THE DEGLACIATION OF THE EAST CHEVIOT AREA, NORTHUMBERLAND.

CHALMERS M. CLAPPERTON

Thesis submitted to the Faculty of Social Sciences, Edinburgh University, for the degree of Doctor of Philosophy.
June, 1967.

TABLE OF CONTENTS

ACKNOWLEDGEMENTS

On numerous occasions during the period of research leading to the production of this thesis the writer has received invaluable assistance from various people. A grant of money was provided by the Department of Scientific and Industrial Research during the first year of research. Informal discussions on glacial geomorphology were enjoyed with colleagues from the Geography Departments of Aberdeen, Edinburgh and Glasgow Universities. Acknowledgement is made to the staff of the air photographs library in the Scottish Development Department and to the Institute of Geological Sciences (Northern Division). Mr. S. E. Durno of the Macaulay Institute for soil research very generously spent four days with the writer investigating peat formations near Wooler; his enthusiastic interest on that occasion and subsequently is greatly appreciated. The writer is indebted to the technical staff of the Aberdeen University Geography Department for printing photographs and duplicating maps, and to Yvonne Wilson for typing the thesis. At all stages of the work, the unlimited encouragement and field assistance of Morag Clapperton has been deeply appreciated; thanks are due also to Allan Clapperton for help in levelling. Finally, the writer is particularly grateful to his supervisors, Professor J. Wreford Watson and Dr. J. Brian Sissons for enthusiastic guidance and stimulating discussion throughout the period of research.

LIST OF FIGURES

INTRODUCTION

Methods of Field Investigation

Those parts of the east Cheviot area appearing on vertical aerial photographs (scale approximately 1:10,000) held by the Scottish Development Department were studied stereoscopically as a prelude to work on the ground. The glacial landforms were mapped on a transparent overlay and then transferred to Six-inch maps (Ordnance Survey Provisional Edition) after approximate corrections for scale and distortion had been made using a Grant Projector. The drift boundaries published by the Geological Survey on the One-inch Drift maps of the area were copied and inserted on the Six-inch maps. The latter were taken into the field so that amendments could be made where necessary as the entire area was systematically investigated on foot. Since it was considered necessary to produce accurately levelled profiles on one of the more extensive landforms, a Watts microptic level (SL100) was used. All of the natural and artificial sections, together with the majority of minor exposures were systematically inspected to establish the nature of the drift and of the underlying bedrock. Since the sections are generally poor and small in number, a 3-foot soil auger was used to provide supplementary drift information in certain places, and all available bore-hole records for the area were consulted. Where good till sections were exposed in sites that appear to be relatively undisturbed by mass movements, the preferred dip and orientation of stones were measured with a Suunto compass in an attempt to establish the former directions of ice movement. The following procedure was adopted. Sites were chosen at least 3 feet below the ground surface, and after a vertical face had been cleared, a trench was dug inwards for at least one foot. This was to avoid all possible interference by plants and animals. At every site the dip and compass orientation of 100 stones were measured and recorded. Each stone selected for measurement possessed axes that were in the ratio of at least 1:2. Stones smaller than 1 cm. and larger than 20 cms. were avoided. Contiguous particles and particles with long areas dipping in excess of 60° were rejected, as were those in the vicinity of large boulders. The lithology and shape of each stone were also recorded. The measurements were plotted on polar equal nets (Figure 1.1, Chapter 1), on which the radial scale represents dip and the circumferential scale azimuth (Kirby 1961). The resulting scatter diagrams of axial plots give a good visual impression of the fabric characteristics.

The analysis of peat layers occurring in close juxtaposition with fluvioglacial phenomena necessitated techniques of investigation beyond the writer's scope. For this reason, pollen analyses of the peat were undertaken by S. E. Durno who also produced the pollen diagram presented with this thesis.

Although levelling and fabric analyses provided useful information, it should be emphasised that the majority of conclusions reached in the following chapters are based essentially on the analysis of morphological maps. The density of fluvioglacial phenomena in many parts of the east Cheviot area is so great that a valid interpretation of the deglaciation could not have been attempted if the complex interrelationships of the landforms had not first been plotted on the Six-inch scale. The landforms were later also plotted on the scale of 1:25,000, so that the overall patterns could be more readily discerned. The latter are presented in the map pocket accompanying this volume and are referred to throughout the text as Maps 1 to 11. The list of symbols used on these maps is also located in the map pocket. Where it has been necessary to illustrate the landforms in more detail, figures drawn to a scale larger than 1:25,000 and photographs are included with the text.

Photograph a The Cheviot massif viewed from the south-west.
By courtesy of Aerofilms Ltd.

LOCATION OF THESIS AREA

Figure 1

<u>The Topography</u>

The area referred to in this thesis as the Cheviot massif is located at the eastern extremity of the prominent range of uplands known as the Cheviot Hills. Aligned predominantly from south-west to north-east, the Cheviot Hills form the southern limb of the amphitheatre of high country that girdles the Tweed drainage basin in the Southern Uplands of Scotland (Figure 1 and Map 12). Rising occasionally to heights of over 1,800 feet, the Cheviot Hills form an important watershed, dividing streams draining generally south-east and east into the North Sea from those flowing northwards into the Tweed. Because the eastern extremity of the Cheviot Hills range is a structural entity, giving rise to quite distinctive topography in places, it has been considered as a separate unit and termed the Cheviot massif. It covers an area of approximately 250 square miles.

From marginal foothills, 700 to 900 feet high, the hill summits in the almost circular massif gradually attain greater heights towards the centre, ultimately culminating in the broad dome of The Cheviot (Map 5 and Photograph a). While the highest point on this mountain lies at 2,676 feet, there is almost a square mile of ground above 2,500 feet. Whereas the western and southern flanks of the massif have been heavily dissected into steep-sided rounded hills, the northern and eastern parts contain much ground that is gently sloping, with broad areas of plateau land between the deep valleys. Steep-sided hills do occur on the northern and south-eastern peripheries, however; for example, Yeavering Bell, Akeld Hill, Harehope Hill and Humbleton Hill in the north, and the hills about Fawdon in the south-east (Maps 5 and 8). Smooth, grassy hillsides predominate over much of the massif, for outcrops of bedrock are rare, occurring chiefly as tors and crags, and as precipices in the Bizzle valley and Hen Hole. The higher summits are frequently covered with blanket peat, the ragged edges of which are slowly wasting away. The black fringes of the dissected peat and the brown heather/bilberry association on the surface form a striking contrast with the bright green and yellow slopes below. Considerable depths of peat are known, for up to 20 feet has been reported from the broad summit of The Cheviot (Carruthers et al. 1932). The massif is drained by a fairly dense network of streams, many of which flow off The Cheviot to compose a pseudo radial pattern. These generally become principal streams; others rise in various places on the massif and become tributary to the larger rivers. Most of the streams are deeply entrenched within narrow, steep-sided valleys, but a few occupy broad embayments and appear to be quite unrelated to the topography; for example, the Threestone Burn (Map 5).

The Cheviot massif is bounded on the north-west and north by the broad lowland of the Merse, beyond which rise rolling foothills that eventually culminate in the Lammermuirs. Curving gently round the eastern perimeter of the massif, a series of low-lying basins is situated between its lower slopes and the assemblage of pronounced escarpments a few miles to the east. The basin floors are mostly below 300 feet. Arranged concentrically around the eastern flanks of the Cheviot massif, the escarpments are disposed en echelon in places and rise steeply to between 500 and 800 feet on the north and east. Their precipitous slopes rise to over 1,000 feet in the south-east, however. The rivers Aln and Coquet have breached the escarpments to reach the North Sea more directly than the river Till which flows northwards between the dip-slope of a minor cuesta and the scarp-foot of one to the east; the river then turns abruptly westwards to break through the ridge before draining northwards along the scarp-foot towards the river Tweed. The belt of lowland that fringes much of the Cheviot massif is absent on its south-west margin, so that a distinct topographical break between the massif and the range of hills to the south-west

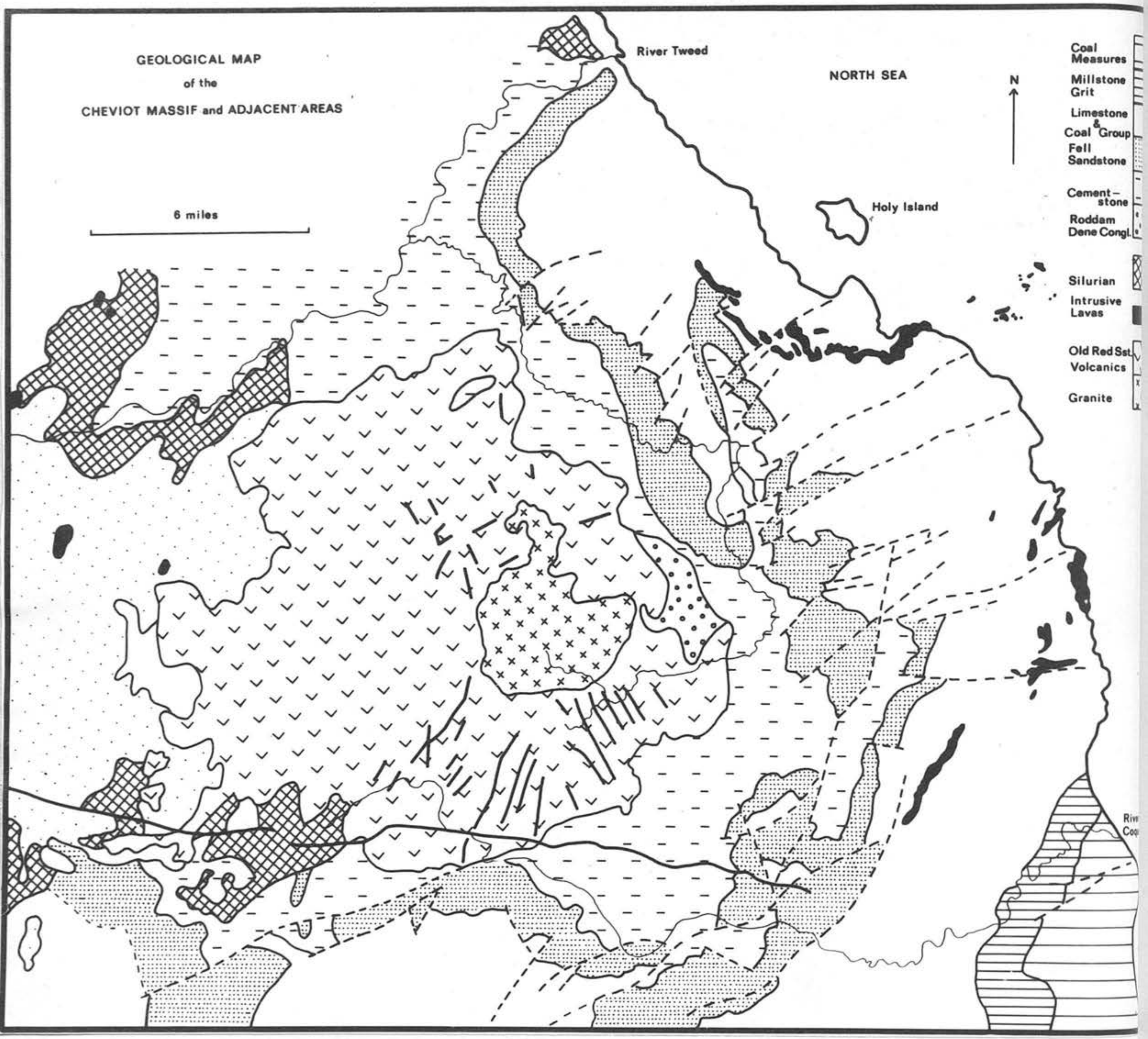

GEOLOGICAL MAP
of the
CHEVIOT MASSIF and ADJACENT AREAS
6 miles
River Tweed
NORTH SEA
N
Holy Island
Coal Measures
Millstone Grit
Limestone & Coal Group
Fell Sandstone
Cement-stone
Roddam Dene Congl.
Silurian
Intrusive Lavas
Old Red Sst. Volcanics
Granite

Figure 2

is generally lacking. Nevertheless, the hill summits south-west of the Cheviot massif are much more tabular and a closely spaced network of deeply incised streams similar to that characteristic of the massif does not exist.

Structure and Lithology

The basic structure of the Cheviot massif is relatively simple, consisting chiefly of an almost circular expanse of andesitic lavas extruded during the Lower Old Red Standstone period (Figure 2). Covering an area of approximately 250 square miles, the lavas are entirely subaerial in nature and consist of a variety of types including mica-felsite, glassy or pitchstone andesite, oligoclase-trachyte and augite-hypersthene-andesite (Carruthers 1931). Since much of the rock is porphyritic it has figured as a useful indicator stone in the glacial drift of England as far south as Cambridge. The composite volcanic cone built up by prolonged eruptive activity was compared in its reconstructed extent to Mount Etna in Sicily by Garwood (1922). Subsequent to the phase of volcanicity, a large mass of pink augite-granite appears to have formed possibly in the former magma chamber that once fed the volcano. This characteristic granite presently outcrops over an irregularly shaped area of approximately 22 square miles, slightly east of the centre of the massif. Dykes cross the granite and abound in the lavas, particularly on the south, and were apparently intruded along master joints and crush lines during a later stage of the Lower Old Red Sandstone period. They seem to be arranged in two principal swarms, running north-north-east - south-south-west and north-north-west - south-south-east. Faults, crush lines and shatter zones occur in considerable numbers in the centre and south of the massif, where they are aligned chiefly from north-east to south-west and from north-west to south-east. Sections of the Harthope and Breamish valleys are clearly related to these structures.

The broad belt of low-lying country partially enclosing the Cheviot massif is composed of sedimentary strata. In the Merse, these are principally sandstones of the Upper Old Red, overlain unconformably by rocks of the Cementstone Group of the Lower Carboniferous. The latter occur chiefly in the east. The Cementstone rocks also underlie the broad basins marginal to the eastern foothills of the igneous massif. A wide variety of Lower Carboniferous rocks, including those from the Limestone Group, the Scremerston Coal Group and the Fell Sandstone Group occur to the east and it is the massive sandstone of the latter group which has produced the impressive escarpments facing towards the Cheviot massif. The sedimentary rocks are arranged in a series of asymmetric anticlines and synclines, partly off-set by faulting and generally pitching northwards. These structures have evidently been important in the development of the cuesta forms.

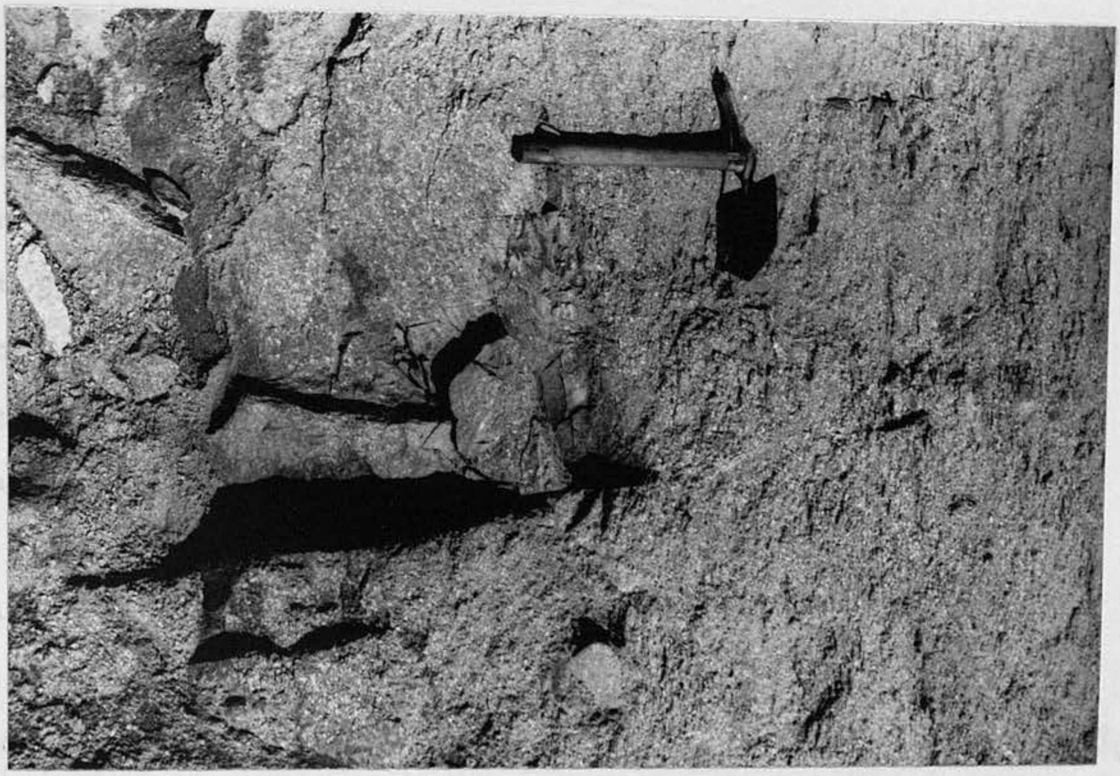
In view of the detailed accounts of the area's geological history by Garwood (1922), Hickling (1931), Carruthers (1931, 1932) and Common (1953) and a recent summary of the events by Clapperton (1963), it is considered that a repetition of such geological work is unnecessary; it would also be irrelevant to this thesis.

Evolution of the Relief

The structural arrangement of the area is believed to be predominantly Hercynian in origin, but the present relief was presumably fashioned during and subsequent to the phase of late Tertiary/early Pleistocene uplift that affected much of Europe. Whether or not any elements of the relief are remnants of ancient topography re-exhumed from beneath Mesozoic or Tertiary sediments is impossible to determine. Many of the major streams flow in narrow valleys, the sides of which rise precipitously for between 300 and 600 feet. Although

Photograph b — Deeply rotted bedrock, left bank of the New Burn.

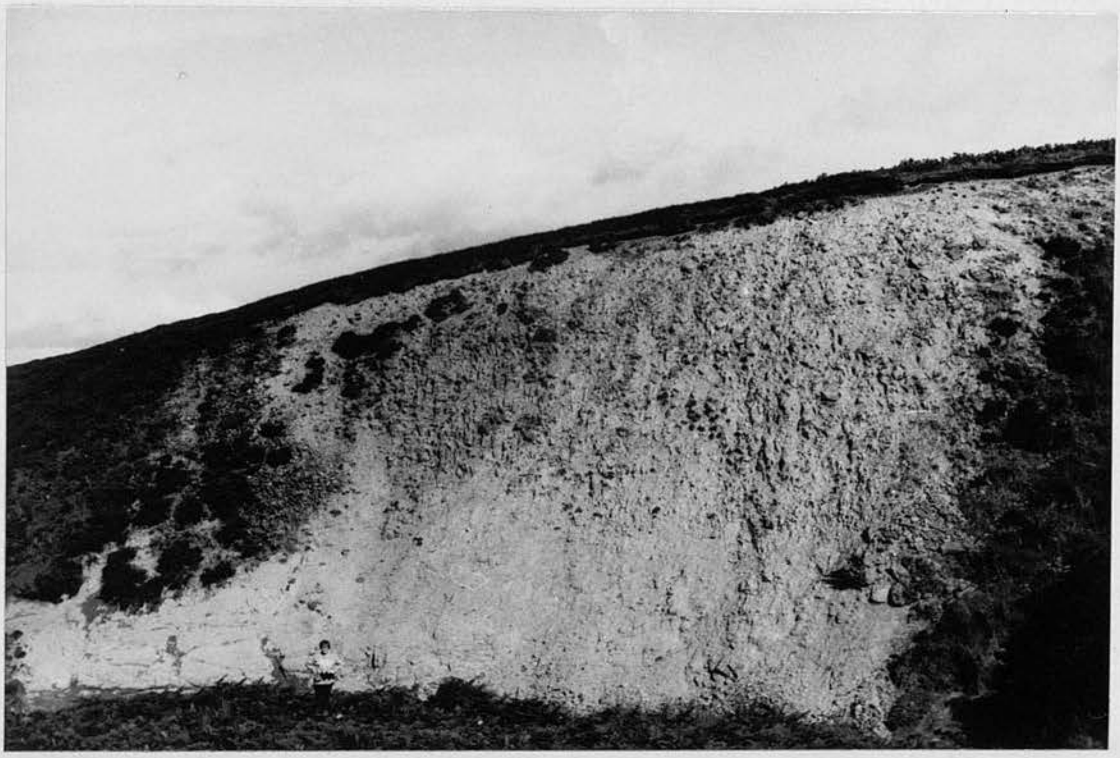
Photograph c — Deeply rotted bedrock, right bank of the Broadstruther Burn.



fluvioglacial rivers probably contributed in some degree to the development of these valleys, much of their present form is preglacial. This suggests that the latest phase of uplift was accomplished fairly rapidly, because valley widening by slope retreat has not progressed very far. The well-defined Cementstone basins and the Fell Sandstone escarpments were probably etched out into relief consequent upon stream incision and slope retreat following the uplift. Concerning the older elements of the relief, Common (1953, 1954) recognised no fewer than eight separate erosion surfaces in the Cheviot massif, and observed that "the relief rises in terrace-like fashion into the core with "treads" separated by "risers" of about 200' height". The accordance of summit levels in some parts of the massif is indeed striking, but it is difficult to recognise the entire suite of erosion surfaces mentioned by Common.

In Scotland, the occurrence of relief forms remarkably similar to those normally developed by morphogenetic processes in semi-arid environments has recently been recognised by Fitzpatrick (1963), Godard (1965) and Walton (1966). In the Cheviot massif many of the somewhat isolated conical hills, the broad cols and embayments, and residuals rising above plateau surfaces (e.g. Hedgehope Hill) possibly owe much of their form to similar processes operating mainly in the late Tertiary. Deeply weathered bedrock is also widespread in Scotland, and has been interpreted by Fitzpatrick (1963) as the product of profound chemical weathering before the Pleistocene. In this respect, it is interesting that deeply rotted bedrock also occurs in several places within the Cheviot massif. The exposures are mostly located above 1,000 feet, in the upper reaches of several streams. Over 50 feet of this material is present on the left bank of the New Burn and 20 to 30 feet commonly occurs along the Broadstruthers and Harthope Burns (Map 5, Photographs b and c). That this soft, growan material is not the product of Pleistocene solifluction is indicated

Photograph d	Deeply rotted bedrock passing upwards into broken, but relatively fresh bedrock, right bank of the Broadstruther Burn.
Photograph e	Deeply rotted bedrock passing upwards into broken, but relatively fresh bedrock, right bank of the Broadstruther Burn.



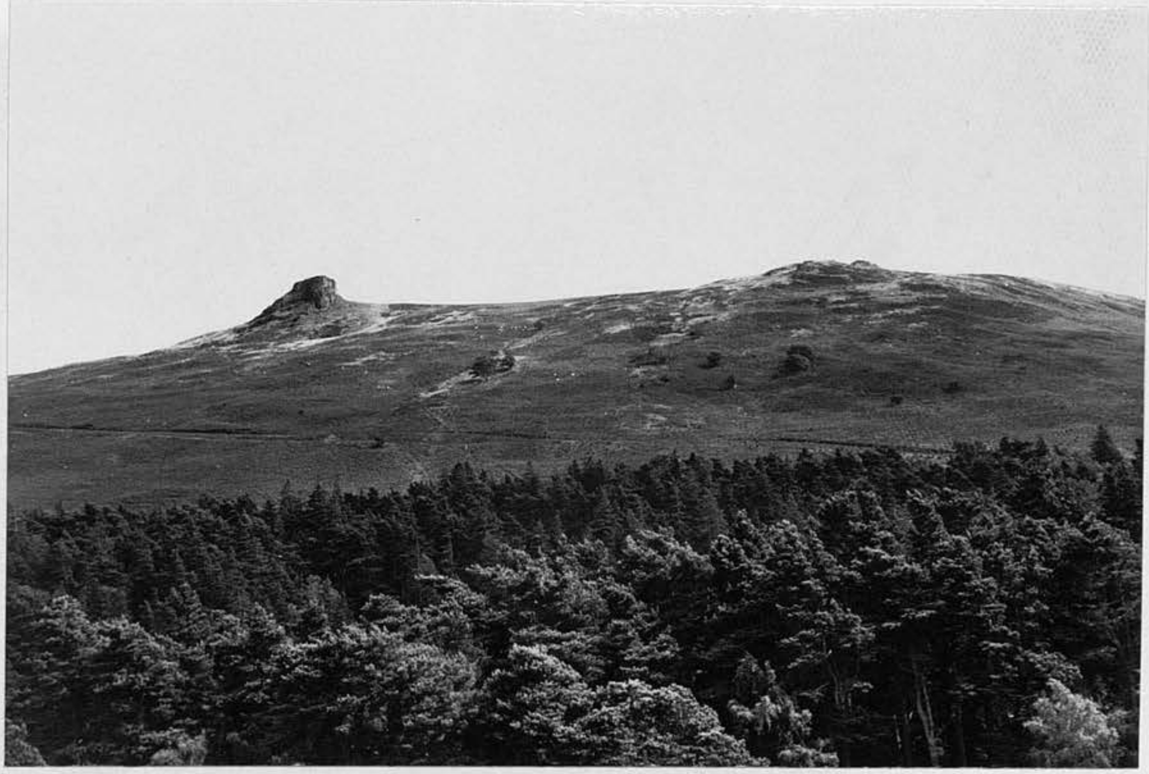
Photograph f   Deeply rotted bedrock (Andesite) near Old Middleton.

Photograph g   Deeply rotted bedrock (Andesite) near Old Middleton.

