rock (Fig. 1). The sides of this valley are very steep, the northern side being proved to slope at an angle of more than 45 degrees. This contrasts greatly with the adjacent gentle rockhead slope of the Royal Mile tail and strongly suggests meltwater incision into the floor of the ice-moulded depression. The 8 m sand with some gravel that have been recorded in the bottom of the valley may be fluvioglacial deposits.

The other principal area in central Edinburgh where there was considerable fluvioglacial activity is at the eastern end of the George Street rise and in the adjoining trench on the northwestern side of Calton Hill. Although these two areas merge into each other

their deposits differ considerably. In the trench the great majority of bores show no glacial till, but almost all show sand and gravel (up to 15 m thick), usually resting directly on bedrock. On the adjacent higher ground to the west till, often thick, is recorded in all bores and is often covered by sand or by sand and gravel, or is interbedded with these deposits. In this area Fleming (1859) observed that a layer of large boulders occurred in one plane in the till, dipping east at about 15 degrees. He also stated that he saw in several excavations in the area an upward transition from till through sand and gravel to fine stratified sand. Tait (1939) described an excavation in which till $4\frac{1}{2}$ to 9 m thick had through its whole thickness vertical fissures about 1 m wide with irregular but roughly parallel sides, the fissures being filled with sand and a little gravel. The origin of these fissures is unknown. The location of much of the sand and gravel on relatively high ground, its interbedding with till, and the gradation from till to sand and gravel demonstrate its fluvioglacial origin.

The areas of considerable fluvioglacial activity are closely associated with the ice-eroded trenches. Other evidence of fluvioglacial deposition not considered above also occurs in such locations. However, it is not suggested that glacial river erosion contributed significantly to the excavation of the trenches since (i) the form of the trenches as a whole appears inconsistent with river erosion, (ii) where both till and fluvioglacial deposits occur in the trenches the former lies beneath the latter, implying glacial formation of the trenches followed by fluvioglacial activity, (iii) in the one location

where there is good evidence for fluvioglacial erosion the channel thus formed is cut sharply into the ice-moulded form. Rather is it suggested that meltwater rivers utilised some of the features already produced by glacial erosion.

Divergence of the basal ice as it approached Calton Hill and the Castle Rock may have been responsible for part of the courses of the two major glacial rivers being located directly up-ice from the stoss ends of these features.

Lacustrine and related deposits

Evidence of former lake sites in central Edinburgh is provided by areas of silt and clay, which are often associated with peat. The areas where these soft deposits are known or believed to be 2-3 m or more thick are shown in Fig. 3, which thus indicates the probable minimal extent of the former lakes and associated areas of badly-drained ground.

Holyrood Loch The deposits of the former Holyrood Loch extend up to a fairly constant maximal altitude along their western margin, values between 37.5 and 39.0 m being recorded in 12 bores. Descriptions of the deposits in bore records include "clay", "sandy clay" and "organic sandy silty clay", while peat nearly 2 m thick has been encountered. Henderson (1888) described the upper deposits as exposed in a trench in Queen's Park south of Holyrood House, where they included peat, peaty clay, clay, and marl with abundant freshwater shells. By far the most interesting records, however, are the descriptions by Taylor

(1888) and Cadell (1893) of the section revealed by a gasometer excavation 47 m in diameter and averaging 14 m deep. On the bedrock exposed near the base of the section was a bed of marl 0.6-1.7 m thick containing freshwater shells and aquatic plant remains. This was succeeded by 0.3-0.8 m of peat containing wood fragments. Above this was a layer, perhaps 2-3 m thick, of "till" or "rough stony clay" containing boulders up to 0.6 m in diameter. This was in turn overlain by a series of beds including marl, sand, clay and peat. It can be inferred that the bed of stony clay is a solifluction layer from the steep slope of Salisbury Crags to the south accumulated under cold climatic conditions (Sissons, 1967, p. 222). The following reasons point to this conclusion. (i) Taylor stated that the stones were mainly of rock types found on Salisbury Crags, while Cadell (writing at a time when the significance of periglacial processes was not appreciated) concluded that the stony layer had been "probably washed down by torrents from the adjoining steep talus of Salisbury Crags". (ii) Taylor pointed out that the stones were essentially angular and that the layer was distinctly bedded. (iii) The distance of the section from the base of Salisbury Crags scree slope along with the data given by Taylor and Cadell show that the gradient of the stony layer is between 4 and 5 degrees, which contrasts with the present 35 degree scree slope below the crags. The low angle mass movement of this coarse debris indicates solifluction when a periglacial climate prevailed. The most likely time for this is Zone III, implying that lateglacial lake deposits exist beneath the coarse layer and postglacial ones above it.