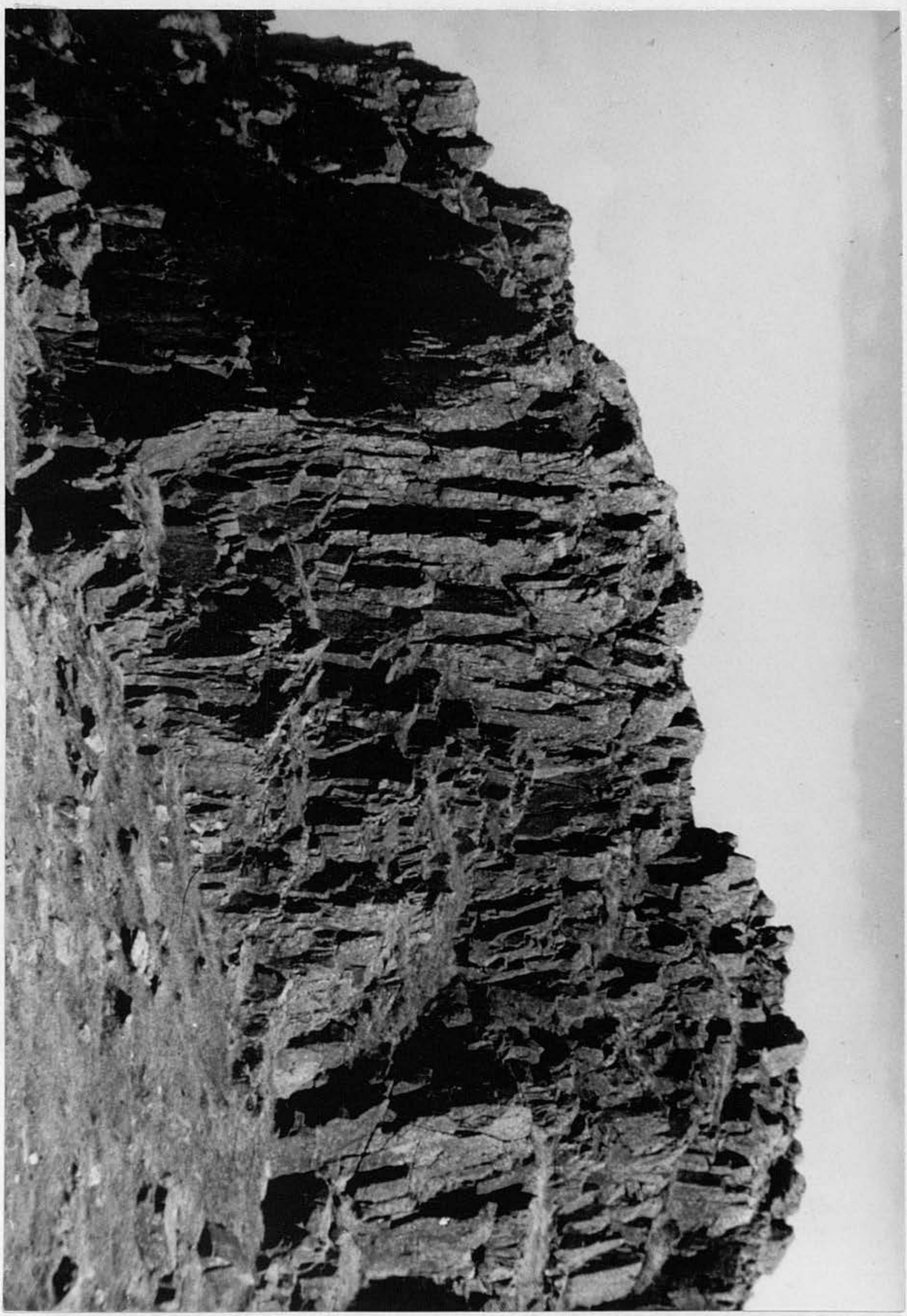
by the disposition of narrow veins of quartz and calcite, clearly in situ. Furthermore, the friable rotted material frequently passes laterally and vertically upwards (Photographs d and e) into bedrock that has remained comparatively fresh. This occurs particularly in areas where the granite/andesite contact throws different rock types into close juxtaposition; for example, in the valleys of the Broadstruthers, Common and New Burns. In these areas the granite has clearly rotted to a much greater degree than the andesite, but the latter is extensively broken along joints and shows some signs of chemical disintegration. At Old Middleton, the andesite is completely rotted (Photographs f and g). Since the writer did not have the scope nor techniques for a closer investigation of these intriguing sections, it is not proposed to consider whether chemical weathering or hydrothermal action was mainly responsible for the deeply rotted bedrock. The significance of the material to the glaciation of the Cheviot massif is considered in Chapter 7.

Of perhaps equal interest in a consideration of preglacial morphogenesis in the Cheviot massif is the occurrence of at least eleven well-defined tors (Photographs h and i, Maps 4, 5 and 7). Lying at various altitudes between 1,300 and 2,500 feet, five are composed of granite, the others of andesitic lava. The tors are mostly from 15 to 30 feet in height but are frequently adjacent to small craggy remnants of what appear to have been large tors formerly. The tors nearly all occur on broad, sloping spurs and they are normally elongated parallel with the axes of the spurs. It is also striking how consistently the tors are located at a pronounced convex break of slope, so that they all have asymmetrical cross-profiles, the "downslope" side of each tor being the higher. The bedrock exposed on each of the tors has clearly been severely weakened by weathering and probably pressure release, for wide fissures are very common (Photograph j). On the granite tors in particular, Photograph h Tors on the south-east side of the Harthope valley.

Photograph i Great Standrop tor.



Photograph j Jointing in the Housey Crag tor.



Photograph k   Great Standrop tor.

massive blocks several feet in size have weathered out in situ and large clitters of tumbled boulders frequently form aprons round the bases of the tors (Photograph k). The bedrock and adjacent boulders on each tor are chemically fresh. Evidence of the deep rotting observed in the bedrock exposures previously described is entirely lacking. In this connection, it is perhaps also relevant to mention at this stage that the blockfields and stone lobes which could be distinguished amongst the vegetation on The Cheviot and other hill summits, are composed of boulders derived chiefly from chemically fresh bedrock.

From the foregoing observations, it seems probable that in preglacial times an upper layer of deeply rotted bedrock, irregular in its vertical and horizontal extent, covered much of the Cheviot massif. Whether or not the bedrock had become rotted by deep chemical weathering or hydrothermal action accompanying the emplacement of the granite is unknown. It also remains uncertain whether or not tors composed of bedrock unaffected by rotting protruded from the hillsides and summits, and the extent to which the rotted layer was removed by mass movements before the Pleistocene. It is very probable, however, that the occurrence of periglacial conditions during the Pleistocene caused the rapid downslope movement of the relatively soft and permeable rotted layer, characterised by a high percentage of fine particles in its matrix. Presumably much of the rotted layer was also removed by glacier ice, for it appears to have contributed significantly to the red, gritty nature of the local till. That such events exposed the present tors and the hill summits is suggested by the fact that the observed bedrock and the blockfields consist of material that is chemically fresh. The tors were probably much more extensive formerly, and it is considered that a substantial layer of bedrock was removed from many of the hills during the Pleistocene. Depending on the irregularity and depth of the original rotted layer (small fragments of which now remain in the relatively sheltered localities described previously) the recognition of pre-Pleistocene erosion surfaces in the Cheviot massif may be more difficult than formerly acknowledged.

Previous Work

A substantial volume of literature concerning the landforms and deposits resulting from the former presence of glacier ice in Northumberland has already been written. The early workers in this field, for example, Tate (1849, 1866a), House (1862), Curry (1867) and Lebour (1878) were concerned chiefly with the drift and phenomena such as polished and scratched rocks (although W. Buckland (1839) had earlier referred to moraines at Middleton and Kirknewton). The pioneer work of these authorities enabled J. Geikie (1876), in conjunction with his own astute field observations, to present the broad pattern of glacial events in the Cheviot Hills. The detailed work accomplished by Miller (1887), Clough (1887, 1888, 1889, 1895) and Gunn (1895, 1897, 1900) for the Geological Survey added much constructive evidence for a more accurate assessment of the direction of former ice movements over the Cheviot area. Their lucid records of drift sections exposed during the construction of railway lines have proved invaluable for subsequent workers. Bulman (1891) and Lebour (1891) contributed further to the knowledge of the drift deposits of Northumberland. The first comprehensive accounts of channels cut by glacial meltwater in the Cheviot area were written by Kendall and Muff (1901, 1902). The tempo of investigations into the glacial phenomena of Northumberland apparently quickened in the early part of the present century, for a great number of papers dealing with "superficial deposits" and "glacial phenomena" were written by Dwerryhouse (1902), Woolacott (1905), Butler (1907), Smythe (1908, 1912), Merrick (1909) and Garwood (1910); a special "Boulders Committee" was even established by the University of Durham Philosophical Society, the reports of which were published in the Proceedings between 1905 and 1910. Following the discovery of Scandinavian erratics and so-called interglacial deposits on the coastal fringe of Durham, there appeared several papers discussing the significance of these materials. Trechmann (1915, 1920, 1931), Woolacott (1920, 1921). Raistrick (1931) provided an admirable synthesis of all the previous work, but a much greater body of evidence concerning the glacial phenomena soon became available following the Geological Survey's programme of revision. The work and ideas produced by Carruthers (1927, 1930, 1931, 1932, 1939) were of outstanding quality, and admirable reports were also written by Burnett, Dinham and Anderson in the memoirs. Little original research appears to have been accomplished on the topic of glaciation in the east Cheviot area during the succeeding decade; a generalised account by Eastwood (1953) simply summarised the previous literature. A thesis dealing with the general geomorphology of the east Cheviot area was written by Common in 1953 and included some observations on features produced by glaciation. The same writer later produced a paper in which various types of meltwater channels were discussed (1957). Derbyshire (1961) subsequently applied newer concepts to explain the landforms produced by fluvioglacial streams in the north-east Cheviots, and employed arguments that had previously been applied convincingly in Scotland and northern England by Sissons (1958<u>a</u>, 1958<u>b</u>, 1960, 1961<u>a</u>). Much more recently, the present writer (1966) has discussed some of the results emerging from the detailed field mapping of meltwater channels in the north-east Cheviots. A similar, but expanded paper will be published in 1968.

Although many references were made specifically to the east Cheviot area in the literature outlined above, the great profusion of meltwater channels, ridges and terraces of sand and gravel, kettle holes and deposits of till suggested that there was ample scope for a re-investigation of these landforms. Concepts evolved during the last two decades concerning the genesis and interrelationships of landforms resulting from deglaciation in Scandinavia and Britain indicated the inadequacy of previous work in the east Cheviot area. For this reason a programme of detailed field investigation was planned in 1961 for that part of north Northumberland; the field-work was concluded in 1966. The approximate limits of the territory mapped on the Six-inch scale are indicated on Map 12.

CHAPTER 1.

THE DIRECTIONS OF FORMER ICE MOVEMENT

The first general map illustrating the direction of former ice movement in the east Cheviot area was that presented by J. Geikie (1894), on which he showed the "British Isles During the Epoch of Maximum Glaciation". Based principally on the distribution of erratics in the till and on the direction of glacial striations, Geikie's map clearly illustrates only one general direction of ice movement over that part of Northumberland north of the river Tyne; the movement was from north-west to south-east. Observing that the drumlins in the Tweed valley "show that the ice-sheet of Teviotdale and Tweed gradually turned away to the east and south-east as it swept round the north-eastern spurs of the Cheviots", and that "The same curious deflection affected the great ice-stream that occupied the basin of the Forth", Geikie concluded that Scandinavian ice advancing into the North Sea Basin compelled the Scottish ice to flow south-eastwards into England. Geikie clearly formulated these ideas much earlier, for a similar account of ice movements in the Tweed valley was written by him in 1876; a map did not accompany that particular discussion, however.

Subsequent to Geikie's early writings, the detailed field investigations of Miller and Clough (1887) led to the first modification of his lines of former ice movement for the southern flanks of the Cheviot Hills. They suggested that the general glaciation of the open Fell country south of the Border watershed was "in a direction a little north of east". In support of this statement they included a table of all the observed sets of glacial striations and concluded that "This mass of moving ice can have been nothing less than a general ice-sheet." Evidence from the transport and distribution of erratics endorsed the conclusions derived from the glacial striations, because the erratics from both distant and local sources indicated a movement from the west. In a subsequent memoir, Clough (1888) described erratics in the south-east Cheviots that could have come only from the south-west; he also mapped "many glacial striae on the Harbottle Hills (Quarter-sheet 108S.E.), which point from this direction." Clough did not discuss the wider significance of these observations in relation to the glaciation of Northumberland, however, even although Geikie's generalised lines of ice movement were approximately at right-angles to those established by Clough's evidence.

The first detailed discussion of the direction of former ice movement was put forward by Smythe (1912). Following an analysis of evidence supplied by glacial striations, erratics in the till and meltwater channels, he was able to present a map illustrating the sequence of glacial events in Northumberland. The striations (Map 12) showed that two general directions of ice movement could be recognised, one from west to east, the other from north to south. Smythe acknowledged that the former direction varied locally, however, (presumably on the basis of Clough's earlier writings), becoming east-south-east in the valley of the North Tyne, and east-north-east in the country between the Rede and the Aln; but on the open Fell country between the Pennines and the Cheviot Hills, he noticed that the west to east series of striations "frequently bears no relation to the surface features, crossing deep valleys and the tops of lofty hills". The other principal movement was recognised chiefly in the coastal province of Northumberland, where the striations gradually turn towards the south-east and then south, more or less parallel with the coast. For example, Smythe stated, "The north-to-south series is well developed in the country between Wooler and the coast due east of that town, thence southwards along a strip of land about ten miles wide." A small number of striations in that area were observed to trend slightly to the west of south, suggesting some degree of movement landwards from the North Sea Basin. It was therefore clear from the striations that glacier ice had enveloped the Cheviot massif in the form of two major streams that appear to have merged together east of the massif. Further evidence concerning the relative ages of these ice movements, however, was also grasped by Smythe, for he stated, "Frequently the same rock surface shows both series of striations, and in one or two favoured instances, ... it is apparent that the southerly trend is later than the easterly one. At Little Mill the directions are to the south and west of south and the former is of earlier date than the latter." The distribution of erratics strongly supported the directions of former ice movements deduced from striations, and allowed Smythe to make the following conclusions. "At the beginning of glacial conditions, it is evident that the Border Hills between the North Tyne and Cheviot sent forth considerable streams of ice in all directions. The ice from the porphyrite area was hemmed in its western progression by the ice from the Carter and Peel Fells (Carter Ice) and driven down the left bank of the Redewater. The Carter Ice was similarly barred by the great western sheet of ice from the Solway district and driven along the left bank of the Tyne. The three great streams converging near Redesmouth were impelled in an easterly direction along the Wansbeck and then south-east towards Tynemouth ..... At maximum glaciation it is probable that the western ice sheet held complete sway almost as far as the coast. On the northern side of the Cheviots, ice flowing down the Tweed valley seems to have checked the flow of Cheviot ice in that direction and to have surmounted the outlying spurs of the hills.

There is some evidence that a sheet of ice flowed southwards along the coast at an early stage of glaciation. Towards the end pressure from the

North Sea had become the prime factor, its effect being recognisable to a distance of 14 miles from the coast. As a result of this the Tweed and local ice was thrust up the valley of the Till, but was barred near Hedgeley by the Cheviot ice."

A subsequent account of the glaciation of Northumberland and Durham by Raistrick (1931) is very largely based on Smythe's earlier work and there was no new evidence or difference of opinion put forward concerning the east Cheviot area. Of considerable significance to the glaciation of the Durham area and of Britain in general, however, was the discovery of glacial drift plugging fissures and valleys in the Magnesian Limestone of the Durham coast. The drift contains no stones from Scottish or English bedrock and consists chiefly of debris that was apparently dredged off the North Sea floor by Scandinavian ice. Plio-Pleistocene fauna and flora, chalk and flint erratics and rock-types peculiar to Norway testify to a movement of glacier ice from the north-east or east. This remarkable drift is overlain by till similar in nature to that occurring in the coastal zone of Northumberland and the former is undoubtedly the earliest glacial deposit in northern England.

The re-examination of north-east Northumberland provided further accounts of striations by Gunn (1927) and Carruthers (1930). The former observed that from the Doddington area, where striations point between east and east-south-east, they generally trend progressively towards the south farther east, particularly in the vicinity of Belford and the off-shore islands. This simple picture is complicated locally, however, as on Weetwood Moor and near Chillingham, where striations pointing south and 15° west of south were recorded. Carruthers (1930) was able to conclude that two directions of ice movement were apparent; firmly emphasising that "This does not necessarily imply two distinct periods of glaciation, separated by an interval of time", he suggested that more probably, there was only one glaciation, the direction of which was later "radically altered by outside influence". The earlier movement was predominantly in an easterly direction and is represented in the southern part of the area by striations trending north-east and 20° north of east on Bewick Moor, and in the northern part, by several striations aligned between east-south-east and south-south-east. Carruthers recognised the second direction of ice movement as being very different; "the trend is more or less to the south, and the abrupt diversion is no doubt due to the presence of Scandinavian ice in the North Sea". Unfortunately, it is not clear exactly upon what premise the latter statement is based, for Carruthers does not refer to cross-striae and his evidence appears to be founded on the fact that "the general south-east trend of more northern areas (Belford, etc.) veers slightly east of south", and "Farther south this changes to 23° west of south." Nevertheless, his general conclusion agrees with that reached some 20 years earlier by Smythe.

Much more recently, Eastwood (1953) has summarised the general sequence of glacial events in northern England and it seems to be generally agreed that "the onset of glacial conditions on the north-east coast was heralded by the arrival of Scandinavian ice". There is some controversy concerning subsequent ice movements, however, for Eastwood continued, "Some observers believe that subsequently northern ice, i.e. from the Cheviots, Tweed and east Scotland, twice occupied the coastal regions of south Northumberland and Durham, with an intervening stage during which western ice, from the Lake District and the south-west of Scotland, and local ice from the Pennines and Cheviots, reached the coast. Others, however, maintain that only one British glaciation affected this region, with changes in the direction of the ice." There is no recorded evidence in northern Northumberland, however, for two advances of northern ice, separated by an advance of western ice, and the basic sequence established by

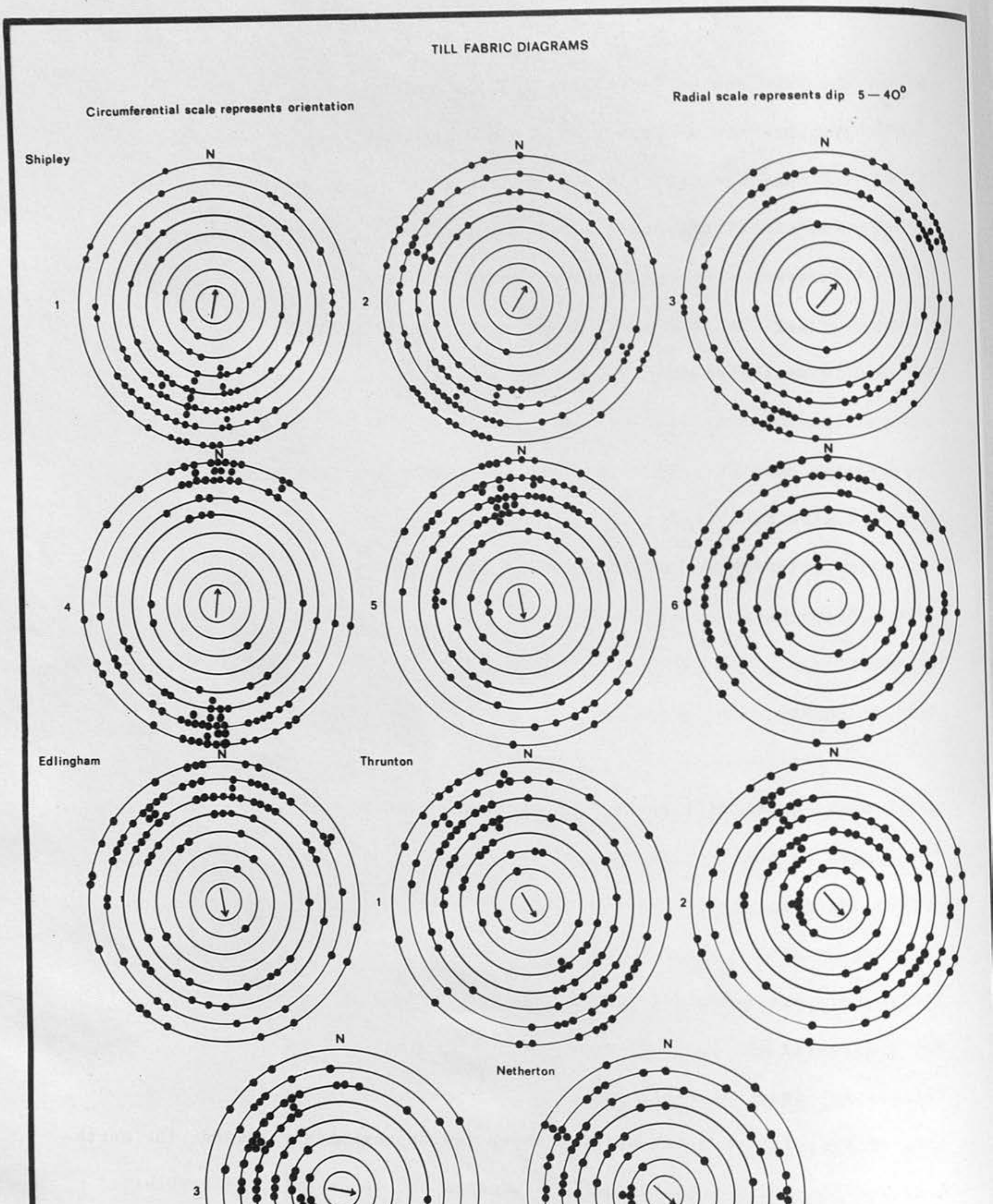

TILL FABRIC DIAGRAMS
Circumferential scale represents orientation
Radial scale represents dip 5 — 40°
Shipley
N
1
2
3
4
5
6
Edlingham
Thrunton
Netherton

Figure 1.1

Smythe and endorsed by Carruthers remains unmodified.

Since the programme of field work undertaken in connection with this thesis was concerned chiefly with fluvioglacial phenomena, a detailed search for glacial striations and analyses of the drift for erratic content were not made. However, since good exposures of till occur on the eastern fringe of the area under investigation, a series of fabric analyses was made in an attempt to establish the former direction of ice movement. The method of analysis has already been outlined (Introduction). The drift sections along the Shipley Burn (Map 9) were fresh and almost vertical in 1964 when the analyses were made at the six selected sites. The drift sequence is particularly interesting in this locality because two distinct tills are clearly separated by beds of laminated clay and fluvioglacial sand and gravel. The thickness of the exposed drift varies from 12 feet to over 60 feet, depending upon how deeply the stream has become incised. The stratigraphic details of each site are listed in the appendix. Since there has been some doubt on whether the upper till represents the ground moraine of a separate glaciation or simply the ablation moraine of the glacier which deposited the lower till, it was hoped that a comparison of the preferred orientation and dip of the constituent stones would perhaps provide some enlightenment on the controversy. The results are illustrated in Figure 1.1. At the three sites selected in the lower till (1, 3, 4), there was a strong preferred orientation and dip of the stones along axes that vary from 10° west of south to 30° west of south, suggesting a former movement of glacier ice approximately towards the north-north-east. Of the three analyses made in the upper till, one exhibited a completely haphazard arrangement of stones (site 6), another showed a poor preferred orientation indicating ice movement towards the north-north-east (site 2), and the third had a strong orientation and dip in a direction 10° west of north, suggesting ice movement towards the south-south-east. The validity of these results is difficult to assess, particularly in view of doubts recently cast by Young (1966) upon fabric analyses as a method of accurately reconstructing the former direction of ice movement. He stated, "There may be several different pebble alignments in a till of any thickness and any one orientation analysis may not be representative of the till." Concerning the lower till, there is evidence supporting the deduced north-north-easterly flow of ice in adjacent areas where striations trend between north-east and 20° north of east (Bewick Moor and Titlington Pike). It is therefore possible to interpret the lower till as the product of the general easterly movement of ice, recognised by Smythe and Raistrick, before the later and more southerly movement superceded. S

Adjacent to these sites and to the north and east, the majority of striations and fluvioglacial phenomena indicate a former direction of ice flow towards the south or slightly east of south, and it may certainly be suggested that the latter reflect the latest movement of ice in this area. With regard to the upper till, the three analyses tend to be confusing. The fact that the fabric arrangement of one is completely haphazard suggests ablation debris in which there is normally little or no preferred dip and orientation of the stones. It is possible that the poorly defined preference at one of the other sites also suggests ablation till, but the third site strongly indicated that glacier ice moved towards the south-south-east. One explanation is that the englacial detritus, which subsequently became ablation moraine, was in places arranged into a preferred alignment, reflecting the direction of sheer planes in the ice. As such, it indicates the latest direction of ice movement and is in keeping with evidence from all the adjacent fluvioglacial phenomena showing that melt-water drainage was directed towards the south and south-east. In this respect, the evidence agrees with the conclusions reached by Smythe and Carruthers, in that the upper till represents the ablation moraine of the same ice mass which deposited the lower till, and also suggests a later phase of ice movement in a south-south-easterly direction.

The site on the right bank of the Edlingham Burn was in an 8-foot section that showed 3 feet 5 inches of purply-brown till at the base, overlain by alluvial sediments. The preferred orientation and dip of the stones is strong in the broad sector between 50^{o} west of north and 30^{o} east of north, so that the average alignment of former ice movement on this basis is towards 5^{o} east of south. This is consistent with the trend of adjacent striations and meltwater channels.

Another three analyses were made in the broad embayment by the Coe Burn, a south bank tributary of the Aln. A considerable thickness of till lies in this embayment and is currently being quarried for a brick and tile factory. The till is variable in colour, with blue, red and brown patches creating a mottled effect. Pockets and lenses of sand, from a few inches to several feet in size, commonly occur, and some laminated clay was also observed in similar positions. The stone content of the till appears to be somewhat variable, being almost stoneless towards the centre of the valley, but containing blocks up to 6 feet by 4 feet in size at other parts. Two of the preferred orientations were very strong along axes 30^{o} west of north to 30^{o} east of south and 40^{o} west of north to 40^{o} east of south. The preferred dip was less pronounced, but indicated a direction of ice movement towards the south-east. The dip and orientation of stones at the third site showed a strong preference in the broad segment between 40^{o} south of west and 60^{o} north of west, the average of which suggests ice movement roughly towards 10^{o} south of east. The directions of ice flow suggested by the Thrunton sites at first appear problematical in relation to the alignment of meltwater channels on the adjacent south-east flanks of the Cheviot massif; but since the sites occur precisely where the glacier ice from the south-west came into conflict with that from the north, the fabric analyses probably reflect that part of the western ice mass which was deflected sharply towards the south-east by the more powerful ice from the north. A consideration of the lithology of the constituent stones in the till tends to endorse this interpretation. Had the ice mass which deposited the Thrunton till come directly from the north-north-west, as suggested by the stone orientations, then the till would undoubtedly have contained a much greater percentage of igneous material than that shown by the random site samples (7%, 11%, 3%). Furthermore, pieces of the characteristic red mica-porphyrite from the Biddlestone laccolith are relatively common in the till, indicating that it was deposited by ice from the south-west.

The remaining analysis of the till was made in a section exposed on the right bank of the Netherton Burn (a tributary of the river Coquet) approximately one mile south-east of Netherton village. The preferred dip and orientation of the stones was slightly amorphous, but an alignment towards the south-east was evident, suggesting that the western ice mass, even in the vicinity of Netherton, was being pulled round towards the south-east under the influence of ice from the north. That the ice which deposited till in the Netherton area had come directly off the flanks of the Cheviot massif is indicated by the remarkably high percentage of igneous fragments it contains (82%).

On the basis of glacial striations and erratics observed by earlier workers and recent fabric analyses of the till, it may be concluded that the Cheviot massif was partially enveloped by two major streams of glacier ice from the west. The ice flowed round the north-east and south-east flanks of the massif and converged in the vicinity of North Charlton. Subsequently, the northern ice mass was caused to flow in a southerly direction over the territory east of Thrunton, causing the distal extremity of the western ice mass to swing round in a similar direction. Reasons for the change in movement of the northern (Tweed) ice mass remain conjectural, since evidence upon which to base a valid explanation was not observed in the thesis area. It is possible that the approach of Scandinavian ice over the North Sea Basin caused a southerly deflection of the Tweed glacier, as envisaged by Geikie (1876). However, since there is considerable doubt as to whether or not the Scandinavian ice sheet expanded sufficiently during the last glacial maximum to influence the flow of British glaciers, this interpretation is perhaps the least probable. An alternative hypothesis concerns the joint Highland/Southern Uplands ice sheet flowing out into the North Sea Basin north of the Lammermuir Hills. Since the dimensions of this combined ice sheet were sufficient to enable it to spread laterally over the eastern extremity of the Lammermuir Hills (indicated by striations, erratic transport and meltwater channels) then it is reasonable to expect that it was the progressive extension of that ice sheet which eventually forced the Tweed glacier to flow in a southerly direction down the east coast of Northumberland. The latter event may have occurred relatively late in the glaciation, following a period of more easterly-directed flow over north-east Northumberland as envisaged by the earlier workers. It appears that the central part of the Cheviot massif was contemporaneously affected by a local ice cap, but further discussion of this is deferred until a later chapter. It should also be mentioned that if the last period of maximum glaciation was characterised by minor retreats and readvances of ice, then the various directions of flow suggested by the above data possibly result from such events. However, since there is an absence of other evidence in the east Cheviot area indicating oscillations of the last ice sheet, further discussion concerning that possibility cannot be undertaken.

CHAPTER 2.

MELTWATER CHANNELS IN THE NORTH-EAST CHEVIOTS

Introduction

The north-east fringe of the Cheviot Massif below 1,100 feet is extensively furrowed by glacial meltwater channels whose size and complexity attracted a paper by Kendall and Muff as early as 1902.

The features vary considerably in form and dimension. Many are small depressions only a few feet wide and deep throughout their lengths; some begin in this way and then develop larger proportions. Several channels are impressive features over 70 feet deep with narrow, V-shaped cross profiles, while those marked 7, 11, 13, 14, 24, 29 and 30, Maps 5, 6 and 8, reach over 100 feet in maximum depth. Between these extremes many display various characteristics of form and size. The channels are cut either entirely in bedrock or in drift, or partly in drift and then into underlying bedrock. Channel floors are covered with various deposits which include peat, alluvial fans and other fluviatile deposits, scree and solifluction formations. Derbyshire (1961) described "considerable amounts of till in col gullies", the till having a "superglacial and englacial" origin, but nowhere was this confirmed. The floors may be completely dry, or occupied by marshy vegetation or small streams clearly unrelated to the size, form and other characteristics of the valleys through which they flow. Seldom do the meltwater channels exist as isolated features on the hillsides; they are generally grouped in systems with tributaries, distributaries and abandoned sections. Furthermore, the most impressive systems occupy pre-existing cols and valley-heads.

Previous Work

The presence of glacial meltwater channels in the Cheviot Massif was first acknowledged by Clough in 1888 who wrote, "At various places we meet with dry steep-sided little valleys or deans crossing over watersheds ..... it has been supposed that they may have been formed by streams from glaciers". The first detailed discussion on the origin and significance of these features was presented by Kendall and Muff thirteen years later. Appearing the year previous to the publication of Kendall's classic monograph on the Cleveland Hills this paper described selected channels, mainly those which are the most impressive. In agreement with current principles of channel formation the authors envisaged a system of lakelets ponded up in small valley-heads which radiate from the fringe of the Massif. The margin of a large glacier from the Tweed valley sweeping round the north-eastern flanks of the Cheviot Hills was postulated as the barrier to normal drainage from ice-free hillslopes above. The presumed impenetrable ice controlled a sequence of lake levels whose impounded waters breached intervening divides as they overspilled from one to another eastwards. Upon ice recession deep, dry, "overflow" channels, cut in cols through watersheds, marked the former courses of water draining the ice-dammed lakes. From this evidence the writers concluded that "while 'foreign' ice was rising along the flanks of the Cheviots to an altitude of 1,000 feet, not only were the spurs free from any native ice-sheet, such as Cheviot or Hedgehope might have been expected to support, but even the lower ends of the intervening valleys were occupied, not by great native glaciers, but by lakes."

Smythe's detailed account of the glacial geology of Northumberland in 1912 contained a chapter on glacial meltwater channels which he termed "The Forsaken Water Courses". He observed that "They occur in two positions; firstly, and most frequently, cutting the watersheds between the pre-glacial valleys or the spurs and subsidiary water partings connected therewith; secondly, as trenches running along a hill-side, roughly parallel to the water parting. Occasionally ..... the two types are combined, the upper part of the water course cutting along one side of the ridge, then swerving sharply across the divide and dropping abruptly in cascade-fashion down the other side." Although Smythe observed many of the intricate details such as "branched courses which isolate steep-sided rocky hillocks ....." perhaps suggesting an explanation for the channels more complex than Kendall and Muff had assumed, he entirely accepted the latters' hypothesis. Adding only the category of "marginal trench" to that of "overflow channel", Smythe described the retreat of various contemporaneous ice masses in great detail, indicating on a map various positions of the retreating glacier margins; these were established on the basis of marginal and lake-overflow channels. Since this paper followed only eleven years after that by Kendall and Muff, Smythe just briefly referred to channels north of the Breamish: "..... most of these have been described by Kendall and Muff, and reference for details may be made to their paper."

Detailed work carried out in the north-east Cheviots for the Geological Survey led Burnett (1932) to believe that Kendall and Muff had given "A completely satisfactory explanation of Clough's 'dry denes' .....", for "A re-examination of the ground confirms Kendall and Muff's work in almost every particular." His only criticism was 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." The validity of Kendall's lake-overflow hypothesis for meltwater channels which had so readily been accepted in Britain, where it endured as the dominant theory until severely challenged in 1958, was clearly questioned by Burnett, but he did not fully elaborate his criticism. Impressive members of the channel system were mapped and described systematically as Burnett depicted the gradual emergence of the Cheviot foothills from the receding Tweed glacier in considerable detail.

Much more recently, Common (1957) has reconsidered the significance of Cheviot meltwater channels paying particular attention to their remarkable variety in form; but their significance to the pattern of deglaciation in the east Cheviot area was not discussed. Acknowledging the presence of channel types recognised by Kendall in the Cleveland Hills, Common added three new categories based on morphology and genesis. Although considerably influenced by Kendall's theories, Common questioned the widespread existence of ice-dammed lakes in the Cheviots suggested by the former. The recognition of up/down longitudinal profiles in many channel floors, the possibility of meltwater superimposition from ice to the ground surface and an awareness of the possible importance of subglacial drainage, led Common to conclude "The last word on the subject of meltwater channels has by no means been written ..... there is still a great deal to be learnt about them and from them." Consequently, a recent paper by Derbyshire (1961), in the light of current trends in glacial geomorphology, further investigated the question of genetic interpretation of Cheviot meltwater channels. Applying arguments similar to those developed by Sissons (1958) for the Eddleston valley some years previously, Derbyshire severely criticised Kendall's interpretation of the channels. Whereas Common described meltwater features mainly in relation to their variety in morphology, Derbyshire grouped them genetically according to their form and location. Three main groups were recognised; "The first group consists of channels which have breached pre-existing cols in the upland. ..... The second is made up of channels running down hill-side slopes, to join accordantly the present valley flood-plains. The third group of channels lies sub-parallel to the contours ....." Indicating how previous explanations are irreconcilable with the complex channel forms and citing up/down floor profiles and the presence of till in the majority of "col gullies" as positive evidence, Derbyshire concluded that channels in the first group were formed subglacially under hydrostatic pressure; he called them "subglacial col gullies". Steep long itudinal profiles and again the presence of till in the channels were considered sufficient evidence to justify the second group being termed "Subglacial channel systems". Admitting that "The rarity of true marginal drainage is very striking," Derbyshire declared "poorly defined benches" near Wooler to be marginal in origin, while "occasional channels of submarginal origin" were also recognised in the area.

Channel Types

According to Sissons (1963) streams can occupy six possible positions in relation to glacier ice. They can be supraglacial, englacial, subglacial, open ice-walled, marginal and proglacial. He pointed out that supraglacial and englacial streams obviously cannot cut channels on the ground since by definition they flow entirely in ice-walled and ice-floored channels.

Proglacial: In the north-east Cheviots a proglacial explanation can be immediately excluded also. The system of channels as a whole lies almost at right-angles to stream valleys opening out from the massif (Maps 5, 6 and 8). The channel system is thus explicable only by glacial diversion just as Sissons (1963) found with the Carlops system on the lower slopes of the Pentland Hills. Furthermore, individual channels and parts of channels within the system can only be explained if glacier ice existed in their immediate vicinity during their formation.

Marginal: Meltwater channels that trend parallel with or at small angle to the contours of a slope have frequently been explained by subaerial drainage along the ice margin. Marginal channels in the north-east Cheviots were described by Smythe (1912), Burnett (1932), Anderson (1932) and Common (1957). On the other hand, Derbyshire (1963) commented on the rarity of true marginal drainage and believed the most common form of marginal drainage feature to be the marginal bench. He tentatively explained channels at 10, Map 5, in this way, suggesting that the absence of true marginal channels is expected since they are easily destroyed by rock and soil creep. A detailed discussion of marginal meltwater drainage and channels formed by such streams has been presented by Sissons (1961) who at the very outset emphasised that while a channel that runs along or exists at small angle with the contours of a slope ".....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." The implication is that such a channel might well have been formed by a stream flowing along the hillside in a tunnel beneath the ice.

The only meltwater channels in the north-east Cheviots that are aligned parallel with or at small angle to the hillslopes are found in two places; (a) on the north-east slopes of Humbleton Hill (10, Map 5);

(b) on the hillslopes rising south-west from South Middleton (19, 20, Map 5).

Those on Humbleton Hill have been interpreted as marginal in origin by Burnett, Common and Derbyshire; those by South Middleton are considered marginal by Common but Derbyshire included them within the category described as "subglacial systems". In both areas the channels can be demonstrated to be subglacial in origin.

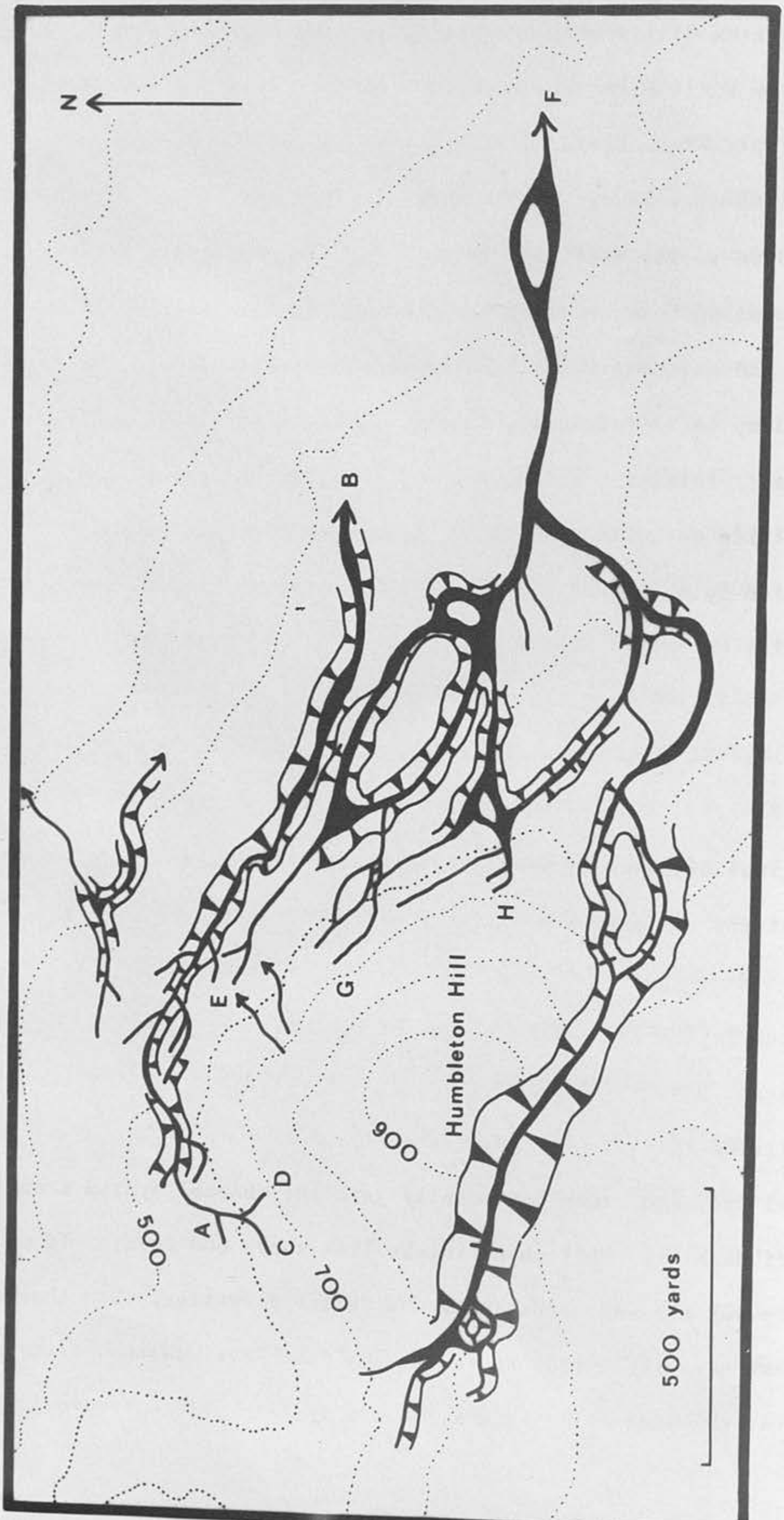
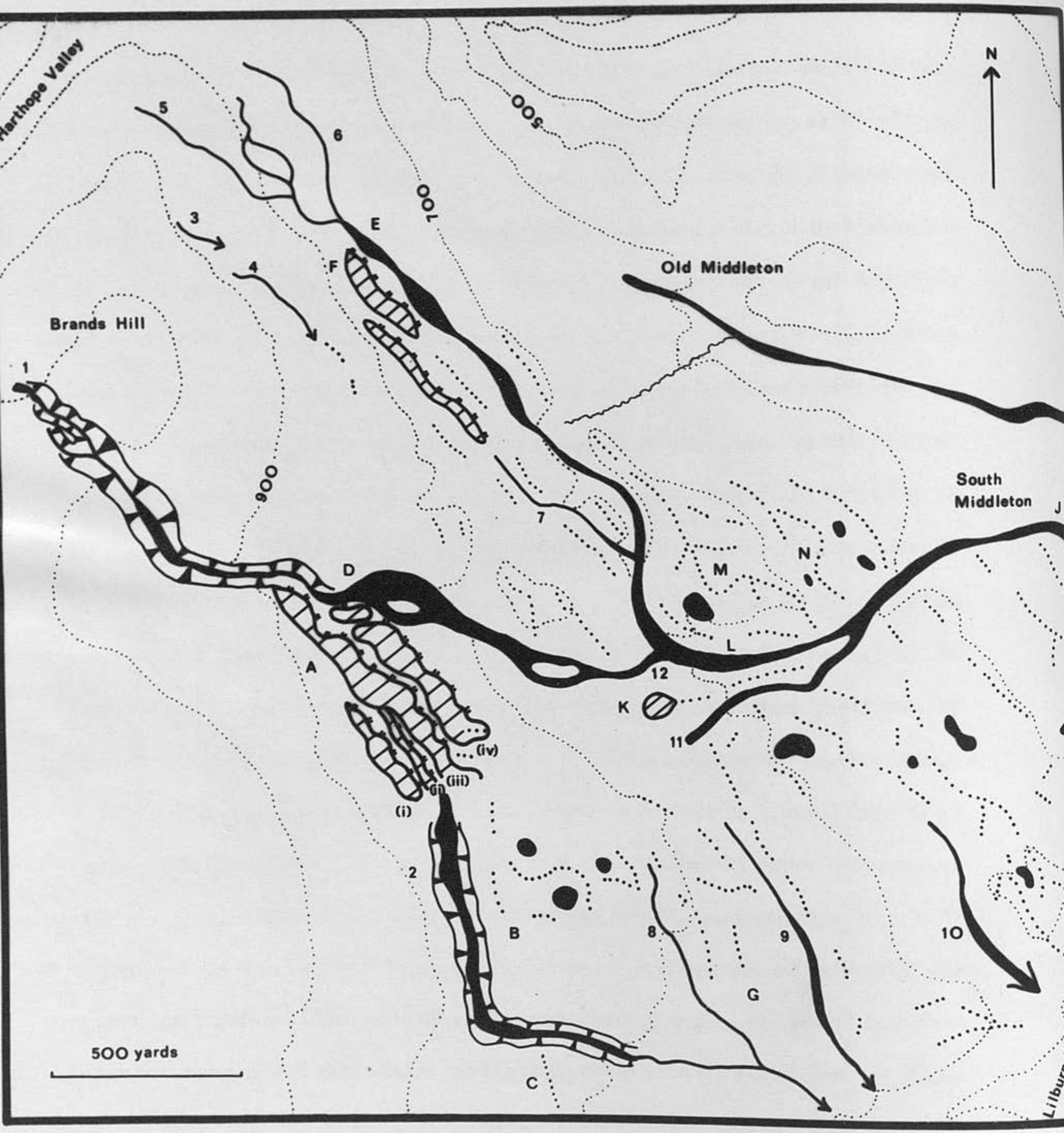


Figure 2.1

(a) The Humbleton Hill system (Figure 2.1): The highest feeders of this system are small rock-cut notches a few feet deep and wide lying on the steep slopes of Humbleton Hill between 550 and 700 feet. Below 500 feet the hillside levels out into a definite shoulder before sloping quite gently down to the Milfield Plain. The channel system attains its fullest development on this shoulder where a complex anastomosing network of benches and channels with tributaries and distributaries, mostly cut in rock and up to 30 feet deep, furrow the hillside almost parallel with the contours. Intakes C and D, are dry gullies that plunge obliquely down the hillside at a steep angle. Since their courses have plainly been controlled by the immediate proximity of glacier ice, they are interpreted as subglacial chutes similar to those described by Mannerfelt (1945), Sissons and others. They join channel AB accordantly, and since the latter lies at a lower level, it too must have been formed subglacially. If the complicated system of channels between points E and F was formed by marginal drainage it would be necessary to invoke highly improbable oscillations and shapes of the ice margin. The small rock-cut features at G and H run parallel with the contours, resembling marginal channels described in the literature. The highest of these lies at 700 feet. They cannot have formed marginally to the ice since after plunging down the hillside in chute-like sections, themselves quite out of keeping with marginal drainage, they accordantly join the channel system from the col south of Humbleton Hill whose intake lies above 800 feet. If meltwater from ice was cutting channels at the latter elevation, then the ice surface must have stood well above the 700 foot level on the north-east slopes of the same hill; since the two systems join accordantly, that lying about 700 feet at least must have formed subglacially. Finally,



Harthope Valley
N
500
700
900
Old Middleton
Brands Hill
South Middleton
1
2
3
4
5
6
7
8
9
10
11
12
A
B
C
D
E
F
G
H
J
K
L
M
N
(i)
(ii)
(iii)
(iv)
500 yards
Lilburn Burn

Figure 2.2

it will be shown later that the system at 800 feet was subglacial in origin and so the lower formations are also subglacial meltwater channels.

(b) The South Middleton system (Figure 2.2): Marginal meltwater channels are generally most clearly developed on slopes of gentle gradient, as on steep hillsides the marginal streams tend to erode laterally downslope into the ice rather than into the more resistant bedrock surface. The slopes above Old Middleton and South Middleton are less steep than those normally developed on the igneous massif, whose faultline edge is frequently precipitous, and an area of gently undulating relief lies between 700 and 850 feet. The hillsides are extensively covered with fluvioglacial formations, generally interpreted as marginal meltwater channels in previous literature, apart from channel 1 which Derbyshire included in his col gully category. However, the four bench-like features at A are unlikely to be marginal meltwater channels, since the presence of ice-contact slopes and a few exposures revealing sand and water-worn gravel and cobbles, indicate that they are fluvioglacial terraces constructed by deposition. They all begin, at successively lower altitudes, in direct alignment with channel 1 almost precisely where the latter feature loses much of its conspicuous form, and so they may be interpreted as depositional phases of the same meltwater stream. Channel 1 begins at approximately 975 feet on the upper slope of Harthope valley as a shallow, dry depression and rises 10-15 feet into the col at Brands Hill. Here it deepens abruptly, as rock-cut plunge-pool sections continue its course through the col, and steeply descends the hillside beyond directly across the contours of the slope. Mid-way through the col an abandoned channel loop 12-15 feet deep hangs over 20 feet above the main gorge on its right wall. These characteristics displayed by channel 1 and its location in the floor of a deep col clearly favour the subglacial interpretation implied on Derbyshire's map (he did not refer to this channel in the text). In particular, the uphill gradient at its intake infers a subglacial river flowing under hydrostatic pressure.

Channel 2 is aligned parallel with the contours of the slope for about 620 yards before it turns at right-angles and runs directly downhill for a further 450 yards. It reaches 25 feet in maximum depth and bedrock is exposed on both of its steep walls. Figure 2.2 illustrates how terraces (i) and (ii) in particular are plainly connected with channel 2 which was therefore undoubtedly cut by the same stream of meltwater. A sequence of erosion, deposition and erosion is thereby revealed by channel 1, the terraces and channel 2. Since the two former groups of phenomena have been interpreted as subglacial in origin, it follows that channel 2 must be accounted for in the same way, and although it slopes at small angle to the hillside it was probably cut by a meltwater stream flowing in a subglacial tunnel.

Features 3, 4 and 5 are rock-cut channels up to 10 feet deep and they descend the hillside too steeply to be considered as having formed along the ice margin. They are much more satisfactorily explained as submarginal channels representing courses assumed by small subglacial streams that were controlled by ice tunnels. Feature 6 is aligned along the hillside at a much smaller angle to the contours of the slope and is a conspicuous, two-sided channel, becoming 30 feet deep where rock is exposed on its left wall. Beyond point E, however, it continues as a broad, shallow feature with a flat, marshy floor and occupies a position between ridges of fluvioglacial sand and gravel. Channel 6 cannot have formed subaerially at the ice margin because it is joined accordantly by the subglacial channel 5 which descends from a higher level, and so channel 6 must have been cut contemporaneously by a subglacial river flowing at a lower level. Furthermore, together with channel 7, it links up with channel 1 in an accordant confluence suggesting that the whole system of channels above Old and South Middleton was cut by a network of subglacial streams flowing simultaneously.

Elsewhere in the north-east Cheviots there are no channels or bench-like features aligned parallel with or at small angle to the contours of any hill-slope. It is therefore concluded that there are no meltwater channels of undoubted marginal origin in the north-east Cheviots.

Open Ice-Walled: While it is possible that some small channels and perhaps parts of other larger features may have been cut by meltwater streams flowing in open ice-walled courses, such an interpretation is improbable for the majority of meltwater channels in the north-east Cheviots and should not be considered significant in any general explanation of deglaciation in the area. The following illustrations may emphasise the point.