Cowgate depression Hugh Miller (1864) referred to a South Loch whose former existence was proved by beds of silt, impressions of aquatic plants and layers of lacustrine shells in the valley now occupied by the Cowgate and Grassmarket. Cadell (1893) stated that there were no conclusive data for the existence of such a loch, suggesting that there were a few marshy pools. MacLaren (1893) and Tait (1939) reported excavations in the floor of the trench east of the Grassmarket showing marl, "blue clay", "black mossy earth" and "black mossy soil". MacLaren concluded that the deposits were thickest some 10 to 20 m north of the Cowgate and Tait mentioned a number of old buildings on the north side of the Cowgate that were supported on wooden piles, which in one instance at least went to "a considerable depth". The latter also stated that the piles were driven into city rubbish, which is of considerable thickness in places, a point strikingly demonstrated by the discovery of an upright water butt embedded in clay at a depth of 5-6 m.

Only a few boreholes have been put down on the floor of the Cowgate depression, the main concentration being those in the east that reveal the meltwater channel referred to previously. These bores show that the sand with some gravel that occupies the lower part of the channel is covered by lake deposits up to 4 m thick recorded as clay or silty clay containing organic matter. A bore farther west on the floor of the depression (at the National Library) proved sandy clay at least 6 m thick. The limited evidence available accords with the view of MacLaren and Tait that the soft deposits are thickest on the north side of the Cowgate, while altitudinal considerations tend

to favour Cadell's opinion that marshes and pools existed rather than a single lake.

Nor' Loch Only a few bores have been put down in the area of the old Nor' Loch, which occupied the depression that lies north of the Castle crag-and-tail. A bore near the western end of Waverley Station revealed 3.0 m of organic silt and clay and a layer of peat 2.7 m thick. Westwards the soft deposits increase in thickness and a bore close to and northwest of the Castle Rock shows that they attain the remarkable thickness of 21.3 m. The log of this bore may be summarised as follows:

	m	m
Fill	6.6	6.6
Very peaty silty clay and peat, with wood fragments in upper part	21.3	27.9
Fine sand with traces of clayey peat, and fine gravel	1.2	29.1
Sandstone (bedrock)	1.5	30.6

Silt, presumably deposited in the Nor' Loch, occurs west of the Castle Rock, where it was seen in the excavations for the railway tunnel (Cadell, 1899). Silty clay and peat up to 3 m in thickness also extend along the depression on the southern side of the Castle Rock (Fig. 3). Thus the original Nor' Loch may have been of considerably greater extent than has been previously assumed. However, the surface of the soft deposits on the southern side of the Rock is at a considerably higher altitude (50-58 m) than on the northern side (42-47 m). It is therefore possible that there were two lakes separated by a drift barrier not yet revealed by bores.

Burgh Loch This former lake coincided approximately with the large elliptical area of flat ground known as the Meadows, although the present extreme flatness is due to a metre or two of debris deposited and levelled by human agency. Beneath this made ground 2 to 3 m of lake deposits are recorded in some of the bores in the western part of the Meadows, but in the eastern part, where the rock basin is deepest, silty organic clay has been proved to a depth of 9 m below the ground surface.

The vicinity of Calton Hill Numerous bores in the depression immediately northwest of Calton Hill reveal lake deposits overlying sand and gravel and underlying made ground. The lake deposits vary in thickness and may attain about 10 m. They include silt and clay with freshwater shells, along with a peat layer up to 2 m thick. Since Greenside Row runs along one edge of this former lake, the latter may be called Greenside Lake.