(a) None of the meltwater channels with up/down longitudinal floor profiles can have formed in subaerial courses since considerable hydrostatic pressure is considered essential to explain such a profile satisfactorily. The meltwater streams must therefore have occupied courses beneath the ice.

(b) Several channels occupy the floors of narrow, steep-sided cols. Ice lying in the floor of a col of this nature is unlikely to have developed deep, long-lived crevasses or ice-walled channels penetrating to the col floor since ice from either side would tend to move down the steep col slopes towards the lowest point, thereby closing up or collapsing any open cavity that might have developed.

(c) The majority of meltwater channels in the north-east Cheviots descend over 100 feet from intake to outlet; several channel outlets lie up to 300 feet below intake level. Crevasses over 100 feet deep are not commonly reported from present glaciers and open walls of ice rising 300 feet above continuous stream courses are unknown. The lower courses of channels with such relief amplitude are therefore more likely to have been cut subglacially although it is conceivable that the intake sections were formed subaerially.

(d) Maps 5 and 6 illustrate the complicated nature of some meltwater drainage systems in which individual branches are frequently separated by narrow, sometimes sharp-crested, inter-channel ridges. For such intricate networks to have formed in open ice-walled courses it would be necessary to postulate equally narrow and elongated wedges and pinnacles of ice perched precariously on the channel divides. This explanation is hardly satisfactory to account for the establishment of the large composite systems in which branches up to 80 feet deep may be separated by narrow ridges only a few yards wide.

Lake Overflows: Kendall and Muff (1902) concluded, "The existence of the series of overflow channels points clearly to the former presence of a chain of small lakes held in the radial system of valleys of the Cheviots by a barrier of ice." This classic concept of Kendall's, which he fully developed for similar channels in the Cleveland Hills, was widely accepted for many years and generally acted as a basis for the interpretation of most channels subsequently described by other writers. It is only within the last decade that serious criticism of Kendall's concepts has arisen. This has been primarily by Sissons for the general application of these concepts to explain the majority of meltwater channels and to establish ice limits, and by numerous other

Photograph 2.a The Akeld channels; Gleads Cleugh on the left. (5, Map 5).

writers for specific areas, including the Cheviot Hills. It will be shown later in this thesis that the largest meltwater channel in the east Cheviot area was last in use as a spill-way for water held up by an ice barrier and evidence will be presented for the former temporary existence of three ice-dammed lakes, but no meltwater channels were eroded solely by overflowing lake waters.

Glaciers in many parts of the world today are known to pond up drainage from time to time, both marginally and proglacially on reverse slopes. Consequently, the channels described by Kendall and Muff should be reviewed bearing in mind the possible validity of their interpretation of them. However, the writers appear not to have mapped or observed all the meltwater drainage features in detail. Only the deep, impressive channels are mentioned and even these were wrongly observed. For example, the authors believed "that the lower end of one dry valley corresponds in altitude with the upper end of another one situated on the spur succeeding in the direction in which the valleys slope." Map 5 illustrates the complete error made by Kendall and Muff in this observation and even in the field the writer fails to appreciate how these authors did not see the outlet of one channel lying far below the intake of the next in sequence on the opposite side of each particular valley (Photograph 2.a). On this error alone their thesis falls, for only the presence of dry channels at a similar elevation formed the basis of their proposal that lakelets were ponded up in the valley heads aligned in sequence the one spilling over to the next and cutting channels through cols in the spurs. An additional objection to their theory is that even if lakes had been impounded by an ice barrier the valley heads which would have contained them are so small in area that any overflowing stream could hardly have eroded the majestic rock canyons fully 100 feet deep in two instances and over 40 feet deep in the majority of others. Furthermore, multiple intakes, abandoned segments, channels occupying positions other than the lowest point available over a spur and the absence of features normally associated with the former presence of lakes, such as shorelines, floor deposits and deltas, all constitute arguments precluding the former widespread existence of glacier-dammed lakes in the Cheviots.

Direct Cuts: In previous literature describing Cheviot meltwater channels occasional reference is made to "direct cuts". This type of channel is invariably located in cols breaching narrow stream divides and the descriptive term used implies that meltwater streams issued from an ice margin abuting against one end of the col, flowed "direct" through the col which was ice-free and entered another ice margin beyond at a slightly lower elevation. This concept would appear to have some possible validity in the north-east Cheviots.

As an extended lowland glacier downwastes over irregular relief, the highest hilltops are normally first to become ice-free and as ablation proceeds, more and more of the higher ground appears above the ice surface. In an area such as the north-east Cheviots where closely spaced valleys, separated by relatively narrow divides, are orientated approximately at right-angles to the direction of ice movement, then the following events are to be expected.

Intervening watersheds become ice-free while the valleys are still occupied by glacier ice. The level of ice in valleys proximal to the source area would normally be slightly higher than that in valleys next in sequence downstream (relative to ice movement). Since the direction of meltwater drainage is closely related to the slope of the ice surface, there would be a distinct tendency for meltwater streams to flow from one valley to the next in sequence in a downstream direction. Where conspicuous cols are located in the crests of intervening spurs it is possible that at certain levels during downwastage part of the ice margin would be situated at the entrance to a col and any meltwater stream issuing from the ice edge at that point would flow subaerially through the col. A single channel should be cut in the col floor, although a multiple intake may result if more than one stream issues from the ice edge. Once through the col the meltwater stream may flow supraglacially, englacially or subglacially as it meets the ice mass occupying the valley beyond; and since glacier margins usually exhibit convex slopes down to adjacent valley sides, englacial and subglacial courses are more likely to be assumed than one which is supraglacial. Streams that become subglacial may be expected to erode chute-like sections for some way down the valley side. With progressive downwastage, meltwater routes through cols higher up the interfluves would become abandoned as lower cols emerged above the ice surface.

In the north-east Cheviots a system of short valleys and narrow divides is orientated at right-angles to the direction of former ice movement in this area and prominent cols occur at various elevations on the spur crests (Map 5). Conspicuous meltwater channels are located in almost every col, and the validity of the above hypothesis must be considered. Close inspection of the meltwater channels, however, reveals that the majority cannot be explained as "direct cuts" for the following reasons:

(i) Several channels exhibit up/down longitudinal floor profiles, a characteristic generally accepted as evidence of subglacial drainage of meltwater.

(ii) In others, channel sections that would have been cut subaerially show peculiar anomalies such as abandoned and interlinked loops and isolated knolls in the channel floor; such phenomena are difficult to interpret by subaerial drainage and require the close proximity of ice to be adequately explained.

(iii) In some examples, tributary channels enter far beyond the main intake and since they can only have come from the ice, then much of the entire channel must have been covered by ice, at least as far as the tributaries' intakes.

(iv) Some tributary channels slope directly or obliquely downhill and are clearly subglacial chutes. In view of their accordant confluences with the main channels it is impossible for the latter to have been cut subaerially while adjacent slopes above were beneath glacier ice.

For these reasons the majority of meltwater channels cut through cols in the north-east Cheviots cannot be interpreted as "direct cuts".

Common: Common mapped meltwater drainage features in the Cheviots in much more detail than previous workers and observed the difficulty of explaining channels with up/down longitudinal floor profiles and those with accordant tributaries by the conventional hypotheses. He suggested (1957) that subaerial and subglacial streams flowing simultaneously in opposite directions along the same channel accounted for up/down gradients in channel floors; oscillating ice margins and multiple glaciation were theories put forward for reticulate tributary systems. Such complex and unusual explanations are considered unlikely since all the channels indicate drainage predominantly in the same direction and glacier margins have not been reported to oscillate in such a peculiar manner. Although unable to offer reasonable alternatives to interpret Cheviot meltwater channels in ways other than those propounded by Kendall and Muff, Common's work was extremely important in that it was a major challenge to the conventional views of the time in Britain. The suggestion of subglacial drainage and superimposition appeared in his paper, concepts that were greatly elaborated in subsequent years by other writers for the widespread interpretation of meltwater channels in Britain.

In the light of the foregoing discussion it follows that meltwater channels in the north-east Cheviots can be interpreted only by the subglacial drainage of meltwater; the process of elimination leaves this explanation and it now remains to consider this concept in more detail and offer more direct evidence in support of its validity.

Subglacial: The majority of meltwater channels in the north-east Cheviots were mapped by Derbyshire (1961) and interpreted as subglacial in origin. Evidence quoted in support of this hypothesis consists of "frequent reverse gradients" and "considerable amounts of till in col gullies". On his accompanying map twelve of the col gully category displaying uphill sections in their floor profiles are indicated. (Channel 72 on Derbyshire's map is not considered here because it drains in the opposite direction and does not belong to the system under consideration, but it is discussed in Chapter 7.) Inspection of these channels in the field reveals that three of them do not in fact possess true uphill sections, while those of a further two are doubtful. There remain only seven channels whose profiles require hydrostatic pressure during formation according to Derbyshire's interpretation. It seems rather a doubtful principle to apply this interpretation by analogy to the other nineteen col gullies lacking reverse gradients indicated on his accompanying map.

In conjunction with the presence of reverse gradients as proof of subglacial drainage in the Cheviots, Derbyshire considered that "Additional evidence includes the presence of glacial till in the majority of the col gullies". Detailed examination of these channels in the field reveals that their original cross-profiles have been extensively modified subsequent to their abandonment by the meltwaters that formed them. Frequently the upper slopes of channel walls are extremely precipitous, occasional exposures of

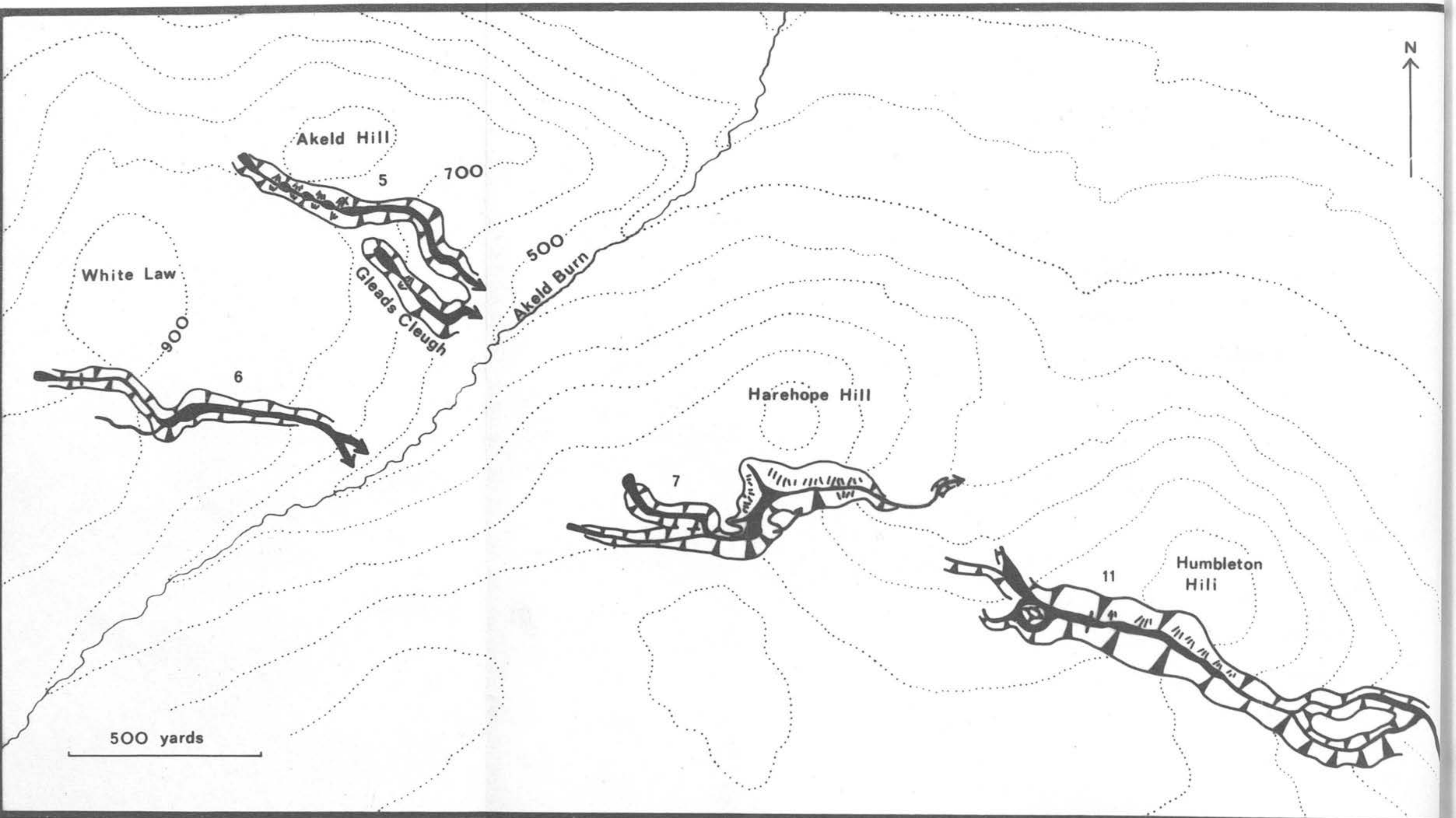

N
Akeld Hill
5
700
500
White Law
Akeld Burn
Gleads Cleugh
900
6
Harehope Hill
7
11
Humbleton
Hili
500 yards

Figure 2.3

bedrock forming near-vertical faces. Lower slopes, however, are normally much more gentle and exposures revealing bedrock are few or completely absent. The lower slopes are chiefly talus slopes, mainly in the form of fossil screes. Vegetation clothes much of these formations but in places the scree fragments have been too large for soil development and the establishment of the vegetation cover, so that they remain as bare rock screes. In the majority of "col gullies" the original channel floor is completely obscured by these accumulations which frequently create irregular profiles with relief amplitudes of up to 10 feet. Scree lobes may extend from both walls of the channels or else spread out mainly from southerly-facing sides, where freeze-thaw cycles were presumably more frequent, allowing more extensive scree development. Channels 5, 6, 7 and 11, Map 5, illustrate this point particularly well. No evidence of till deposits is apparent. Scree formation, solifluction, soil creep and other slope processes have led to such extensive accumulation in the majority of channel floors subsequent to their formation that any till deposits therein must be completely obscured. The possible presence of erratics is not considered reliable evidence for ablation moraine since they could have been derived from a variety of sources. These include soliflucted bedrock fragments of different lithology from slopes above the channels, soliflucted till, fragments carried in by the meltwaters that eroded the channels and anthropic sources such as dry-stone dykes or ancient hill forts and earth-works frequently perched above channel walls. Finally, it is undoubtedly relevant that the one-inch drift map of the Geological Survey shows no till deposits on or near the floors of any of the channels under consideration.

A detailed description of channels 5, 6, 7 and 11 (Map 5 and Figure 2.3), referred to above will perhaps illustrate the extent to which these major col gullies have been modified since their formation and emphasise the

Photograph 2.b Scree causing an irregular floor profile in the Akeld col channel.

improbability of any till deposits being exposed in their floors.

Channel 5: The col between Akeld Hill and White Law is trenched by a meltwater channel cut entirely in andesite bedrock. Beginning with a broad, trumpet-like intake 15-20 feet deep, it cuts through the interfluve with a straight section reaching about 40-45 feet in maximum depth. Considerable scree formations, many of which remain bare of vegetation and include blocks up to 4 feet across, are prevalent along this section (Photograph 2.b). Present scree activity is confined to quite minor developments where small particles only a few inches in maximum size tail down from very restricted rock outcrops on the upper slopes of both channel walls. The bright pink colour of these particles shows that they are relatively fresh in comparison with the grey, lichen and moss-covered fragments elsewhere in the channel. Scree development has been so extensive that lobes of debris have encroached from either wall to meet in the centre of the channel floor. This has created an irregular floor profile and three small basins up to 8 feet deep have been enclosed between confluent scree lobes. Stagnant water and marshy vegetation periodically occupy these basins.

A close vegetation cover, mainly of turf and bracken with some heather, clothes obvious scree slopes in many parts of the channel; a few sheep-scars reveal loose, angular scree fragments and the development of soil from the comminution of them below the turf. For example, at point x on the talus slope of the south-facing wall, 12-15 feet above the channel floor a hole was dug which penetrated through 24 inches of brown sandy soil in which a few flat angular fragments of andesite were present. At a depth of 24 inches the debris began to include particles coarser than sand and these probably continue to a much greater depth.

Photograph 2.c Gleads Cleugh. Intakes to the Harehope channel in the background.

After 500 yards the channel curves southwards through about 70 degrees and plunges steeply into the valley of the Akeld Burn where it attains its maximum depth of 70-80 feet. The steep channel sides along this section are very smooth and closely covered with vegetation. Minor exposures reveal at least 12 inches of material consisting of angular stones $\frac{1}{2}$-4 inches in size contained in a light-brown sandy soil matrix, all of which seems to have been derived from andesite. A block of this rock-type 4 feet by 4 feet lies on the channel floor immediately before the profile plunges over the valley side and no bedrock outcrop from which it could possibly be derived occurs on the slopes above; it may therefore have dropped from the ice as it dissipated by downwasting. Alternatively, it may be derived from the channel wall now obscured by scree and vegetation. Apart from this one block of doubtful origin there is no evidence to suggest the presence of till in any part of the channel.

The feature called Gleads Cleugh is very similar to that section of the Akeld channel just described and plunges down the Akeld valley side as a 40-foot dry gorge (Photograph 2.c). The feature begins abruptly with an amphitheatre plunge-pool section out of which the floor rises by 15 feet before continuing downslope. Although this hump in the floor profile is covered with vegetation, numerous rock fragments visible amongst the turf and grass suggest that at least part of this 15-foot bar across the plunge-pool probably resulted from scree development. From the plunge-pool downwards the channel floor is completely covered by bare scree fragments and resembles a rock glacier. This material has been derived from the channel sides, which rise almost vertically above the floor, and fingers of scree tail downslope from bedrock exposures on the upper parts of both walls.

If till lies in any part of Gleads Cleugh it is completely obscured by scree and other debris; there is certainly no evidence to suggest its presence.

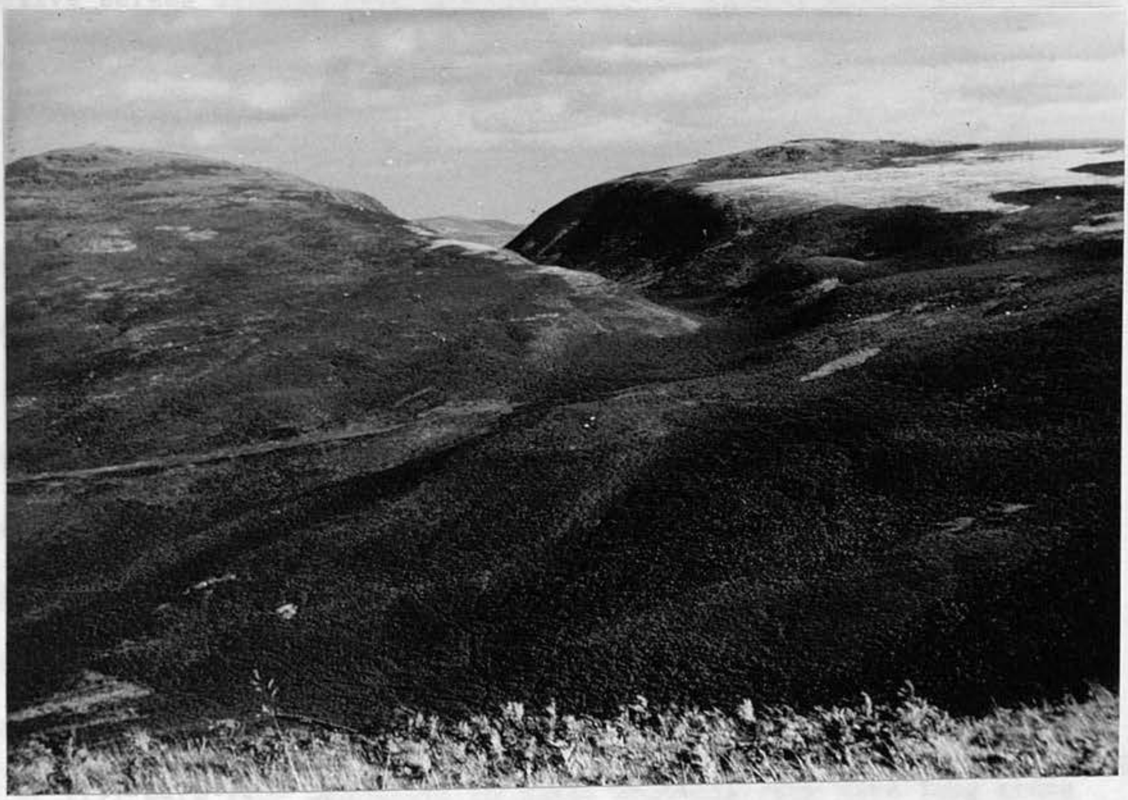
Channel 6: The ridge from Tom Tallon's Crag to White Law forms a divide between two valleys, that of the Akeld Burn and one to the west which opens out northwards. A prominent col in this watershed south of White Law is occupied by a meltwater channel. Beginning with a wide intake whose gentle sides rise up from a flat floor, it quickly narrows and becomes constricted between more precipitous walls rising to a maximum height of about 50 feet. Bedrock is occasionally exposed along this steep-sided section which extends for 340 yards through the ridge crest. Beyond a marked step in the floor profile, leading the channel into a sharply-curving loop section, the feature becomes much broader and shallower as it slopes gently across a bench-like rampart on the valley side. The channel walls are less prominent along this section of its course and finally fade away when the feature terminates with a double, chute type of outlet as the ground drops steeply into the Akeld valley.

Scree development along the first section of the channel has been so extensive that an irregular floor profile with a relief amplitude of about 5 feet is characteristic. Rock fragments up to 4 feet in size litter the floor and lower slopes, but although bedrock outcrops in a few places, any evidence of fresh shattering and screeing involves only small particles a few inches in maximum dimension. The grey, lichen-covered surfaces of larger blocks suggest that they are mainly fossil features and their edges appear to be rounding off under present weathering processes.

Although the longitudinal floor profile of the channel is irregular in detail, a distinct, uninterrupted rise of approximately 20 feet occurs over the first 80 yards from the intake. Beyond this major crest in its profile the channel slopes gently downwards again for the remainder of its length. Scree debris probably contributes to the height locally attained by this crest, but there is an overall upward gradient from the intake to this point which

Photograph 2.d Extensive screes on the south-facing wall of the Harehope channel (7, Map 5).

Photograph 2.e The Humbleton col channel, showing the uphill profile of the main intake (11, Map 5).



cannot be explained by scree infill.

Along the last 460 yards of the channel its floor is extremely flat owing to the extensive growth of peat and marshy vegetation of unknown depth; the gentle sides are also thickly obscured by a vegetation cover.

It is therefore almost impossible to determine whether or not till is present in any part of this channel and since there is no evidence of till anywhere in the vicinity such deposits are considered unlikely to lie hidden beneath other superficial debris.

Channels 7 and 11 are two of the most impressive meltwater channels in the north-east Cheviots. They are sharply incised gorges cut through narrow cols south of Harehope and Humbleton Hills respectively in a most spectacular manner. The precipitous walls of these rock-cut canyons rise over 100 feet above their narrow floors and contrast markedly with the smooth outlines of adjacent hillslopes (Photographs 2.d and 2.e). Along much of their lengths massive screes descend from bedrock walls to obscure the original lower slopes and occupy the floors to unknown depths; well over 10 feet of debris must occur in most places. The screes are developed to their greatest extent on the south-facing sides and are comprised of fragments varying in size from a few inches to several feet across. As in the previous examples there is little evidence to suggest that much of the scree is contemporary, apart from minor fans of quite small material, that frequently descend from places where overlying turf has been removed rather than from outcrops of bedrock, and is evidently debris which already exists in a comminuted state beneath the mantle of vegetation. Processes currently acting on these screes appear to be confined solely to the weathering of individual fragments for their edges, which were presumably angular and sharp originally, are now being rounded off. If quantities of till were ever deposited in these channels they have long since been completely obscured by later debris along the greater parts of their lengths. At the intake and outlet extremities the sides are lower and less steep so that scree development has been less significant, but even here, solifluction and soil creep have obscured the original profiles to such an extent that any till deposits must lie buried if, indeed, they ever existed.

The above are only four examples which illustrate the difficulty of trying to recognise till deposits in the floors of Cheviot "col gullies" in an attempt to establish criteria that will prove them subglacial in origin. Since the majority of other such channels are fully described later in the chapter no further examples will be quoted at this juncture.

Seven meltwater channels in the north-east Cheviots have up/down longitudinal floor profiles which are due only partly or not at all to the accumulation of debris; these are channels 1, 2, 3, 6, 11, 17 and 36 (Map 5). The various explanations put forward in the literature to account for such channels have been fully discussed by Sissons (1961). He concluded that only two of these explanations are likely to be tenable for the majority of such channels in Britain:

(a) The reversal of flow on both a small and large scale; in the former situation local collapse of ice is responsible and may be quite common where a network of subglacial chutes has developed; the latter situation is a much more fortuitous occurrence and is restricted to areas which have been affected by two (or more) ice sheets from different areas.

(b) The subglacial flow of meltwater under hydrostatic pressure. This concept was first considered at length by Sissons (1958) in a paper concerning meltwater channels in southern Northumberland which had previously been mapped in detail by Peel (1951). He quoted evidence from accounts of present glaciers proving that streams are known to flow subglacially

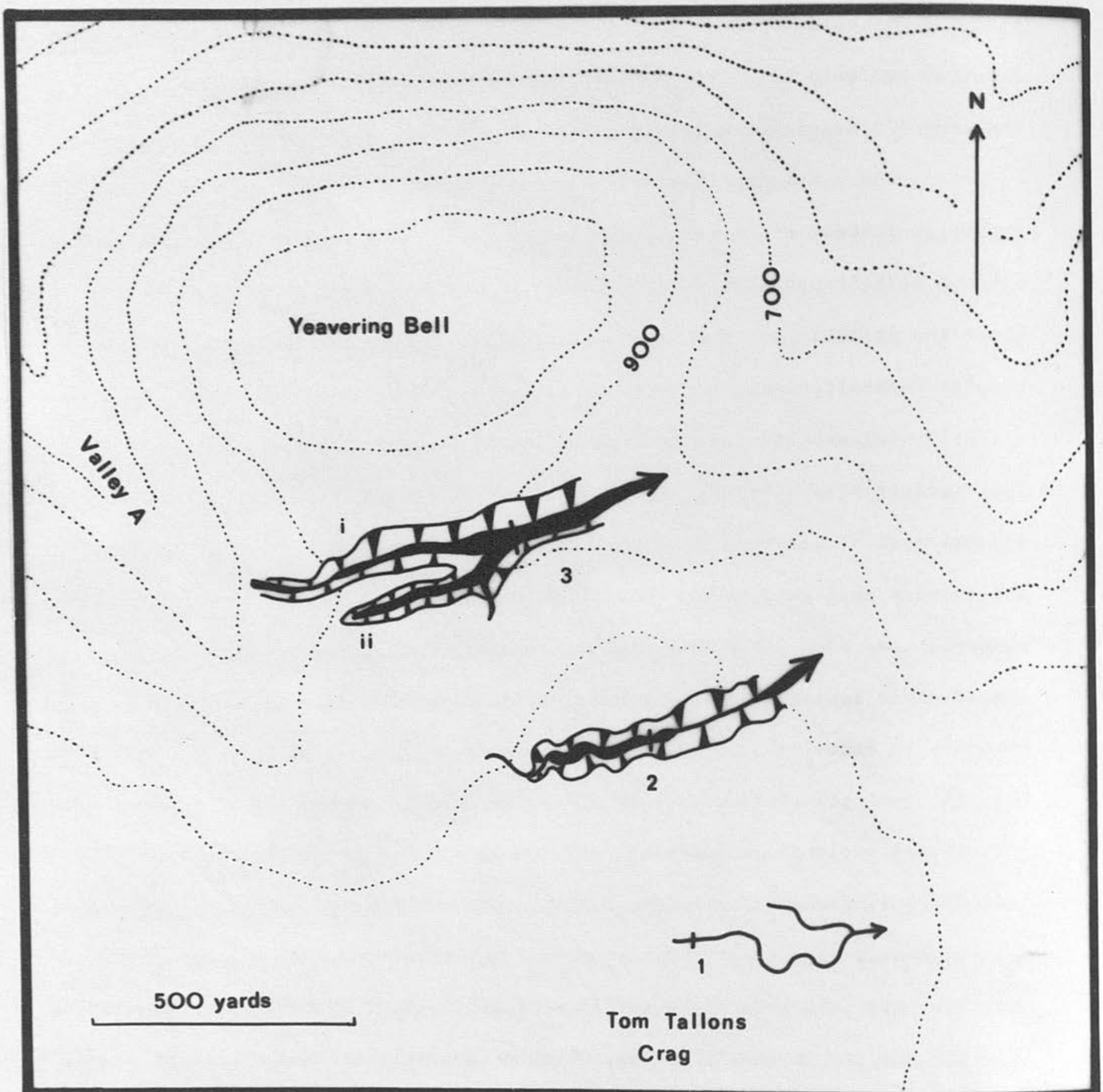

N
Yeavering Bell
900
700
Valley A
i
ii
3
2
1
500 yards
Tom Tallons
Crag

Figure 2.4

uphill under hydrostatic pressure and pointed out how the great meltwater channels of Denmark (Tunneldale) and Germany (Rinnentäler) frequently display up/down floor profiles and are commonly interpreted as subglacially-formed features. The application of this concept has subsequently become quite widespread in Britain, particularly for parts of Wales, England and Scotland. Perhaps the only disadvantage of this theory is that observed by both Peel and Sissons, the latter stating it clearly in the following words; "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." But the substantial weight of evidence from detailed work in Britain since Sissons' earlier papers leaves little doubt as to the probable validity of this concept.

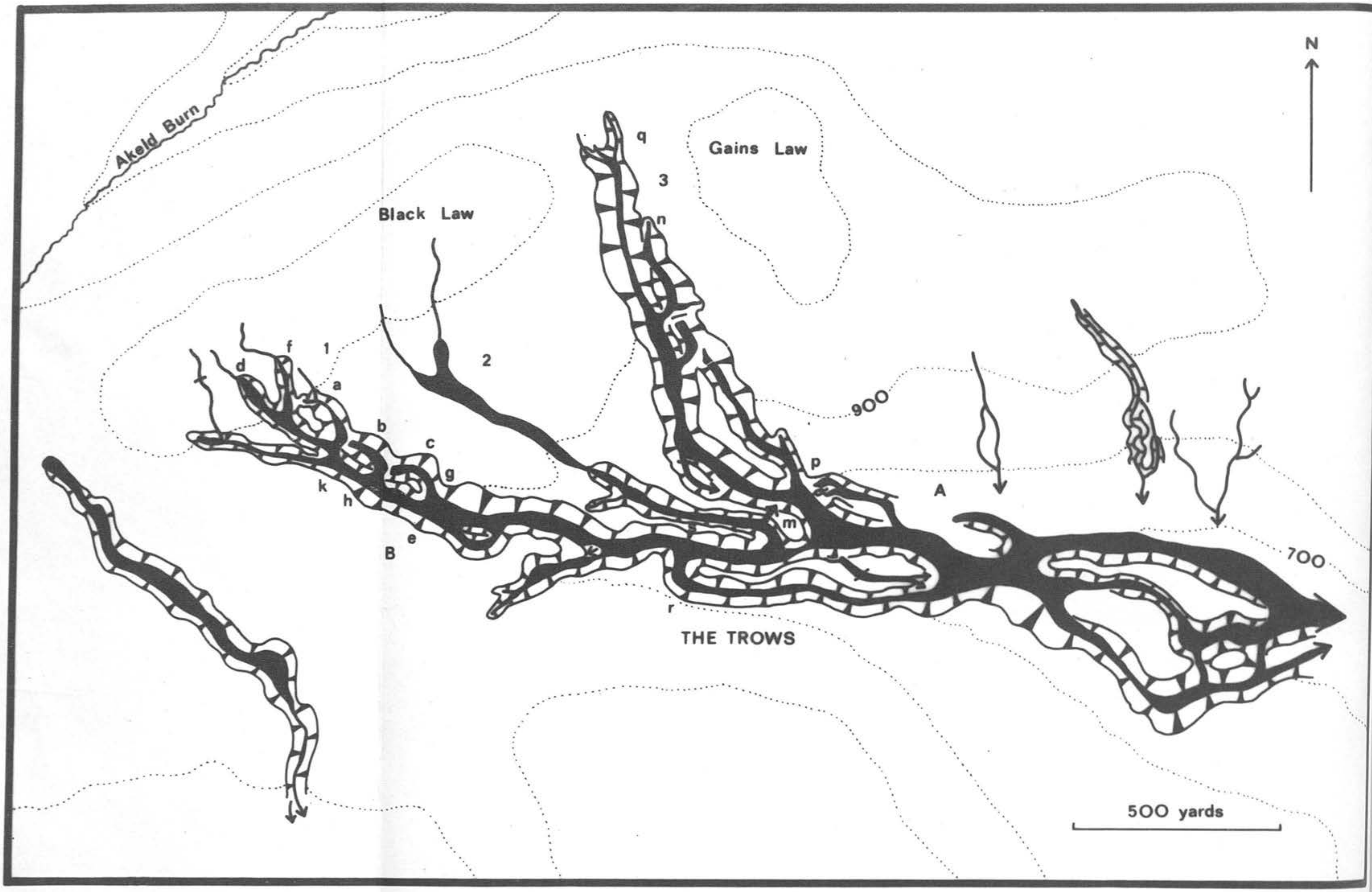
Channels 1, 2 and 3 (Figure 2.4) are located at respectively lower levels on the ridge which extends north-north-west from Tom Tallon's Crag to Yeavering Bell. Number 1 is a small feature no more than 10 feet deep cut entirely in andesite bedrock. From its intake at about 1,125 feet, the channel floor rises a few feet along the initial 40 yards to a crest in its profile precisely mid-way through the shallow col within which the channel is located. It descends the steep eastern flank of the ridge in a somewhat sinuous course and is accordantly joined by a prominent tributary entering from the left just before the outlet at 1,050 feet. There is no accumulation of debris at the channel's outlet, which occurs well above the base of the slope. These characteristics are typical of channels classified as subglacial chutes and there is little doubt that the entire length of channel 1 was cut subglacially. Channel 2 is located in a more conspicuous col lower down the ridge crest at about 1,025 feet. It intakes at 1,000 feet west of the ridge crest with a neatly-defined winding course and soon becomes 20 feet deep. The channel floor rises very gently along the first 300 yards until it is at least 15 feet above intake level. A considerable amount of peat appears to lie in the channel floor and probably contributes in part to the up/down profile, but is unlikely to be fully responsible for it. Beyond the crest the channel narrows and deepens to at least 30 feet as it slopes more steeply down the eastern side of the ridge. It terminates on a valley-side bench at 950 feet, 350 yards from the crest in its floor profile. Channel 3 lies in the deep col south of Yeavering Bell and is comprised, for the most part, of two distinct segments, (i) and (ii) which join in the valley-head east of the col. Segment (i) begins at 825 feet as a distinct furrow on the upper slopes of valley A, well below the level of the col. It climbs quite steeply uphill, becoming wider and deeper and ultimately breaches the col as a 40-foot gorge cut entirely in bedrock. An obvious crest in the present floor profile occurs beyond the confluence point of the two segments, but there has been such extensive peat development at this place that it is unlikely to represent the true crest, which probably lies further west. Even so, the channel floor has climbed uphill for at least 50 feet to a position mid-way through the col and over 400 yards beyond the intake. The cross-profile is asymmetric, the northern wall being about 10-15 feet higher than the southern, and this has resulted from channel incision into the lower slopes of Yeavering Bell above the actual floor of the col. The inter-channel ridge between segments (i) and (ii) continues the line of a reconstruction of the original topography, so that segment (ii) appears to be located in the col floor proper. Rather surprisingly, segment (ii) is the smaller feature whose intake level is about 20 feet above the adjacent floor level of segment (i). It does not exhibit an up/down profile and slopes continuously down from its intake to join segment (i). A wide, chute-like channel enters from the right at this juncture, and the main channel becomes broad and somewhat less distinct as it enters the head-reaches of a pre-existing valley. Channel 3 terminates here at approximately 825 feet. Two distinct channel segments, separated by a narrow bedrock ridge, are thus located in a pre-existing col and clearly formed while ice was also present there, otherwise only one channel would have been created. This evidence supports the contention that the uphill gradient described in segment (i) was eroded by a subglacial river.

Channel 6 has already been described.

Channel 11 (Figure 2.3) cuts through a col on the south-west side of Humbleton Hill as a magnificent gorge 125 feet deep along its central part. Although it is aligned through a col between two pre-existing re-entrants on either side of the Humbleton Hill spur, it is entirely the work of glacial meltwater and there is no evidence suggesting that more than one period of deglaciation has been responsible for it. Three principal intakes feed into the main channel. The lowest and most important begins at 725 feet and climbs quite steeply uphill to meet the other two (Photograph 2.e); the latter are prominent benches cut into the slope of the pre-existing valley-head west of the col. The uppermost has a peculiar double outlet into the main channel; both are steep, chute-like segments that isolate a rock knoll in-between. They are unlikely to have functioned simultaneously and local collapse of ice was possibly responsible for a diversion in meltwater flow. A distinct crest in floor profile occurs beyond the confluence point of all three feeders, and lies over 50 feet above the intake level of the lowest, 20 feet above that of the intermediate one and 20 feet below that of the uppermost. Beyond this crest minor irregularities in the gentle floor profile are produced by the great scree lobes previously described, but nowhere is the general downward slope interrupted. A much steeper profile begins as the channel cuts down the eastern flank of the Humbleton ridge and culminates in a spectacular plunge-pool 60 feet deep. At this point the channel bifurcates round a prominent rock ridge and both segments are obliged to rise 15-20 feet out of the plunge-pool before continuing the normal downslope course and rejoining 300 yards further beyond on the gentle hillside above Humbleton hamlet. A broad and shallow channel continues the line of meltwater drainage across this hillside and becomes accordantly confluent with the intricate network from the northern flank of Humbleton Hill previously described. The latter system has already been interpreted as subglacial in origin and lies at a much lower level. The complex of feeders at the intake of channel 11 can only be accounted for satisfactorily if ice was present during their formation, in the same way as the bifurcation lower down. There is therefore considerable evidence which strongly suggests that channel 11 and its associated features were formed subglacially, so that the uphill gradient of its initial section must have been eroded by water flowing uphill beneath the ice.

Channels 17 and 36, Map 5, are more relevantly discussed in full later in this chapter, but it may be observed here that both display conspicuous up/down floor profiles, each of which is an integral part of the channel eroded by a subglacial stream flowing in one direction.

The up/down longitudinal floor profiles exhibited by the above channels were clearly not produced by reversal of meltwater flow through the local collapse of ice on a small scale; neither were they formed as a result of the interaction of two separate ice sheets, for all the evidence from striations, erratics, drumlins and the alignment of fluvioglacial phenomena proves that only ice from the Tweed valley affected this part of the Cheviots. Some of the channels display characteristics other than up/down gradients that



N
Akeld Burn
Black Law
Gains Law
q
3
n
f
1
d
a
2
b
c
g
k
h
e
B
p
A
m
900
700
r
THE TROWS
500 yards

Figure 2.5

prove them to have formed subglacially and so their floor profiles must have been produced in such an environment, presumably under hydrostatic head of pressure.

The two criteria used by Derbyshire (1961) to demonstrate that almost every channel in the north-east Cheviots was formed subglacially have now been discussed at length and the following conclusions emerge. The recognition of till in channel floors is a doubtful principle to employ in view of the extent to which later debris has accumulated through screeing, solifluction and creep. The presence of up/down longitudinal profiles in channel floors is much more reliable evidence of subglacial formation and the seven channels which possess such phenomena have been referred to. However, these represent only a small part of the very extensive network of glacial meltwater channels in the north-east Cheviots. Those channels lacking up/down floor profiles cannot therefore be assumed subglacial in origin and require further discussion before an interpretation is suggested.

The four major channel systems, 8, 13, 14 and 15/16 (Map 5), exhibit no anomalous gradients in their floor profiles nor do they contain visible deposits of till; but the very complexity of each system merits detailed description and analysis.

System 8: An extremely intricate system of meltwater channels occupies the col and slopes which form the broad head of the Humbleton Burn valley. This system (Figure 2.5) consists of three main branches that join and extend partway down the valley presently drained by the small, misfit Humbleton Burn. The highest point at which meltwater erosion is perceptible amongst the several intakes is approximately 1,040 feet; the definite channel form of the system does not extend below 675 feet. Channel forms vary throughout from intake furrows only 2 feet deep to impressive gorges over 70 feet deep along middle sections of the system. As far as point A the system is incised mainly in bedrock, although a thin veneer of drift may overlie rockhead in some places. Except for the main south wall, which is cut chiefly in bedrock along its entire length, wide channels below A have dissected a drift plug in the pre-existing valley floor. Exposures are extremely poor, but unsorted sub-angular particles, from 3 to 16 inches in size and embedded in a gritty matrix can be observed at a few sites. Lithology is mainly local andesite and the material is probably the "earthy angular drift" of the early Geological Survey officers. Although this drift could possibly be a coarse, unsorted fluvio-glacial deposit, it is probably till since it predates meltwater erosion and forms no definite constructional feature.

Three shallow intakes to branch 1 begin on the crest of Black Law ridge which forms part of the watershed between the Akeld and Humbleton Burns. The two intakes to branch 2 begin on the south-east slope of the ridge just below its crest and three short feeders of branch 3 head on the col crest between Black Law and Gains Law. All feeders begin as minor features only a few feet deep, but each one suddenly expands into a plunge-pool section that varies in depth from 10 feet in branch 2 to 60 feet in branch 3; 20 feet is the maximum depth reached in branch 1. Shortly below these plunge-pool sections the feeders unite to form their respective main branches of the channel system.

Branch 2 is the least complex; two feeders join to form a wide shallow channel that continues for about 450 yards before narrowing and deepening. The only tributary to this branch enters mid-way along; it heads only two or three yards from the north wall of branch 1 and joins branch 2 accordantly. A short distributary "hangs" about 25 feet above branch 3. Of the three arterial stems which comprise the Trows system, branches 1 and 3 present highly complex patterns. Indeed they exhibit peculiarities found in other channel systems in the east Cheviots and elsewhere in Scotland. Their common characteristic consists of several abandoned loops at various levels above the main floor; they occur on both channel walls but dominantly on one. This has the effect of partially isolating rocky knolls varying in height from a few feet to over 40 feet. A similar braided pattern has been described in a channel system south of Edinburgh (Sissons 1963) where some branches are aligned along faults; no faults have been mapped in this area of the Cheviots.

By point B in branch 1 all feeders have united and the channel is about 60 feet deep. The main floor is a marshy flat over 30 yards wide and three isolated knolls separate it from three channel segments cut into the north wall like abandoned meander loops. Loop (a) begins 15 feet above the main channel floor and becomes 10 feet deep before steeply descending to re-join it discordantly; the inlet to loop (b) lies only 3 feet above the main channel floor but the loop deepens to at least 25 feet before curving sharply to drop several feet into the main channel; loop (c) begins 10 feet up, deepens to almost 30 feet and plunges nearly 15 feet into the main channel with an extremely steep profile. Each abandoned segment is the undercut bend of a pseudo-meander loop and it is perhaps significant that the breach through to the main channel occurs at the contiguous "downslope" undercut bend in each case. A possible explanation is that two meltwater streams originally flowed side by side, stream (d-e) and stream (f-g). The "downslope" meanders of stream (f-g) undercut the interstream divide to such an extent that it was eventually breached at point (h) thereby isolating loop (c); the same process would account for loops (a) and (b) while undercutting by the neighbouring stream (d-c) on the other side of the divide probably assisted the process. Loop (c) "hangs" higher than (b), and (b) higher than (a); this illustrates the progressive abandonment of the loops from (c) to (a).

Abandoned segments in branch 3 probably formed in a similar fashion. Again, there appears to have been two streams (q-m and n-p) originally flowing side by side and separated by only a narrow divide. Three feeders lead into (q-m) which occupies the floor of the col and cuts 70 feet through bedrock, while (n-p) lies on the lower slopes of the col and reaches 20 feet in depth. In terms of its volume and location stream (q-m) must have been superior to (n-p) and was able to erode more quickly and deeply. Breaches through the divide, created possibly by an undercutting process similar to that described for branch 1, enforced the progressive capture of water from (n-p) further and further upstream so that lower segments were first abandoned and isolated at higher levels. Eventually all meltwaters were concentrated in the col floor as gravity dictated.

Other small forsaken segments occur in various parts of the Trows system, but perhaps the most intriguing is the largest loop of all, loop (r). Branching off abruptly at right-angles into the right wall of the main channel, the floor of loop (r) climbs about 40 feet to the point where it resumes a course parallel to the main channel. The volume of water which cut loop (r) was sufficient to erode an impressive narrow gorge deepening from 20 feet to almost 70 feet in bedrock. Weathered bedrock is exposed on the interfluve whose crest is only a yard wide at one point. The main channel is nearly 100 feet deep along this section and is rejoined after 1,500 yards by loop (r) descending in a steep profile. The most puzzling feature about the loop is that its intake is a continuous erosional channel sloping steeply uphill from the main channel floor at right-angles to it; there is no evidence that post-glacial processes have been responsible for this section for no alluvial fan lies at its mouth and the present tiny ditch, even in flood, could hardly have completely removed the volume of material excavated from loop (r) intake; this seems especially obvious when account is taken of alluvial fans not far upstream remaining untouched by the present Humbleton Burn. Although the intake to loop (r) is almost in direct line with the flow of meltwater down the Gains Law col, there is no evidence of original flow from this source across spur (s) to account for the intake section. It can only be concluded that hydrostatic head of pressure forced water out of the main channel at right-angles to form loop (r). Exactly why and under what conditions this event occurred is difficult to conceive.

The Trows system of meltwater channels descends directly downslope into the Humbleton Burn valley and cannot be marginal in origin since gradients are too steep. Furthermore, the system is too complex and its very location denies the presence of an ice margin in such a position. Neither can the system have formed proglacially since the concept of downwastage implies that the Black Law ridge and Gains Law would have emerged as nunataks while ice still occupied the Akeld and Humbleton valleys. Evidence which precludes an origin for the system from possible ice-dammed lake waters in the Akeld valley includes the absence of features normally associated with lakes, such as floor deposits, shorelines and deltas, and the fact that several feeding channels intake on the crest of a ridge rather than in the lowest col. Since the system descends approximately 365 feet of altitude it is unlikely to have been formed by streams flowing in open ice-walled channels throughout their length, for crevasses of such dimensions are not reported from present-day glaciers. It is conceivable that upper parts of the system had such an origin while lower segments were subglacially formed, but this theory also meets inherent difficulties. If streams which cut the upper segments of the Trows system flowed in open ice-walled channels, then all the interfluves and isolated knolls must

Photograph 2.f The main gorge of the Horsdon channel (13, Map 5).

Photograph 2.g The main tributary of the Horsdon channel.

Photograph 2.h The Horsdon channel; view from the south-east.

Photograph 2.i The Horsdon channel, central section of the main gorge.