A considerable number of the bores put down in the depression northwest of Calton Hill encountered boulders and cobbles of basalt along with some of sandstone and mudstone[3]. In one bore a boulder layer was penetrated for 9 m without reaching its base. On the south side of Calton Hill, extending to a maximal distance of about 100 m from the very steep rock slope, there is an area of boulders and cobbles of basalt mixed with sandstone fragments. The maximal thickness recorded for this boulder layer is 22 m. The location and composition of the two boulder deposits implies mass movement of weathered material from Calton Hill. The size and angularity of the rock fragments and the apparent scarcity of fines indicate frost

shattering, while the large volume of the material and the evidence for periglacial solifluction into nearby Holyrood Loch strongly favour the view that the deposits largely or entirely accumulated when a periglacial climate prevailed.

Made ground

Material deposited by man, or thoroughly disturbed by his actions (e.g. material put back into an excavation), covers a large part of central Edinburgh. Often it is only a metre or two thick and may result from pipe and cable laying and the digging of shallow foundations for buildings, while on the Meadows it was presumably deposited to cover poorly-drained ground. In certain areas, however, made ground attains a considerable thickness, especially in natural depressions and on some steep slopes. Debris tipped into the depression on the south side of the Castle Rock was found to attain a thickness of 13.7 m. The depression northwest of Calton Hill has been considerably infilled by man (the maximal recorded thickness of fill is 8.9 m), while the depression on the northern side of the Royal Mile tail contains thick deposits of made ground beneath Princes Street Gardens and beneath and near Waverley Station (11.6 m recorded). The broad flat top of the Royal Mile tail in the vicinity of St. Giles Cathedral is clearly artificial and a section published by MacLaren (1833) showed made ground up to 9 m thick. On the northern slope of the tail 16.9 m of made ground was proved by one bore (Royal Bank of Scotland). This great thickness is related to the extensive tipping carried out to produce the Mound, the large embankment that links the

Royal Mile with Princes Street. No bores are known to have been put down through the Mound itself, but its approximate thickness can be readily calculated. The greatest depth of made ground here is about 22 m.

Foundation engineering applications

While Figs 1 to 3 are of relevance to studies of landform evolution they are also relevant to the construction of foundations for buildings and other structures. Fig. 3 in particular was devised to show aspects of drift geology that are significant in this context.

Figs 1 to 3 can be of assistance in planning and costing future site investigations. In areas where there is already much borehole evidence and considerable uniformity in the drift deposits it should be possible to carry out less detailed (and therefore less expensive) site investigations than would otherwise have been made. For example, the boreholes show that glacial till (with some made ground) covers large areas of central Edinburgh (Fig. 3). This till is normally a good foundation for buildings. If Fig. 3 is considered in conjunction with Figs 1 and 2 it can be inferred that over considerable areas bedrock underlies the till at limited depth. On the other hand, Fig. 3 shows areas of soft deposits, where foundation problems may arise and piles may be required. Fig. 2 shows that these soft deposits often coincide with areas where the drift is thick or very thick. Since much or all of the drift is poorly consolidated in these locations piles of considerable length may be needed. Such information may help to avoid expensive mistakes such as have been made in the past.

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Notes

[1] Since some boreholes and trial pits are very closely spaced (e.g. 2 or 3 m apart) it is not possible to show all of them on Figs 1 to 3. The great majority of the dots in these figures mark the sites of boreholes, a very much smaller number indicate trial pits, while a few indicate sections measured by the writer or accurately recorded in the literature. For brevity all these data points are simply referred to as boreholes in the text.

[2] The areas shown in Fig. 3 as till over 2-3 m thick resting on bedrock include areas where made ground exceeding this thickness rests directly on bedrock since it is often impracticable to map such areas separately.

[3] This boulder deposit is not shown on Fig. 3 since it is usually overlain by soft lake deposits.

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- Borehole or other data point
- Rock reported in excavation
- Exposed rock

200m National Grid

Rockhead contours at 5m interval

George Street

Calton Hill

Holyrood

Queens Drive

Salisbury Crags

CASTLE

Grassmarket

Lauriston

Cowgate

Royal Mile

The Meadows

Fig.1.

· Borehole or other data point

× Rock reported in excavation

Exposed rock

200m National Grid

Drift thickness: 4m contour interval

Calton Hill

Holyrood

George Street

Royal Mile

Cowgate

CASTLE

Grassmarket

Lauriston

The Meadows

Queens Drive

Salisbury Crags

Fig. 2.