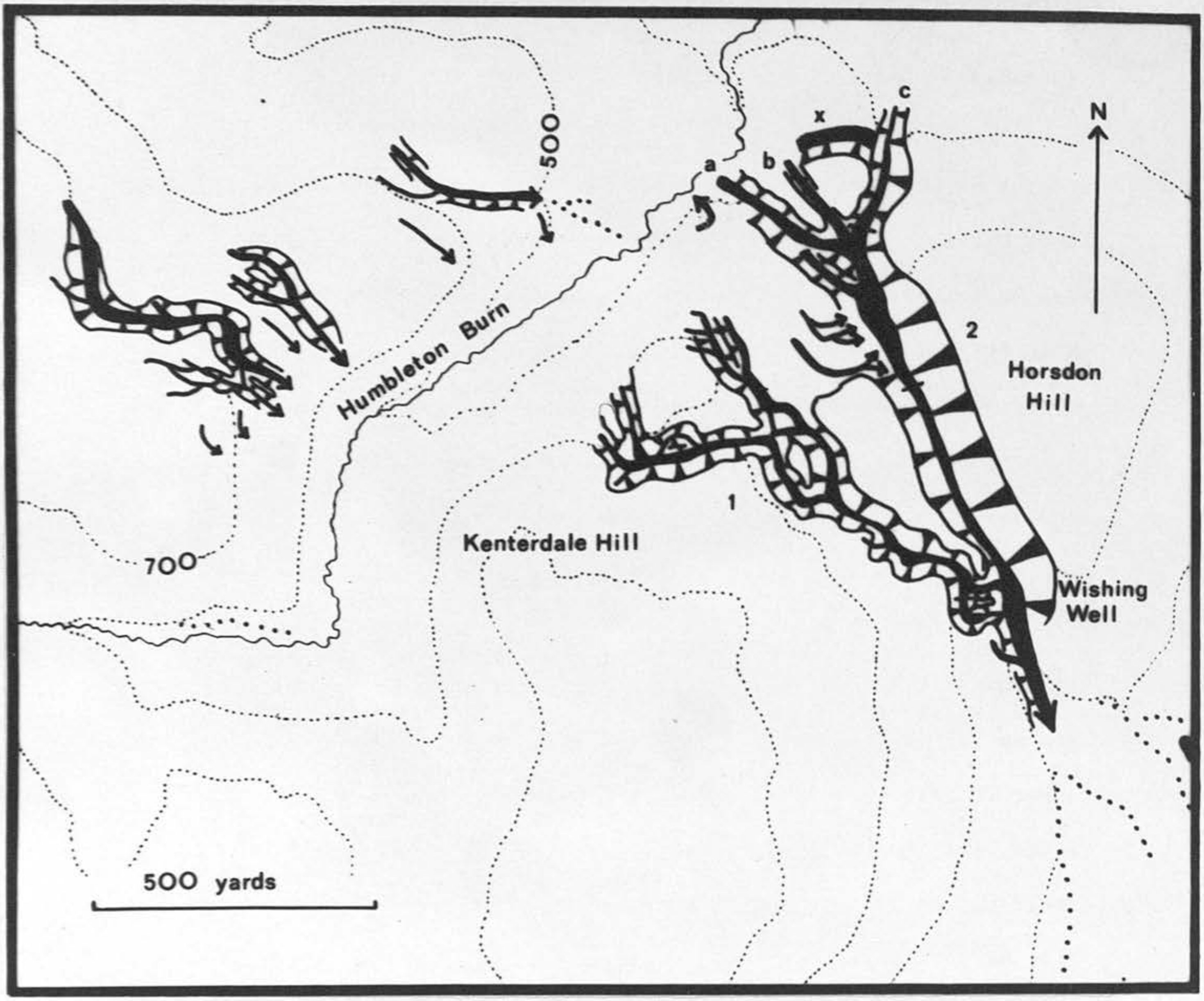
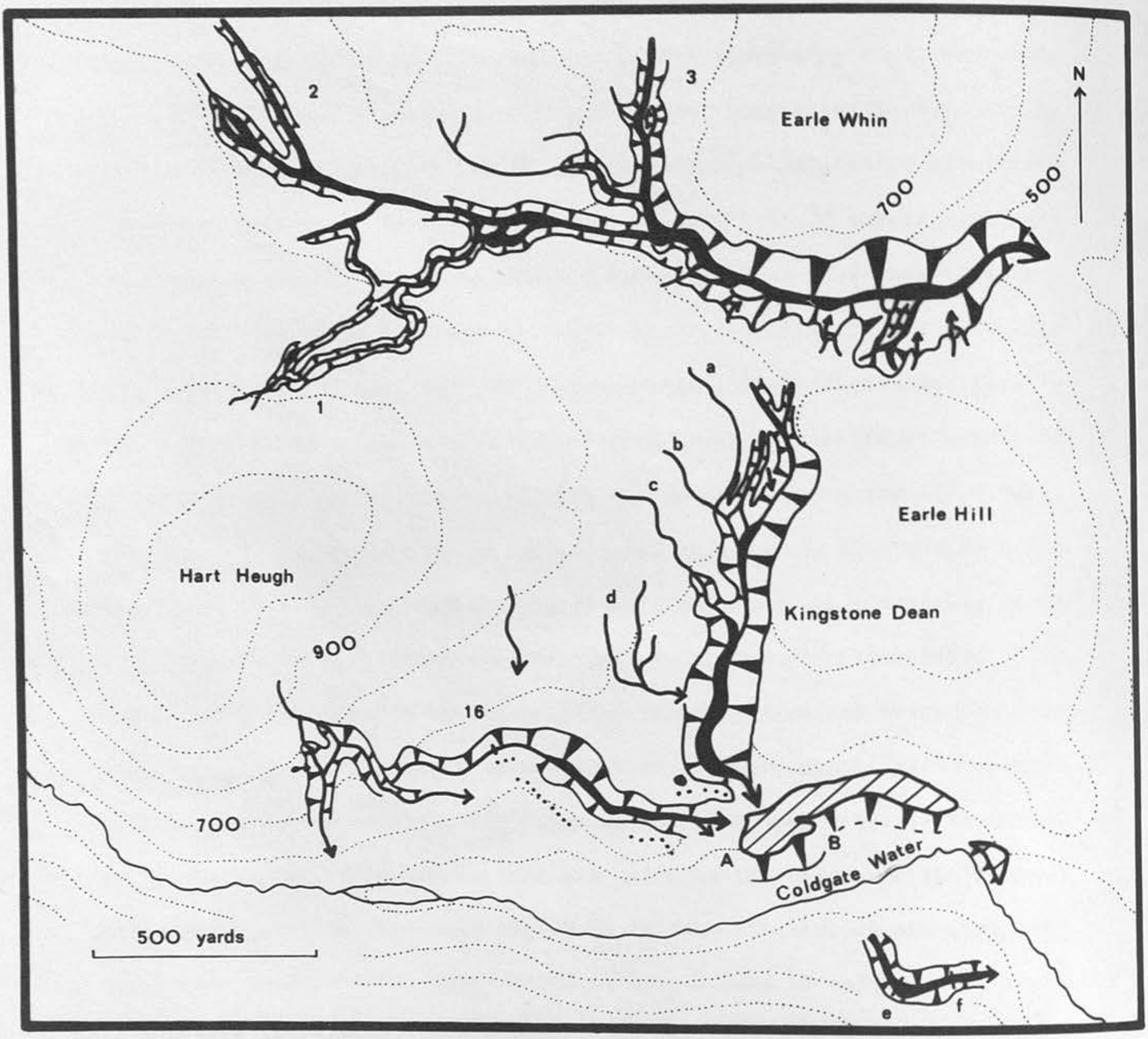

N
500
700
a
b
c
x
1
2
Humbleton Burn
Kenterdale Hill
Horsdon Hill
Wishing Well
500 yards

Figure 2.6

have retained narrow pinnacles and wedges of ice perched precariously on their crests; such situations are highly improbable and well-nigh impossible.

Since the various complexities of the channel system require the presence of ice during their formation, and in view of the foregoing discussion, meltwater streams must have flowed in tunnels beneath the ice. A subglacial origin for the system is thereby established.

System 12: The broad, shallow col between Kenterdale Hill and Horsdon Hill at an altitude between 400 and 500 feet forms part of the watershed between the Humbleton Burn and the Wooler Water. It is breached by a composite channel system (Figure 2.6) whose multitude of feeding intakes join to form two main gorges which meet at the Wishing Well near the system's outlet (Photograph 2.f). The highest intake lies at 600 feet and the outlet at 400 feet. Following an intricate, reticulate network in which knolls are isolated, the feeders unite as branch 1. This slightly sinuous feature has an irregular floor profile and becomes a narrow, steep-sided gorge cut almost 70 feet into bedrock at maximum depth (Photograph 2.g). It joins branch 2 with a double outlet, where one of the segments is an abandoned, high-level loop isolating a rock knoll over 10 feet in height. Branch 2 is one of the most impressive channels in the east Cheviot area. It slices through the spur of land between Kenterdale Hill and Horsdon Hill as a majestic rock-cut canyon whose walls generally rise 80 feet above the floor; along one section the east wall stands over 100 feet high (Photographs 2.h and 2.i). Three main feeders are tributary to branch 2 and three minor chute-like gullies hang high up on the right bank. The latter appear to have been of limited significance and were probably abandoned early. The relationships of the feeders to the main channel and to each other are extremely interesting. Feeder a begins on the lip of Humbleton Burn valley as a 15-foot channel cut in bedrock. Fluvioglacial hummocks and vague depressions on the opposite side of the valley may also be connected with this line of drainage. It enters the main channel with a double outlet where the southernmost artery hangs by 6 feet at its proximal end above the northernmost artery and was forsaken as the latter continued in use. Both slope steeply down to the main channel. Feeder b lies 6 to 10 feet lower in altitude and although broad and shallow, it is well defined; the west wall stands 15 feet high at one point while the east wall declines from 6 feet to disappear as the channel becomes a bench. Feeder b terminates at its junction with feeder c, above which it hangs by 10 feet. The short channel segment at point x may have been the original intake of feeder c for it curves round towards it as a 30 foot wall, but hangs by 20 feet above its floor. Feeder c begins about 100 feet below the point where it meets with a and b in the main channel and its gradient up to this point is extremely steep. It is possible that this uphill section was eroded by water flowing upslope under hydrostatic pressure, but the profile gradient and the height climbed seem excessive compared with the uphill sections of other channels in the east Cheviot area; a more logical explanation is apparent. Feeders a, b and c were most likely contemporaneous to begin with, feeder c having a downslope gradient at this time during which the topmost parts of its walls were cut; in addition, the southernmost outlet of feeder a would have been functioning. The system appears to have persisted sufficiently long to have eroded the vast gorge of branch 2. Ultimately, at a later stage, it appears that the route through branch 2 was abandoned, perhaps due to blockage by collapsed ice. Water from feeder a cut a new outlet northwards thereby reversing the flow along feeder c down to the Humbleton Burn valley. Subsequently, water following this new route cut 10 feet below the outlet of b, which had probably stopped functioning, and 20 feet below the original feeder x thereby truncating



2
3
N
Earle Whin
700
500
1
a
b
c
Earle Hill
Hart Heugh
d
Kingstone Dean
900
16
700
A
B
Coldgate Water
500 yards
e
f

Figure 2.7

their distal ends. This explanation accounts for the apparent reverse gradient in the floor profile of the Horsdon channel. Another apparent crest in its floor profile, one-third of the way along, is explained by scree infill.

Applying arguments similar to those expounding the Trows system, it is concluded that the Horsdon system also can only be explained by subglacial flow of meltwater. Indeed, the minor chute-like gullies cut into the right wall of branch 2 must be subglacial in origin for they terminate before reaching the base of the slope and no detrital fans exist below their outlets. They are presumed subglacial chutes and since they are incised into the walls of a channel already in existence, then this channel must also have formed subglacially, for there is no evidence that suggests it was formed during a previous glaciation.

System 13: The Earle system of meltwater channels (Figure 2.7) is incised into the hillslopes and cols leading to and including the short pre-existing valley between Earle Whin and Earle Hill. Here the main channel attains its greatest dimensions, precipitous rock walls towering over 100 feet above a narrow floor presently drained by a small misfit stream trickling down its steep gradient. Preliminary signs of meltwater erosion associated with the system appear on the northern slopes of Hart Heugh at an altitude of 900 feet; the system terminates at 450 feet. The waters of three major tributary systems appear to have combined and followed the pre-existing drainage line of Earle valley.

Group 1 crosses the contours of Hart Heugh hill obliquely; a diverging-converging network and considerable sinuousity are characteristic of the system which is incised by 25 feet in places. These attributes can only be accounted for by subglacial flow down the hillside. Group 2 consists mainly of two converging channels which trend almost parallel to the contours along the hillside leading over from the Humbleton Burn valley. They attain depths of up to 15 feet before fading away in a shallow marshy depression. Definite walls reappear as the channel form resumes prior to its accordant junction with group 1. Two knolls are isolated by abandoned loops before a 20 foot gorge leads group 2 into the main Earle channel. Group 3 begins in the col leading south from the Humbleton Burn valley. A vague crest in its floor profile near the intake is most likely due to solifluction and creep from surrounding hillsides. From its shallow intake the channel rapidly deepens to a 40 foot rock gorge and is joined by two chute-like tributaries on its right bank. A further peculiarity is the rock knoll formed by an abandoned loop hanging 15 feet above the main branch.

The main Earle channel itself displays several isolated rock knolls in its floor and on the lower slopes of its south wall; but perhaps most significant are the numerous dry gullies which furrow the upper slopes of the south wall imparting a crenellate aspect to its outline. The gullies vary from shallow features 5 feet deep to more impressive forms incised by nearly 20 feet. No deposits spread from their outlets and they normally terminate above the floor of the Earle channel. They are most logically explained as subglacial chutes.

Tributary system 1 was formed subglacially and the complexities of systems 2 and 3 are explained only by subglacial flow; the main Earle section is furrowed by subglacial chutes and so must also be subglacial in origin. The entire Earle system of channels was therefore eroded by meltwaters flowing subglacially.

System 14/15: The broad col between Hart Heugh and Earle Hill contains a deeply incised network of meltwater channels (Figure 2.7). The main gorge, called Kingstone Dean, is located in the col floor and descends quite steeply southwards to its outlet which lies over 100 feet above the Coldgate Water. Beginning at 700 feet with a double intake, each of which is only a few feet deep, the channel rapidly deepens until its east wall rises 70 feet above the rock-strewn floor. The cross-profile is asymmetrical along the greatest part of its length; the higher, east wall is cut entirely in bedrock whereas the opposite wall may be partly in drift. A short tributary branch is aligned parallel to the main feature on the right. Beginning a short distance downstream from the two intakes, it lies about 15 feet above the main channel floor. It plunges quite steeply into the latter and was probably abandoned a considerable time before Kingstone Dean ceased to function. A narrow ridge of bedrock lies in between. Tributaries a, b and c cross the moderately steep flanks of Hart Heugh at an angle oblique to the contours of the slope and descend into Kingstone Dean with sharply incised chute-like sections. The junctions appear to be relatively accordant. Tributary d is a minor system within itself, for three shallow channels cross the hillside obliquely to unite as one feature that slopes directly downhill towards Kingstone Dean; the tributary terminates on the upper part of the right wall.

This somewhat complicated pattern of meltwater channels requires the immediate presence of ice during its formation for streams to have assumed such courses. Since the system begins at 700 feet and terminates at 525 feet, these courses are unlikely to have been open ice-walled, and a subglacial explanation is more satisfactory. The difficulty of narrow wedges of ice on inter-channel divides is also obviated by this interpretation.

It may be argued that Kingstone Dean predates the main gorge of the Earle Hill system since it lies at a higher level only 200 yards from the latter. Although no connection between the two systems is evident in the field, meltwater might have initially flowed from some of the higher feeders to the Earle system round the north-east flank of Hart Heugh to join Kingstone Dean. The absence of definite evidence makes this suggestion speculative, however, and a more probable interpretation for this peculiar juxtaposition of channels aligned at right-angles to each other is either, (a) Kingstone Dean was formed earlier than the Earle channel and was later abandoned when the latter system became established; or (b) the two systems functioned simultaneously, meltwater streams feeding the Kingstone system flowed englacially at a higher level than the subglacially-engorged Earle system. The latter was controlled by local topography to assume its particular alignment. The latter theory of several meltwater streams operating contemporaneously in close juxtaposition is perhaps supported by the remarkable concentration of equally composite systems in adjacent areas.

The outlet to Kingstone Dean is partly cut through a ridge of fluvioglacial deposits, and at A a delta-shaped mass of sands and gravels, rising over 100 feet above the present floodplain of Coldgate Water, seems obviously related to the channel, although Derbyshire (1961) considered that it "bears no resemblance to a water-laid delta". While this deposit has been mentioned in previous literature, little attention appears to have been paid to adjacent fluvioglacial features to the west. Here, two subglacial chutes, cut almost 25 feet deep in bedrock plunge down the southern slope of Hart Heugh. The more easterly seems to have truncated an earlier line of meltwater drainage trending in sinuous fashion along the hillside at a small angle to the contours. At first it is a rock-cut bench, but subsequently becomes a two-sided channel incised through sand and gravel ridges apparently associated with the same line of drainage. This channel terminates close to the outlet of Kingstone Dean, with which it seems to have been contemporaneous, and similarly merges into the fan-like mass of deposits. Superficial exposures reveal sand, gravel and cobbles which Derbyshire observed to be arranged in contorted beds resting on till and capped with "superimposed glacial till". Careful mapping of these deposits has revealed that the fan-like plan of the mass is largely related to a sharp incision made by a small stream at B, for beyond this gully a level-topped spread of fluvioglacial deposits continues down-valley at a similar elevation. Indeed, the so-called "fan" is a partially truncated fragment of this terrace feature which is very probably related mainly to channel 16, west of Kingstone Dean.

On the spur of land rising steeply above the south bank of Coldgate Water a small, rock-cut meltwater channel describes a remarkable course which has stimulated considerable speculative comment in previous literature on this area. In the form of a loop, this channel intakes at just below 525 feet and curves uphill as a well-defined, two-sided feature; the south wall rises higher than the north and reaches about 15 feet at maximum development. The channel floor rises quite steeply by almost 20 feet to point e beyond which it follows a short, level course; at point f it rises 8 feet to a second crest before sloping down to rejoin the Coldgate valley, on whose side it terminates at 500 feet. Both Burnett (1932) and Common (1957) proposed that waters from Kingstone Dean flowed across valley ice to cut this arcuate channel in meander fashion, but offered no explanation for its uphill gradients. Derbyshire (1961) refuted that hypothesis, arguing that the channel flows westwards and can in no way represent a former extension of Kingstone Dean. He advocated it to be of later date and tentatively suggested that it may mark a low ice-edge lying in the Coldgate valley. This suggestion is extremely difficult to appreciate since no other evidence points to such a situation. Moreover, it is very peculiar that while all other features of meltwater drainage in the area flow in a uniform direction, this arcuate feature should have been eroded by water flowing in the opposite direction. The location of this channel in relation to Kingstone Dean seems highly significant, and when it is observed how the latter terminates at 525 feet while the former intakes at a little below that level, a definite relationship becomes more apparent. The most acceptable explanation for this feature is that it does represent a segment of the meltwater drainage system responsible for Kingstone Dean, as Burnett and Common have already suspected. The uphill gradients displayed by its floor profile, however, exclude the subaerial interpretation envisaged by the two previous writers, and can only be explained by subglacial flow of meltwater. The most probable explanation is that the meltwater stream flowed englacially across the Coldgate valley and became superimposed on the pre-existing slope opposite the outlet of Kingstone Dean. The original floor gradient was probably continuously downhill, but as meltwater was able to cut more easily down through ice occupying the valley between the channel and Kingstone Dean, it became progressively difficult for the stream to be forced uphill along this arcuate channel until it was finally abandoned. Consequent on the opening up of a new englacial route down the Coldgate valley, the meltwater stream in Kingstone Dean seems to have ceased its erosive phase and commenced deposition. This is suggested not only by the presence of fluvio-glacial gravels associated with the outlet of Kingstone Dean, but also by the fact that the Kingstone Dean outlet has not been cut below the abandoned intake level of its arcuate continuation on the opposite valley-side.

It is therefore suggested that the entire suite of fluvioglacial features converging on this part of the Coldgate valley was formed sub glacially.

The Development of Subglacial Drainage

In addition to channels whose up/down longitudinal floor profiles of necessity imply hydrostatic head of pressure during formation and consequently subglacial flow of meltwaters, the four most intricate channel systems in the north-east Cheviot area have been shown to be subglacial in origin with reference to other inherent characteristics.

In recent literature on deglaciation frequent reference is made to drainage channels cut by subglacial meltwaters. This interpretation is often based on such evidence as the presence of up/down longitudinal floor profiles, implying hydrostatic pressure during formation, and quantities of till in rock-cut channel floors. Nevertheless, the means by which these streams became subglacial has not always been clearly emphasised.

Subglacial streams may develop in any of the following ways:

(a) Water from ice-free hillslopes, together with ice margin meltwater, plunging downslope beneath the ice. Streams forming in this way may cross the contours obliquely or go directly downslope as subglacial chutes. If the ice is less than 300 to 400 feet thick subglacial streams may penetrate to the base of the hillslope and continue to flow subglacially. Subglacial channel systems originating in this way on valley floors and broad vales are usually recognised by their relationships to chutes and oblique downslope channels tributary to them. Where thicker ice is known to have existed, such downslope streams seem frequently to have turned along the hillslope to flow parallel or at small angle to it. Alternatively, instead of being diverted along the hillside, the downslope streams may continue in courses entirely within the ice leaving no further trace on the ground.

(b) Water from supraglacial and/or englacial streams descending directly beneath the ice by way of crevasses or moulins. It is doubtful if crevasses and moulins in present-day glaciers generally penetrate to a depth of more than about 300 feet. Indeed, Glen (1953) noted that "crevasses of much greater depth than 20 m. are rare in a temperate glacier". Consequently, during final stages of downwasting, the deepest crevasses and moulins presumably reach the ground only when the ice is considerably less than 300 feet thick, enabling some water to reach the ground and flow subglacially beneath relatively thin ice above. There would be little hydrostatic pressure exerted in the majority of such streams since any crevasses and moulins penetrating stream tunnels would allow release of pressure. Accordingly, this interpretation is unlikely to apply to the majority of glacial meltwater channels with up/down floor profiles which require water flowing under considerable hydrostatic pressure to be adequately explained.

(c) Carey and Ahmad (1961) have demonstrated that "Subglacial channels can only develop from the seepage outlet of the meltwater at the terminus of the glacier." Analogy is drawn with water-saturated sediment where seepage is known to produce open channels well below the surface. Consequently, the writers imply that open subglacial channels may develop by this process of seepage under hydraulic gradient pressure "continuously back from the terminus of the glacier". While subglacial river systems originating in this fashion may have cut meltwater channels on low-lying ground of gentle relief, this interpretation seems inadequate to explain meltwater channels cut through spur crests and cols in areas of varied and considerable relief. The two previous explanations (a and b) are also considered unlikely to apply in such areas, for it is unreasonable to envisage subglacial streams flowing through several hundred feet of altitude up and down spurs and cutting channels only in cols or on the crests.

(d) Since the concept of ice downwastage during deglaciation in Britain is now well-established, there is no need to further elaborate the point at this stage. It is therefore reasonable to expect that meltwater drainage in the upper zones of the ice mass may become superimposed on to the ground beneath as the ice surface lowers in altitude. Englacial streams become subglacial by superimposition and they do so at their point or points of contact with the ground. It is thus possible for a stream to be subglacial in parts of its course and englacial in others. It is doubtful if any meltwater stream is likely to become superimposed by cutting vertically downwards continuously through the ice for many hundreds of feet. The average maximum depth of meltwater penetration in any ice mass is probably controlled by the glaciological properties of each particular mass. Glen (1953) calculated that meltwater should be able to penetrate to a depth of approximately 600 feet in a temperate glacier, but in Scotland, Sissons (1963) has observed "many channels and some eskers ..... 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". Evidence from the Cheviots endorses the observation which may, of course, be valid only for glaciological conditions peculiar to the British environment during the Pleistocene. This approximate value of 400 feet for maximum subglacial penetration by meltwater implies an upper zone of meltwater flow below which persists glacier ice impermeable to meltwater at any one time, or where any tunnel opened by meltwater is quickly closed by glacier flow before a permanent channel is established. As a mass of glacier ice downwastes there is progressive lowering of its surface. Associated with this, a lowering of the base level of subglacial penetration by meltwater (approximately 400 feet below the ice surface) is to be expected. The writer thus visualises englacial streams flowing in this zone of meltwater flow becoming superimposed as the zone impinges on underlying topography.

It is therefore concluded that the majority of subglacial meltwater channels in cols and spur crests were cut by streams that became subglacial by superimposition from former englacial courses.

Although the superimposition of meltwater streams on to emerging bedrock ridges was briefly discussed in 1899 as a possible means by which some channels might form, it appears not to have been applied to channels cut across spurs in Britain until developed by Price in 1960. Following detailed work in the upper Tweed valley Price concluded that as the ice thinned the supraglacial and englacial streams became superimposed on the spur crests beneath. The original drainage, confined entirely upon ice-floored channels, was thus disrupted and became subglacial on the further side of each spur. At about the same time as Price's publication Embleton (1961) independently arrived at a similar conclusion for certain channels in north Wales, but did not fully develop his theory. Subsequent work in Scotland by Sissons further endorsed the superimposition concept and recent publications by Bowen and Gregory (1965) mention its possible validity in south Wales. The writer also has recently stressed the importance of this concept to an interpretation of some meltwater channels in north-east Scotland (Clapperton 1966).

In Price's paper considerable emphasis was placed upon the former persistence of supraglacial drainage along a marginal zone of the Tweed glacier, but since the majority of channels in the north-east Cheviots were formed subglacially along their entire lengths, supraglacial streams evidently played an extremely minor role in that area. Indeed, the extensive development of such drainage on the surface of a glacier decaying over highly irregular underlying topography may be questioned. Although supraglacial drainage is observed on most present-day temperate glaciers, the majority of these streams have relatively short-lived courses on the surface and soon disappear down crevasses and other fissures in the glaciers to assume englacial and/or subglacial positions. It is therefore suggested that in areas where dense networks of meltwater channels are cut across spurs, as in the upper Tweed valley and north-east Cheviots, the superimposed meltwater streams responsible for them are more likely to have been englacial rather than supraglacial. The englacial streams then became disrupted but assumed subglacial positions along only short portions of their total lengths.

The above theory applies particularly to the north-east Cheviots where the majority of channels and channel systems so far described are located either in cols through spurs or in pre-existing valley-heads. The channels are conspicuously aligned in a uniform direction and although they might appear to be arranged as an interlinked sequence on successive spurs, the channel on one spur having been cut by the same stream responsible for channels on succeeding spurs, this is only a false impression which can be gained from a very small-scale map. It has been illustrated quite clearly that in most instances the outlets of channels across any particular spur are normally at considerably lower levels than the intakes of channels on spurs next in procession. The only exceptions are likely to be the possible connections between (Map 5), (a) channel 2, outlet at 950 feet, and channel 6, intake at 850 feet; and (b) channel 4, outlet at 1,000 feet, and channel 9, intake at 950 feet; but the connections remain conjectural. In almost every instance the channels terminate on the lower slopes or floors of pre-existing valleys; erosional form ceases suddenly, quite often before reaching the base of the slope. Such abrupt terminal points of erosion are interpreted as marking the lowest subglacial limit of meltwater erosion during the formation of the particular channel concerned. The streams may have resumed englacial courses and escaped out of these small valleys to join an extensive meltwater drainage system in a submarginal position within the Tweed glacier, where they probably entered the general flow of fluvioglacial drainage south-eastwards through the Wooler Water depression and onwards to the Hedgeley Basin, in which deposition occurred on an immense scale. The levels at which these streams flowed were probably controlled by prevailing englacial water-tables. Alternatively, and less likely, is the possibility that the streams continued to flow subglacially but with no erosive action, perhaps occasionally as disseminated sheet-flow as opposed to concentrated flow in channels.

It is perhaps reasonable to assume that all the systems of meltwater channels in the north-east Cheviots did not form at the one time. For example, the table below illustrates the altitudinal difference between the highest intake and lowest outlet of channels on the major spurs.

i)	Yeavering Bell	- 1,125 feet and 825 feet:	difference = 300 feet.
ii)	Akeld	- 950 feet and 550 feet:	difference = 400 feet.
iii)	Harehope	- 1,040 feet and 500 feet:	difference = 540 feet.
iv)	Humbleton	- 1,040 feet and 500 feet:	difference = 540 feet.
v)	Horsdon	- 900 feet and 400 feet:	difference = 500 feet.

The highest and lowest channels on any one spur are therefore unlikely to have functioned contemporaneously in view of the depth to which subglacial water has generally been proved to penetrate. Even so, the presence of considerable channel systems at similar elevations on each spur - quite unconnected to one another - proves that very extensive networks of englacial drainage permeated marginal zones of the Tweed glacier as it downwasted over highly irregular topography peripheral to the north-east Cheviots. The highest meltwater channels probably formed first, although on any one spur a channel, say 100 feet lower than the one above, might possibly have been superimposed before the higher one if it had originally occupied a lower englacial position than the latter. Any channel interpreted with the superimposition hypothesis cannot therefore be used per se to establish former glacier limits; but it will be suggested later that the upper limit to a marked concentration of these channels may be of some value to approximatley deduce the minimum height of glacier incursion.

The Formation of Crawley/Shawdon Dean

The drainage direction indicated by the large channel network just described illustrates how enormous volumes of meltwater flowed south-eastwards into the area of Hedgeley Basin. The south-east margin of this basin is extremely well-defined by a steep escarpment of Fell Sandstone, rising to over 750 feet in Titlington Pike, and an impressive ridge of sandstone from the Cementstone group. The latter continues the bounding rim of high ground south-westwards and swings through Glanton to merge with the steep foothill zone of the volcanic massif. Several cols occur on this relatively narrow watershed between the rivers Breamish and Aln (Maps 6 and 8). From west to east the lowest of these are as follows:

(a) between The Mile and Mile End, lying at a height of between 450 and 475 feet.

(b) between Mile End and Glanton Pike, lying at a height of between 450 and 475 feet.

(c) between Glanton Hill and Crawley Tower. A deep meltwater channel has considerably lowered the pre-existing col, but a reconstruction of it lies at between 425 and 450 feet.

(d) between Beanly Moor and Eglingham Moor, lying at 400 feet. Cols (a) and (b) show no signs of having accommodated glacial meltwater at any time. Col (d) is the lowest and might be expected to contain some evidence that meltwaters escaped from the Hedgeley Basin by this route, but apart from several small ridges of fluvioglacial sand and gravel, which actually slope north-westwards into the Hedgeley Basin, there is no obvious evidence that this col performed an important function in the evacuation of meltwater. Although the pre-existing col at Crawley appears to have been slightly higher than that at Eglingham, the presence of a gigantic meltwater channel in its floor strongly suggests that this col controlled the escape of meltwater from the Hedgeley Basin and consequently the whole régime of fluvioglacial activity in that area. The topography about Crawley, Shawdon and Glanton indicates that two pre-existing valleys head into the col from opposite sides of the ridge; the larger opens out to the Aln valley while the smaller appears to have been a scarp-face gully formerly joining the Breamish at Powburn. The two valleys are now linked by the huge Crawley Dean meltwater channel.

The Crawley Dean channel (Maps 8 and 9) has been previously mentioned by several authors, the majority of whom explain its origin in terms of overflowing water from an ice-dammed lake in the Hedgely Basin. Derbyshire, however, linked it with his "subglacial col gully" category. Contrary to what Derbyshire assumed, there is abundant evidence to prove the former existence of an ice-dammed lake in the Hedgeley Basin and this will be fully discussed in a later chapter. The writer also fully agrees that Crawley Dean almost certainly functioned as an overflow for water from this lake. However, there is also considerable evidence suggesting that the Crawley Dean channel in its entirety was originally formed in a different manner, and that overflowing lake water subsequently effected little or no modification. Firstly, all evidence in the Hedgeley Basin points to a lake level at or about 300 feet, and although this level may at times have been below 300 feet, there is no evidence that it exceeded this height by more than 25 feet at any time. Since meltwater incision into the pre-existing col began at slightly above 425 feet, it is difficult to reconcile channel cutting with a 300-foot lake level. Secondly, the longitudinal floor profile of the channel rises quite distinctly from approximately 275 feet at the intake to between 300 and 325 feet at a crest in the profile 850 yards further on. From this point the channel floor descends regularly to its outlet at between 175 feet and 200 feet in the Aln valley. Although a small alluvial fan has been built out by a stream entering on the left side at x, this is over 100 yards down-channel from the profile crest and is in no way responsible for it. A tiny stream of water from the right bank does enter almost at the crest, but it is so small that very little debris is associated with it and it has eroded only a shallow gully in the channel wall. There is little doubt that the up/down longitudinal floor profile of the Crawley Dean channel is original and was created by the meltwater river which formed the channel. Consequently, it is necessary to invoke subglacial drainage under hydrostatic pressure to adequately explain this phenomenon. The channel was therefore cut almost entirely in this fashion and the preservation of its up/down profile suggests that subsequent modification by water escaping from an ice-dammed lake to the north was extremely slight.

The Crawley Dean meltwater channel is the largest of such features in the east Cheviot area. It has been traced over a distance of four miles and is almost 400 yards wide at maximum development. Sharp, channel-like form is best preserved where it cuts through the Cementstone ridge; here the feature is over 125 feet deep and cut entirely through bedrock. As the channel emerges from this gorge section into the area of pre-existing valley development, large undercut meander bends appear, and the marshy floor, densely covered with tangled undergrowth is over 30 yards wide. Beyond this section, pre-existing slopes are gentler and channel incision has been much less spectacular. Nevertheless, the east wall has been actively undercut and stands 50 to 60 feet high opposite Lincomb Cottage. From this point onwards the channel is, at first partly and then entirely, cut into a flat-topped spread of sands and gravels that slopes gently as a fan-like terrace into the Aln valley. Gully (i) has already been referred to and enters on the left bank over 100 yards beyond the crest in the floor profile of the main channel. Cut nearly 70 feet into rock at its maximum, this gully heads in a pre-existing depression. It appears to have been eroded chiefly by meltwater for there is a very small catchment area at its head and the present stream is quite insignificant. Much debris must have been removed during the excavation of this deep gully but only a small alluvial fan lies at its outlet into Crawley Dean. This suggests that the detritus was removed during a period when much larger volumes of water flowed through (i) and Crawley Dean. Gully (ii) is a narrow little tributary presently occupied by a tiny stream of water. It is extremely difficult to deduce whether or not meltwater has been wholly or partly responsible for this gully and the writer prefers to remain uncommitted on this feature (which is quite insignificant in the general pattern of fluvio-glacial drainage anyway).

Since no other meltwater channel cuts through the Breamish-Aln watershed (except for a tiny feature at 500 feet in the col between Tick Law and Harehope Hill, Map 6), it is proposed that Crawley Dean was responsible for the evacuation of all meltwater from the Hedgeley Basin when the level of meltwater drainage through the ice in this area was below 425 feet. Previous to the inception of Crawley Dean, meltwater streams flowing from north to south presumably crossed the watershed in either englacial or supraglacial courses. Pre-existing topography probably influenced the gradual migration of meltwater streams into the Crawley col where they became united and superimposed on its floor as one large "master" channel. Since the highest point on the floor of this channel occurs between 300 and 325 feet, this must have been the base level of erosion for contemporary meltwater streams eroding channels to the north. This is supported by the fact that, with only two exceptions, all meltwater channels in the north-east Cheviots terminate at or above 300 feet; the two exceptions are only small features related to a later phase of fluvioglacial drainage.

Meltwater Channels Aligned Downslope

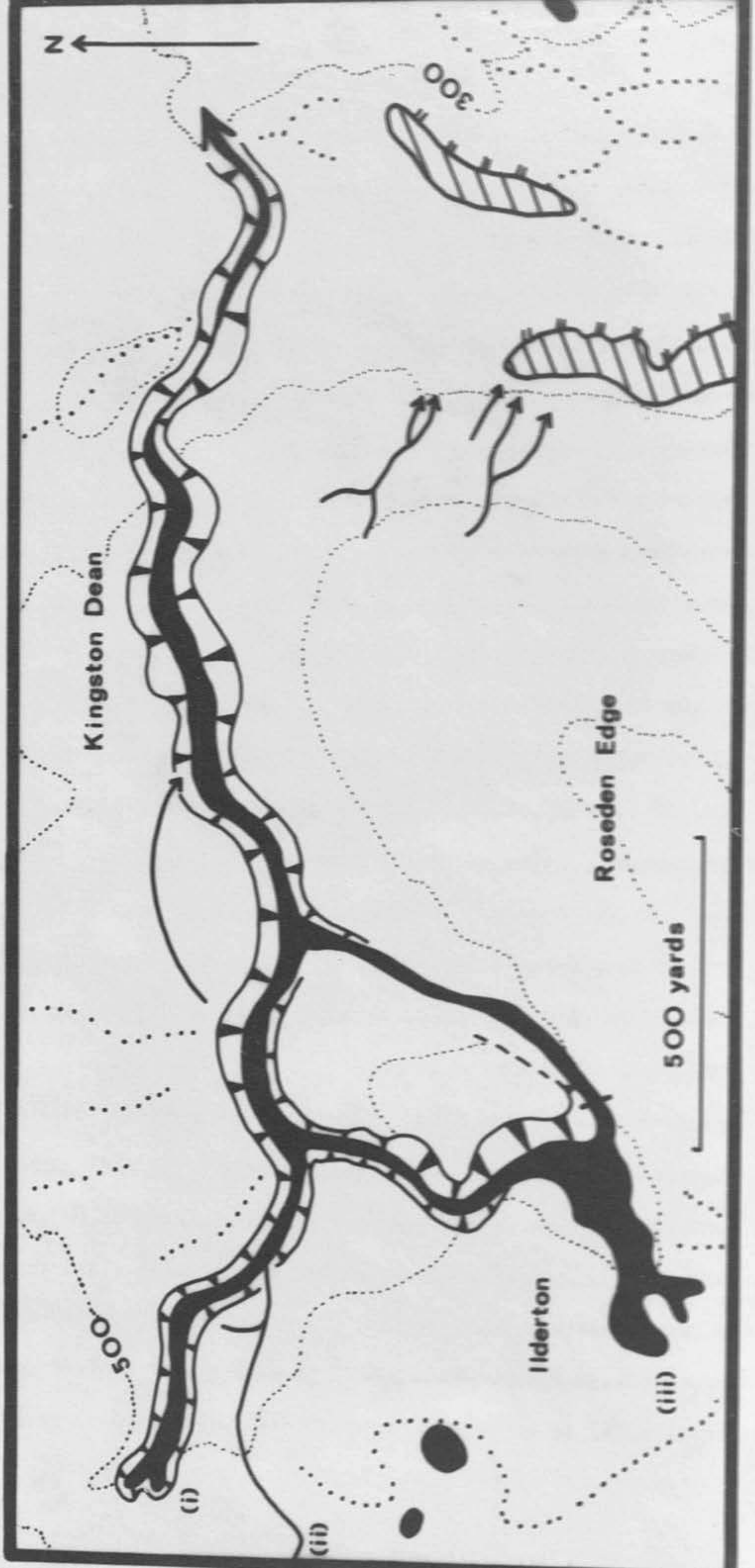
The majority of meltwater channels in the north-east Cheviots so far described are orientated approximately at right-angles to the alignment of pre-existing valleys, and as such, are discordant to the topography. There also exists, however, another set of meltwater features orientated more concordantly to topography, and is comprised of nine channels/channel systems (12, 21, 21a, 22, 23, 24, 25, 26, 27/28, Maps 5, 6 and 8). For the most part they lie below 600 feet, although two examples appear to have some connection with higher lines of meltwater drainage. Unlike the previous group these systems almost invariably follow the easiest downhill route in relation to the general topographic slope. The majority are cut partly in bedrock and partly in drift - especially in fluvioglacial sands and gravels - and whereas small streams presently occupy the main sections of systems 23 and 24, tributary branches and the other systems remain quite streamless. Derbyshire was aware that these channels constitute a morphologically separate group of features and observed "Except for their length-breadth-depth ratios, these channels closely resemble normal stream systems. Like normal streams, tributaries join the main stream in complete accordance, and their lower courses merge imperceptibly with the present valley bottoms." Although generally correct, this observation does not apply to channels 12, 21a and 22, all of which hang over 15 feet above the present stream valleys to which they are tributary. Workers in this area previous to Derbyshire divided these systems somewhat arbitrarily, partly as "contour channels", and, in their lower courses, as "transverse channels". Derbyshire criticised the "contour channel" category with the observation, "More often than not, the gradients of these upper sections are too high to be regarded as having originated strictly marginal to a dying ice-mass, where very low gradients are to be expected.", and, having observed "The presence of till in these features ...", he suggested that "these channels were formed subglacially before the final deposition of till". In view of their marked contrast with the systems elsewhere in the Cheviots, and since no detailed description of these systems has previously been published it is perhaps relevant to provide one at this point before an interpretation is suggested for them.

No. 12 is quite an impressive feature cut to a maximum depth of 30 feet through a mass of fluvioglacial sand and gravel on the northern fringe of Wooler. Its broad intake, at about 260 feet, appears to have been partially truncated by the present Humbleton Burn, but its outlet grades into the surface of a delta-shaped terrace of sands and gravels at about 180 feet on the fringe of the Ewart Basin. The alignment of this channel with the system that curves round the northern face of Humbleton Hill may be significant and it could represent a continuation of that system.

No. 21a is a small feature cut through a veneer of drift into a spur of bedrock. It is at least 15 feet deep and slopes from 250 feet to 200 feet where it fades away on the slopes of the river Till valley. While it appears to continue a line of meltwater drainage traceable from Old Middleton, the very large kettle hole/dead-ice hollow at its intake might have supplied some of the water involved in its excavation.

No. 21 is a narrow channel 12 feet deep cut into the gentle till-covered slopes beside Old Middleton at 600 feet. Although this feature fades on the side of a present stream, which crosses its path at right-angles, the same line of meltwater drainage resumes with a rock-cut intake on the opposite bank. Occupying a pre-existing depression, this shallow continuation ultimately deepens to 25 feet as it cuts through deposits of sand and gravel. It eventually loses the identity of a definite channel feature at 430 feet. From 430 feet to 400 feet, the north side of hill A (Map 6) looks water-trimmed, however, and probably represents a one-sided continuation of channel 21.

No. 22 possibly has some connection with channels 18 and 20 (Map 5) previously described. Feeders to channel 22 begin above 700 feet and cut at right-angles through a system of fluvioglacial ridges before uniting a short distance upslope from South Middleton. A small stream presently occupies this channel and has cut itself a little valley through the fluvioglacial deposits from South Middleton to channel 21. Channel 22, however, swings sharply away from this line of drainage and runs south-eastwards amongst hummocky topography where it finally diverges into two outlets round a massive mound of sand and gravel. Both outlet branches terminate at 400 feet and hang approximately 25 feet above the present Lilburn Burn. For the greater part of its course channel 22 is aligned directly downslope, but the sharp bend it takes at South Middleton causes it to follow a pre-existing depression more or less parallel to the contours of the hillside above. Finally, the double outlet section follows an oblique path across the contours.



N
300
Kingston Dean
Roseden Edge
500 yards
500
Ilderton
(i)
(ii)
(iii)

Figure 2.8

No. 23 (Figure 2.8). The main trunk of this system is fed by three tributaries; one of them bifurcates near its intake into two separate, well-defined segments that independently join the main trunk. Tributary (i) begins abruptly at 500 feet as two plunge-pool features unite; they are incised into conglomerate bedrock. Although it does not quite breach the major watershed of this area, tributary (i) follows the general north-west-south-east alignment of the higher network of meltwater channels which does breach watersheds, but it soon turns sharply eastwards down the regional slope. Tributary (ii) is simply a very minor feature feeding into (i) and merits no further discussion. Tributary (iii) is extremely interesting for it begins as a broad marshy flat with little channel form; this is at 500 feet. Indeed, the intake area bears more resemblance to a kettle hole or dead-ice hollow than to a meltwater channel. Nevertheless, before Ilderton hamlet is reached two well-defined channel forms have branched off from the intake area and diverge round a massive accumulation of sand and gravel. Branch (a) is the more impressive with its east wall cut almost 50 feet deep into the deposits. The cross-profile at this point is quite remarkable in that it illustrates that branch (a) has been incised into the sloping side of the sand and gravel mound. Branch (b) is cut into a depression between the deposit and the smooth, steeply sloping side of Roseden Edge. A partial veneer of till and several small ridges of gravel mask the conglomerate bedrock comprising Roseden Edge, but there is little doubt that this large spur of bedrock controlled the alignment of channel 23. The main trunk (Kingston Dean) is a steep-sided gorge incised through the conglomerate which is sporadically capped by thin deposits of till and fluvioglacial materials. The channel becomes almost 80 feet deep where it is incised into the side of a hill, so that it assumes an asymmetric cross-profile like tributary (iii). Towards its outlet the channel cuts through fluvioglacial ridges at right-angles to their alignment before fading away at approximately 300 feet.

No. 24 (Roddam Dean) (Map 6). The highest level of meltwater drainage associated with the deep gorge of Roddam Dean appears at 900 feet in the col between Dunmoor Hill and Reavely Hill (Map 7), but the direction of this drainage was from south-south-west to north-north-east and belongs to the system pertaining to a more southerly ice-mass which will be fully discussed in a subsequent chapter. There is, however, abundant evidence that meltwater also drained south-eastwards into the Roddam valley; it is contained in the col between Heddon Hill and Roseden Edge. Five small channels and several eskers (discussed later) indicate movement of meltwater down the lower slopes of Heddon Hill and a broad, but shallow, meltwater channel occupies the valley floor. Although these formations tend to hang by as much as 70 feet above the present floor of Roddam Dean they undoubtedly were partly responsible for its inception. Roddam Dean itself is a spectacular rock-cut canyon that gashes the southern end of Roseden Edge. It attains a maximum depth of over 100 feet and its precipitous walls expose the Roddam Dean conglomerate. The present stream flows swiftly through the ravine but is relatively inactive in any further development of it. Terminating at approximately 300 feet, channel 24 is aligned downslope in an easterly direction similar to Kingston Dean.

Nos. 25 and 26 make a peculiar system. Two rather vague, channel-like features begin amongst eskers on the slopes above the south wall of Roddam Dean. They unite to form a more definite channel that fades out in a large depression surrounded by fluvioglacial deposits (x, Maps 6 and 8). This depression is joined by another channel entering from the south-west. An upper line of meltwater drainage in the area appears to have flowed initially south-eastwards through the valley east of Fore Rigg and continued past

Roddamrig House towards the Breamish valley. At a later stage this alignment was forsaken and channel 26 (leading north-east into the depression) was cut almost 50 feet deep. This represents an abrupt change in the direction of fluvioglacial drainage in the area. Ice-contact slopes surround the depression and numerous large kettle holes are adjacent; the depression is therefore thought to be a large dead-ice hollow into which meltwater flowed and from which a deep channel evacuated the water to just below 300 feet. The important property of this somewhat discontinuous line of meltwater drainage is its easterly downslope trend, terminating at about 300 feet.

No. 28 is an impressive gorge orientated from south-west to north-east; this drainage alignment has already been attributed to a southern ice-mass, but such a relationship may not be valid for channel 28. Similar to channels 24, 25 and 26, its lower reaches cut across ridges of fluvioglacial sands and gravels. The east wall rises 80 feet above the floor at one point and is cut chiefly in bedrock, although shallow ridges and mounds of sand and gravel lie on the slopes above the channel wall. This meltwater gorge heads in a broad, marshy flat which appears to have been a point of bifurcation for meltwater flow from the north-east; the channel east of Roddamrig House (referred to above) can be traced continuously south-eastwards past the point of bifurcation into the huge channel called Brandon Dean, which outlets into the Breamish valley. On the other hand, it can also be traced continuously through an angle of 180 degrees into channel 28. In fact, this latter route appears to have been the last in use for the Brandon Dean branch hangs about 8 feet above it. Channel 28 terminates at 300 feet, Brandon Dean at 325 feet.

Derbyshire has observed till in the floors of some of these channels and on this evidence suggested that they are "subglacial channel systems". More certain evidence that may be quoted in support of this hypothesis is that they cross the contours at right-angles and obliquely - topographical relationships that can best be accounted for by subglacial drainage of meltwater, as with subglacial chutes. In addition to volumes of water being furnished by decaying ice lying against the foothill zone of the massif, water from ice-free hillsides and valleys must have been important in augmenting drainage into the ice, particularly when it had downwasted to the lower elevations suggested by the location of these channels.

The above channels differ considerably in size, form and complexity, but they are all roughly parallel with each other and demonstrate the flow of meltwater eastwards and north-eastwards. This alignment is conspicuously at right-angles to the dominant direction of meltwater drainage illustrated by the higher network of channels previously described. Furthermore, they all trend more or less with the slope of normal drainage, a fact which suggests relaxation of the ice control so necessary during the formation of the previous channel network. Since nearly all of the channels terminate at or above 300 feet, it is possible that their lowest level of erosion was controlled by the flow of water through Crawley Dean. It is therefore suggested that an englacial water-table was in existence at that time, at a height of approximately 300 feet, and that the downslope channels were unable to erode below that level. The concept of this englacial water-table will be more fully elaborated in Chapter 3.

The majority of meltwater channels in the north-east Cheviots have been described in two main categories according to their dominant alignment. A small number of channels occur outwith these categories, however, and require separate consideration. Located much more centrally in the massif, these features are aligned differently, so that they do not conform to the general patterns of drainage illustrated by the two former channel networks.

Two such channels (33 and 34, Map 5) are associated with the Lambden Burn valley. Three smoothly rounded, but steep-sided hills - Preston Hill (1,724 feet), Broadhope Hill (1,694 feet) and Scald Hill (1,797 feet) - form a semi-circle of high ground partially enclosing the head of this valley, and the two channels occur in prominent cols separating Broadhope Hill from the others.

The col containing channel 33 lies between Preston Hill and Broadhope Hill; it is over 250 feet deep and 150 yards wide and leads north-north-eastwards from the Lambden valley to that of the Broadstruthers Burn. The peat- and heather-covered floor of this col, where the latter stream rises in a small valley, lies at 1,375 feet. Channel 33 begins at 1,325 feet, on the upper slopes of the Lambden valley south of the col crest and runs uphill into the col as a two-sided feature. Rising steeply at first, the channel gradually widens and deepens as its uphill gradient slackens near the crest in its floor profile. Beyond this point the feature slopes gently down the Broadstruthers valley until it fades away as a distinct channel. Cut almost entirely through bedrock it is incised into the lower slopes of Broadhope Hill and is therefore not located centrally in the col. Because of this the channel's cross-profile is always asymmetric; the eastern wall frequently reaches 30 feet in height while 10 feet is about the maximum development on the opposite side. Peat has grown thickly in the channel floor but is unlikely to be wholly responsible for the crest in its longitudinal profile which lies at least 40 feet above intake level. Channel form is less well-defined in the lower reaches where a bench-like section merges into the gorge presently followed by the Broadstruthers Burn. All definite signs of fluvioglacial erosion end at 1,025 feet but melt-water drainage almost certainly continued down the line of the broad, pre-existing valley.

Photograph 2.j The uphill intake section of the New Burn channel (34, Map 5).

Channel 34 lies in the floor of a much wider and shallower col between Broadhope Hill and Scald Hill at 1,500 feet. The channel begins at between 1,450 and 1,475 feet on the Lambden side of the col, west of its crest, and leads into it as a well-defined trench 40 yards wide and 15 feet deep. The floor rises gently for about 150 yards from intake level to a crest 20 feet higher (Photograph 2.j), beyond which the channel deepens eastwards to a maximum of 40 feet; it is cut entirely in bedrock which only sporadically shows through the mantle of vegetation. Similar to the previous channel, the floor is so thickly buried by peat that it looks remarkably flat. Approximately 9 feet was observed in a hole at one point and while considerable depths of peat have been recorded on the higher summits of the massif, it is unlikely that sufficient depth prevails in this channel to account for its up/down longitudinal floor profile. Like the majority of col channels in the Cheviots this feature becomes incised into the head of a conspicuous pre-existing valley. The cut obviously made by meltwater terminates at 1,325 feet; below this point a waterfall leads into a narrow gorge quite different in its morphology from the meltwater channel at its head. The interlocking spurs and v-shaped cross-profile of this gorge were probably cut by the New Burn which presently occupies it, for the stream appears adequate to have accomplished such erosion in the deeply rotted and fragmented bedrock exposed along its valley. Several small, dry channels occur in the lower section of the valley, hanging on its upper slopes; they suggest that meltwaters did continue down this path, but how far they may have been responsible for excavating the main gorge is difficult to assess.

Channels 33 and 34 each contain an up/down longitudinal floor profile, a phenomenon which so far has been satisfactorily explained only by subglacial flow of meltwater under hydrostatic pressure. Accordingly, both channels are considered subglacial in origin, arriving in their respective positions through the superimposition of englacial streams. Since the direction of meltwater drainage indicated by these channels is outwards from the Lambden valley, it can be deduced that the ice surface there stood higher than to the north and east beyond the watershed. According to Carruthers et al., the massif is unlikely to have acted as "an active centre of dispersion" (of ice), and Common noted that "Cheviot did NOT, indeed could NOT, supply the amount of névé to form a local ice centre". It will be shown in a later chapter, however, that a prominent ice cap almost certainly formed on The Cheviot. The glacier ice that filled the Lambden valley was therefore derived locally.

At 1,500 feet, channel 34 is the highest meltwater formation in the east Cheviot area and probably functioned at an earlier period than channel 33, which, at 1,375 feet, is the next highest. This must have been at a time when the ice surface lay at a minimum altitude of 1,600 to 1,700 feet in the vicinity of that particular col, for channel 34 was eroded subglacially at 1,500 feet under considerable hydrostatic pressure. The general west-east slope of this ice surface is reflected by the location and alignment of channel 34. Channel 33 is orientated almost at right-angles to the latter and lies 125 feet lower in altitude. The two channels probably did not function contemporaneously for the following reason. Channel 34 is cut at 1,500 feet and 33 can be traced down to 1,025 feet; if the two features operated simultaneously, a depth of meltwater penetration 475 feet below the ice surface is implied, and since this is considerably in excess of the deduced depths reached by other subglacial streams in the Cheviots and elsewhere in Britain, this situation is considered unlikely. Observations from present glaciers do not support such a depth of subglacial penetration by meltwater. The orientation of these channels at right-angles to each other is also rather improbable for contemporaneous drainage. The direction of meltwater flow indicated by channel 33 cannot be readily reconciled with the deduced slope of the ice surface above the Lambden valley unless account is taken of downwastage and some measure of control by underlying topography. As the ice surface progressively lowered from the deduced minimum of 1,600 to 1,700 feet, the tops of hill masses girdling the Lambden valley emerged as large nunataks until the col between Broadhope Hill and Scald Hill became ice-free and ceased to function as an escape route for meltwater from ice in the Lambden valley. At that moment, when the ice surface stood at approximately 1,500 feet, the col between Broadhope Hill and Preston Hill would still have been covered by at least 125 feet of ice. It is therefore more acceptable that following the abandonment of channel 34, englacial drainage reaching the head of the Lambden valley was obliged to take the next most convenient route available, which was the col in which channel 33 is located. Furthermore, the tongue of ice occupying the col would presumably have had a pronounced slope towards the north-north-east as it "overflowed" from the Lambden valley.