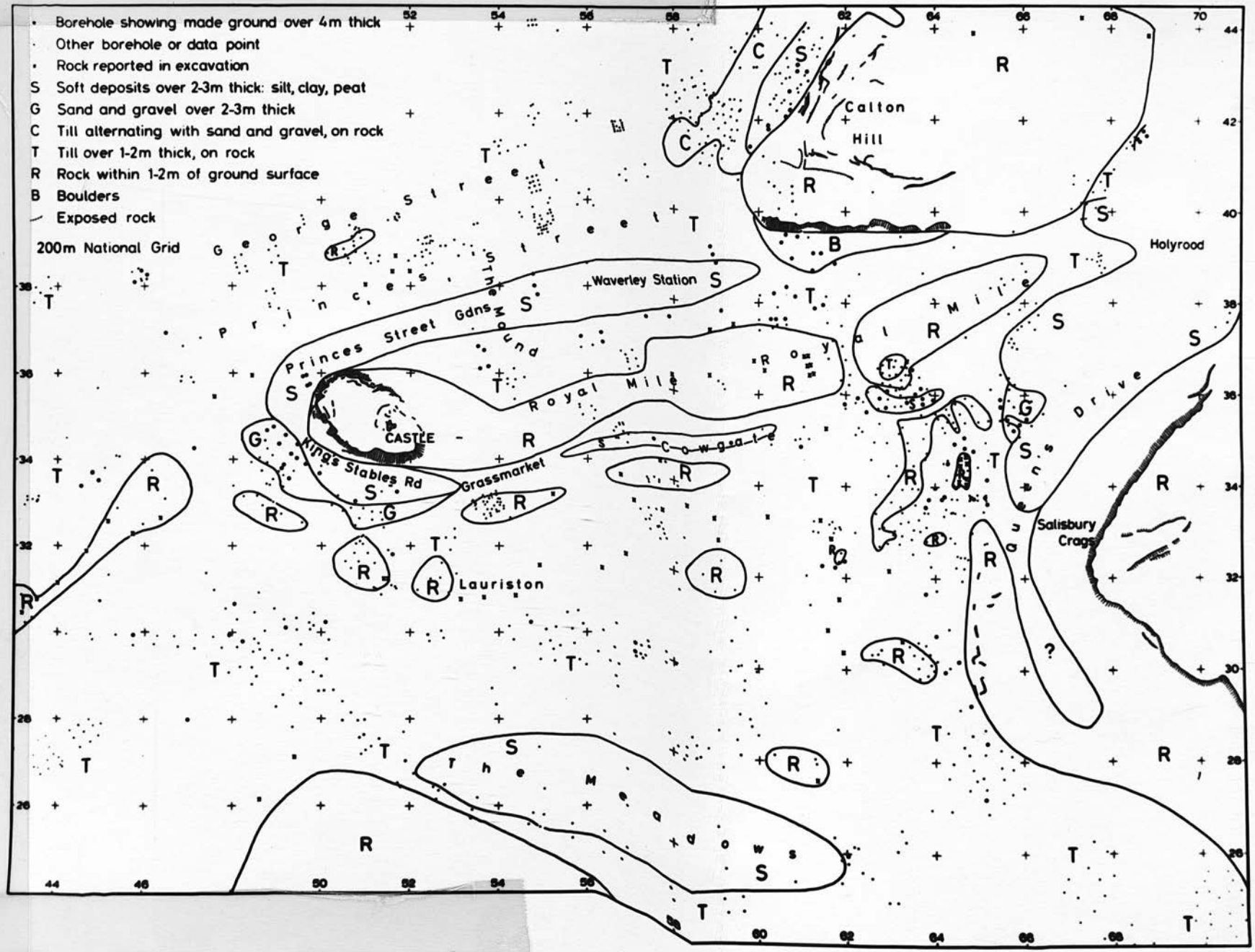

Borehole showing made ground over 4m thick
Other borehole or data point
Rock reported in excavation
S Soft deposits over 2-3m thick: silt, clay, peat
G Sand and gravel over 2-3m thick
C Till alternating with sand and gravel, on rock
T Till over 1-2m thick, on rock
R Rock within 1-2m of ground surface
B Boulders
Exposed rock
200m National Grid
George Street
Princes Street
Princes Street Gdns
Waverley Station
Calton Hill
Holyrood
Royal Mile
CASTLE
Kings Stables Rd
Grassmarket
Cowgate
Lauriston
Salisbury Crags
Queens Drive
The Meadows

Fig. 3.