Perhaps the main significance of channels 33 and 34 is that they are the highest meltwater formations in the east Cheviot area and afford some indication of the extent to which glacier ice built up in the central part of the massif. They also indicate directions of meltwater drainage quite different from those shown by the two major channel networks previously described and presumably belong to a much earlier stage of deglaciation, for there is no evidence suggesting that they were formed by an earlier glaciation.

Except for the two small chute-like channels of little consequence which slope down the northern side of the Threestone Burn valley, the remaining channels outwith the two main groups occur between Langleeford and Langlee in the Harthope valley (Map 5). Numerous dry channels furrow the steep valley sides in this area. Except for channel 35 they all run downslope precipitously, sometimes at right-angles across the contours and sometimes obliquely to the contours. Tributaries, connecting limbs and distributaries are common characteristics displayed by these channels, the majority of which are cut in bedrock. Depths of over 15 feet are unusual. They are all quite streamless (except for part of Easter Dean) and no debris cones spread from their termini; furthermore, they mostly end before the base of the slope into which they are incised. The preceding characteristics suggest that they are all subglacial chutes. Exactly why such a concentration of these features should occur at this point in the valley is not readily understood, but they certainly prove that there was abundant meltwater and highly crevassed ice. Accordingly, the following factors may be extremely relevant:

(a) The Harthope valley is wider along this section, perhaps allowing more scope for the establishment of subglacial chutes beneath remnant ice in this part of the valley.

(b) Two major tributary valleys, those of the New and Hawsen Burns, join immediately above this concentration of channels. Both valleys would have led considerable volumes of water onto and into the decaying ice occupying the Harthope valley, and this water appears to have escaped subglacially at many points beneath the margins of that ice mass.

The other major meltwater feature in this area is the large meltwater channel that is aligned almost parallel to the present stream, from which it is separated by a narrow ridge of ground called The Shank (S, Map 5). The wide, marshy depression leading into this feature lies 25 feet above the Harthope Burn, and although partly modified by an alluvial fan at the mouth of Wester Dean, the Shank channel becomes a sharply incised gorge 25 feet deep.

Its lower section curves round towards the Harthope Burn above which it hangs by 20 feet with a double outlet. The western limb of this outlet was abandoned before the other ceased to function and can be observed hanging 15 feet above the latter. The Shank channel is cut through a veneer of gravelly drift and into underlying bedrock. At the point where the double outlet curves round towards Harthope Burn, a remarkably flat-topped spread of drift continues along the valley side for approximately 250 yards in line with the main trend of the channel. No good exposure is available but sub-rounded pebbles and a gritty matrix lie scattered on the surface in places. This feature is possibly linked with the Shank channel and probably represents a depositional phase prior to its truncation as the meltwater stream cut its final outlet. While it may be argued that the channel and terrace resemble features frequently interpreted as representing marginal drainage, the fact that some subglacial chutes appear tributary to them suggests a subglacial origin. Their conspicuous alignment with the New Burn valley, which contains abundant evidence of former meltwater drainage down it, points to a continuous line of drainage - that section across the Harthope valley being entirely englacial.

<u>The System as a Whole</u>

Apart from the independent courses assumed by channels in the last group, the majority of meltwater channels in the north-east Cheviots are located within a peripheral zone, for the most part below 1,100 feet. Approximately two miles broad, this channelled tract of country swings east and south-east roughly following the margins of the Old Red volcanic outcrop, and extends in a wide curve from Yeavering Bell to Crawley. It represents a remarkable concentration of meltwater channels that is apparently confined below a pronounced upper limit. The highest channel on each successive spur from north to south declines progressively from 1,125 feet (channel 1) to 425 feet (channel 30). Although the channels, with few exceptions, have been interpreted as subglacial formations and as such cannot accurately represent former ice margins, this remarkable upper limit to their concentration may be meaningful. It may well afford an approximate minimum level of inundation by the Tweed glacier round the north-east Cheviots. The ice perhaps extended to higher levels, but there is no evidence to record or prove this. The declination in channel limits is 700 feet over a distance of roughly 12 miles, and converted to a gradient, it may approximately represent a former ice surface which sloped south-eastwards at 1:90.

Channels 33 and 34 of the last group were formed by meltwater associated with an ice surface which sloped north-east and east out of the Lambden valley area where it built up to a level of at least 1,600 to 1,700 feet; this body of ice may have been confluent with that which swept round the north-east periphery of the massif. In view of this it is difficult to assess whether fluvioglacial formations in the Harthope valley near Langleeford are linked with ice from the Lambden valley or with ice from the north-east, for the two are likely to have come in contact about here. However, the slight down-valley orientation of several subglacial chutes, and certainly that of The Shank channel, suggest that glacier ice lying in this part of the valley sloped outwards from the massif. If extraneous ice had penetrated the Langlee area from the north-east, a slope inwards to the massif of its surface and associated fluvioglacial formations is to be expected.

Conclusion

Meltwater channels in the north-east Cheviots may be grouped into three major categories according to their orientation with respect to the massif.

The first group considered consists of those which cut through pre-existing cols and valley heads. The majority of these trend at right-angles to valleys radiating from the massif and are aligned roughly parallel to its curving margin. Evidence indicating a subglacial origin for these channels has been fully discussed and a hypothesis of superimposition from former en-glacial courses suggested as the means by which many of the meltwater streams became subglacial. The number and complexity of the channels and channel systems comprising this group demonstrate how vast networks of anastomosing meltwater streams must have been contained within the marginal zone of the Tweed glacier, and suggest a phase or phases of rapid deglaciation during which immense volumes of meltwater were liberated. Each channel cut on the ground by such meltwater rivers must represent only a small fragment of a stream course; it can therefore be deduced that where any channel terminates, the meltwater stream responsible for it resumed an englacial course or else a sub-glacial course marked by deposition or negative activity rather than by erosion. Indeed, it will be suggested in a subsequent chapter that the extensive deposits of fluvioglacial sands and gravels clothing lower slopes of the volcanic massif and masking considerable expanses of the low-lying Cementstone basins, probably represent depositional counterparts of many channels described in the massif. Fluvioglacial waters responsible for this large group of channels consistently flowed south and south-east into the area of Hedgeley Basin. While the englacial water-table remained at a level above 425 feet these waters appear to have escaped englacially through and across the rim of high ground encircling the basin; this was probably achieved mainly in the vicinity of prominent cols through the watershed. As downwastage progressed, the water-table dropped below the 425-foot level and the Crawley Dean channel began to function; it remained in operation as a subglacial channel until it was cut to a level of 300-325 feet. All meltwater channels in this first group terminate at or above this level and were presumably controlled by it. Since the uppermost channel on each successive spur from north to south declines progressively in height, it is suggested that this upper limit of fluvioglacial erosion is meaningful in estimating an approximate minimum level of glacier inundation round the north-east Cheviots.

In the second group described, meltwater channels are orientated approximately at right-angles to those in the first and appear to trend mainly with topography, assuming the easiest downslope course available. The strong control exerted by ice in determining the orientation and location of the previous channels appears to have been relaxed, so that meltwater and water from ice-free areas penetrated directly downslope under the margins of heavily decayed ice occupying lowland basins, situated between the volcanic foothills and sandstone escarpments to the east. Since the majority of these channels lie below 600 feet, it may be suggested that dead-ice only about 600 feet thick occupied the basins. Because they terminate at or above 300 feet, it seems that Crawley Dean probably controlled the level of an englacial water-table in the Hedgeley Basin, below which the channels could not be eroded. All the channels in this group cut through ridges of fluvioglacial sands and gravels which are orientated north-west-south-east. They therefore post-date a phase of considerable deposition related to the same alignment of fluvioglacial drainage observed in the channels of the first group, but nevertheless were formed subglacially.

Perhaps the most important information yielded by channels in the third group is that a mass of glacier ice accumulated to a height of at least 1,600 to 1,700 feet in the Lambden valley area, and that its surface sloped northwards and eastwards as though radiating from the heart of the massif.

Fluvioglacial formations about Langlee in the Harthope valley are also believed to have been associated with this body of ice and prove it to have extended at least to 1,300 feet in that area. All channels in this group have been interpreted as subglacial in origin.

Whereas both Common and Derbyshire classified meltwater channels according to morphology and genesis, it seems much more practical and meaningful to group such features according to their dominant alignment. In this way meltwater channels may contribute significantly to an interpretation of the glaciation and deglaciation of the north-east Cheviots.

CHAPTER 3.

FLUVIOGLACIAL DEPOSITS IN THE NORTH-EAST CHEVIOTS

Introduction

Between Wooler in the north and Percy's Cross in the south-east (Maps 5 and 8), the peripheral foothill zone of the Cheviot massif slopes generally eastwards down to low-lying basins presently drained by two principal streams, the Wooler Water and the river Breamish/Till. Immediately south of Wooler, volcanic rocks present a steep edge overlooking the Wooler Water; south of this stream, the topography is more broken and varied, and hillslopes of more gentle gradient descend to the foothill basins. Steep slopes are still present in places, however, such as those of Heddon Hill, Roseden Edge and Brandon Hill, but they are much more infrequent than elsewhere in the north-east Cheviots. The underlying bedrock is generally composed of igneous materials and the small outcrop of Roddam Dean conglomerate is quite indistinguishable in its topographic expression from the Old Red volcanics surrounding it and which are unconformably overlain by it. For the most part, the Wooler, Lilburn and Hedgeley Basins, which have been referred to as the sub-Cheviot depression, are underlain by rocks belonging to the Cementstone group of the Lower Carboniferous and lie below 400 feet. Clearly bounded by the volcanic and conglomerate rocks rising westwards, their eastern margin is equally distinct and much more impressive. The sharply-defined escarpment of Weetwood Moor rises precipitously to tower 300 feet above the Wooler Basin. Similarly, an equally abrupt scarp edge, although considerably more irregular in plan, girdles the Lilburn and Hedgeley Basins. This escarpment is surmounted by prominences on Hepburn, Bewick and Beanley Moors where elevations of over 700 feet are common; an isolated mass north of Hepburn Moor forms the highest

Photograph 3.a Rolling topography caused by the tightly packed eskers south of Lilburn Burn. The Fell Sandstone escarpment in the background.

point at 1,025 feet. These conspicuous escarpments are underlain by sandstones of the Fell Sandstone group of the Lower Carboniferous.

From the environs of Wooler, south-east to the river Breamish, the lower slopes of the Cheviot massif are extensively covered by fluvioglacial deposits. Within this roughly triangular-shaped area exposures of bedrock are confined to the steep walls of meltwater channels, and apart from strips of postglacial alluvium along the Lilburn and Roddam Burns, only the hill-masses at X, Y and Z, Map 6, remain as uncovered enclaves in the mass of sands and gravels. The paucity and poor quality of sections and exposures create difficulties in determining whether superficial drift on these enclaves is broken bedrock or till; it is probably a combination of both. North of Coldgate Water the sands and gravels are confined mainly below 400 feet, whereas from Coldgate Water south to the Breamish, they are abundant up to 800 feet; an isolated ridge of gravel at 950 feet represents the highest observed deposit of fluvioglacial material in the north-east Cheviots. This considerable expanse of sand and gravel is arranged, for the most part, in complex networks of ridges. Individual forms vary from broad, low swells only a few feet high, to sharp-crested ridges rising over 100 feet above adjacent depressions. Quite frequently, the ridges broaden into massive mounds, occasionally with flat tops; they twist sinuously and continuously divide and unite in a most irregular manner; tributary and distributary segments are equally common characteristics. In some places the features are so concentrated and confused that they create a rolling landscape (Photograph 3.a) of tightly-packed knolls, ridges and depressions, each merging and uniting with the next in apparent disorder. Elsewhere, order clearly prevails, and neatly-defined ridges run parallel with each other and are occasionally linked by connecting branches. Although this linear form of the fluvioglacial deposits is present in most places these deposits also form considerable expanses of terrace-like topography, particularly at Calder, Brandon and Wooperton. Enclosed hollows varying in depth from 3 feet to 25 feet commonly occur between and on the ridges and frequently pit the terraces. Their floors are quite often covered with water or marshy vegetation but may also be completely dry. Other depressions within the deposits are not quite enclosed; these form large indentations in terrace margins and in the flanks of ridges and mounds. The enclosed depressions are interpreted as kettle holes and the open depressions are termed dead-ice hollows.

Fluvioglacial deposits on hillsides above about 400 feet tend to be relatively thin and predominantly coarse-grained, and, as such, contrast with those in the low-lying basins where they are thick and usually finer grained. This is very much a generalisation, for many exceptions occur in which lenses of fine sand are present in the higher deposits while layers of large cobbles and even boulder beds may be observed in those at lower levels. Clear sections revealing the internal composition of these deposits are few in number and are chiefly confined to the banks of two streams. The Wooler Water has extensively undercut sand and gravel ridges on its left bank opposite Haugh Head, forming precipitous scars up to 80 feet high at some points. Clean sections are present in only three places, however, for slumping and vegetation have obscured many faces that were exposed previous to the artificial entrainment of the stream. Other sections presently available for inspection occur at two places on the right bank of the Roddam Burn just below 300 feet. The fluvioglacial nature of these deposits has not been deduced solely from morphological evidence, however, since limited, but valuable, information may be gleaned from the upper few feet of slumped bankings along major and minor water courses, from minor exposures at tree roots, in rabbit burrows, in sites where sheep have excavated shelter hollows, and in other minor irregularities on slopes. Finally, the nature of materials exposed on the surface of a freshly ploughed field may sometimes indicate whether or not fluvioglacial deposits underlie the soil.

Previous Work

Concerning these sands and gravels, very little was published previous to an admirable, descriptive account by Bullerwell in 1910, although Tate (1866b), Hardy (1872), and Lebour (1886) had briefly referred to them. In spite of ample documentation on the ridge-like nature of these deposits, Bullerwell tended to ignore the significance of their morphology and ascribed their formation to the rapid deposition of rivers "fed from the glaciers and glacial lakes at higher altitudes". He implied that they were deposited in a lake or series of lakes whose margin(s) lay between the 300 and 400 foot contour lines and that "The coarse gravels in the sections near Wooler and Roddam were the littoral deposits, while the finer material and sands of Hedgeley were carried into the quieter and deeper portions of the lake." He did not provide an explanatory map with this account. Smythe (1912) alluded briefly to this belt of deposits, but only within the general framework of glacial sands and gravels in Northumberland and although Dinham (1927) provided a most detailed and exhaustive account of similar deposits in the Belford district, his colleagues were much less enterprising in their appraisal of such formations in the north-east Cheviots. Burnett (1930) referred to "well-defined areas of 'kettle moraine' in the Breamish valley" and described these deposits as "accumulations of dirty, ill-assorted material, gathered on the surface of 'dead ice'", that "were left behind after the ice had melted away". He also observed that "they are characteristically pitted with shallow depressions, or 'kettles', marking the site of detached and isolated ice-remnants, the last to disappear". His explanation of these "Hedgeley Gravels" involved a lobe of dead ice from the Tweed glacier that slowly dwindled away in the confined space of the Hedgeley Basin, where "spreads of gravel or sandy dirt, originally washed on to it from the surrounding hill-sides, are a witness to its former presence". A contribution by Dinham in the same memoir briefly mentions the Bradford-Charlton gravel "string" he had discussed in detail three years earlier, and he tentatively concluded that "the Bradford-Charlton 'string' is comparable to the eskers described by Leverett and Stone". He left the explanation of their deposition an open question, however, "except that they seem to be the work of glacial waters of some kind". In the following paragraph Carruthers provided an alternative hypothesis for the entire belt of gravels, including the Bradford-Charlton "string" and those in the Breamish valley. In this, he said that the deposits essentially represent a moraine formed along the edges of a southern lobe of the Tweed glacier, at a check in its retreat round the foot of the Cheviot Hills. Although reports from Alaskan glaciers and publications on formerly ice-covered areas of the United States described fluvioglacial formations similar to those observed in north Northumberland and explained them as eskers, the Geological Survey officers were clearly hesitant in drawing a full analogy and applying such concepts to Britain. Dinham was almost convinced that the Bradford Kame is an esker but the less clearly defined morphological expression of similar deposits in the Breamish valley evidently puzzled Burnett and Carruthers to such an extent that they were loosely grouped as a zone of "kettle moraine". Much more recently, Derbyshire (1961) has reconsidered the significance of these sands and gravels with respect to the deglaciation of the north-east Cheviots, but apart from briefly summarising Carruthers' description, his only contribution of note is the rather tentative suggestion "that they are subglacial deposits, possibly of esker affinity".

Fluvioglacial deposits in the north-east Cheviots have, therefore, received only generalised description and uncertain interpretation in previous literature on this area. Although the one-inch drift maps of the Geological Survey provide a reasonably accurate delimitation of their extent, a detailed map of their morphological expression is notably lacking. In view of this situation, it was decided to map all the landforms of fluvioglacial deposition in the north-east Cheviots, using a base map on the scale of six inches to one mile, subsequent to an exhaustive stereoscopic study of such forms on vertical air photographs. It was believed that this analysis might unravel the apparent confusion of depositional forms prevalent in some areas and aid a more comprehensive interpretation of their genesis.

Perhaps the most remarkable and far-reaching result emerging from detailed morphological field mapping of these deposits is the clearly-defined pattern of ridges and terraces into which an apparent chaos of forms has been resolved. The nature of the constituent materials may be clearly observed at the few good sections available in which the stones are predominantly water-worn and the distinct sorting and bedding (described more fully later) indicate that the selective action of running water has played a large part during the deposition of these sands and gravels. Various evidence strongly suggests that this water was, in fact, glacial meltwater, and includes the following points:

(i) There is an abrupt range in the calibre of materials contained within adjacent beds; for example, lenses of extremely fine rock flour are interbedded with coarse gravels and boulders up to 3 feet in girth in the section opposite Haugh Head. This demonstrates that stream flow was very irregular and prone to considerable variations in volume -

characteristics that are normally associated with rivers flowing in a glacial environment.

(ii) The beds are arranged in undulations and interrupted by faults and slumps. These characteristics generally suggest deposition in or around glacier ice, so that disturbances in the bedding are produced when the ice melts out.

(iii) Up to 3 feet of glacial till overlies the sands and gravels in some places.

The side slopes of the ridges and mounds are usually steep, although some of the smaller features are simply gentle swells. As a rule, deposits of gravel tend to maintain steeper slopes than those of sand, since the angle of rest for the constituent particles is greater. Such a rule is difficult to apply generally in the north-east Cheviots because of considerable modification to original slopes by various forms of cultivation and farming. The most sharply-defined group of ridges, although by no means the most impressive, occurs on Ilderton Moor, the absence of cultivation on this rough, hill-grazing ground allowing the preservation of slopes in their more original form. Despite centuries of human interference, however, the side slopes of the majority of these fluvioglacial formations can be readily interpreted as ice-contact slopes, and, together with kettle holes, dead-ice hollows and the internal properties of the materials, point to the widespread deposition of sands and gravels in a glacial environment characterised by decaying ice. Elements in the morphology of these ridges such as their sinuousity, their continual dividing and uniting, and tributary and distributary branches, strongly suggest an origin associated with closely spaced networks of braided streams. It can therefore be concluded that the ridges and mounds of sand and gravel in the north-east Cheviots were deposited in an ice-contact environment by systems of braided meltwater streams. They are accordingly interpreted as eskers and not as moraines.

In some parts of Britain, isolated eskers stand out as distinct, sharp-crested ridges that frequently display considerable sinuousities. Examples of these are, the Bradford Kames, the Kames near Greenlaw, the Thankerton Kames, the Strathnairn eskers and an esker on the south-western outskirts of Inverness, perhaps the largest of such features in Britain, for it is almost 200 feet high. Although some of these examples were initially interpreted as moraines, they are now generally accepted as true eskers, but more complex formations, such as those near Carstairs, have only recently been interpreted as an esker system and controversy still exists on the true genesis of these deposits. This perhaps reflects the hesitancy of some workers in Britain to accept that while complex, often braided, networks of meltwater channels may be cut on hillsides, such systems of meltwater streams are able to deposit sands and gravels in a similar pattern. Publications by Price (1964, 1966) admirably illustrate complex esker systems recently formed in Alaska. The majority of fluvioglacial ridges in the Cairngorm mountains of Scotland, formerly interpreted as the moraines of valley glaciers, have been reconsidered by Sugden (1965) who mapped and described them as eskers. There is therefore an increasing awareness in some of the literature on de-glaciation in Britain that anastomosing systems of sand and gravel ridges in some places are indeed esker complexes and not moraines. In the north-east Cheviots, the former presence of great networks of interlinked meltwater rivers has already been demonstrated with reference to their abandoned channels. It is therefore to be expected that the deposits of such drainage systems would also assume, in places, similarly complicated patterns. Consequently, ridges and aligned mounds of sand and gravel in this area are interpreted as eskers that were deposited by meltwater streams comprising these drainage systems.

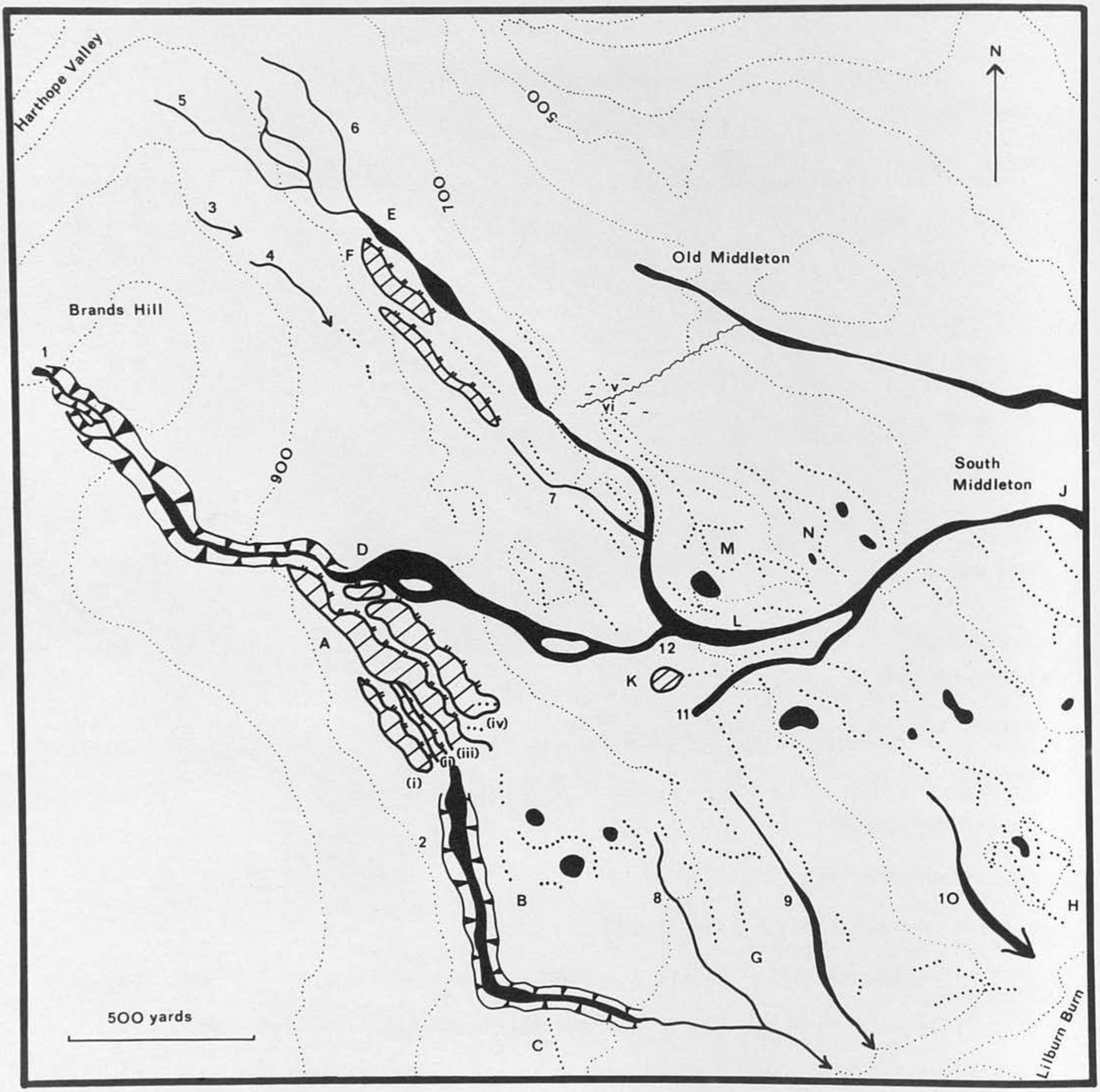
Above 300 feet, the majority of eskers and fluvioglacial terraces are aligned predominantly in directions that vary between south and south-east, reflecting depositional stages of the vast systems of meltwater streams directed into this alignment by glacier ice. Small numbers of eskers in places diverge from this orientation and illustrate that drainage was able to flow more directly downslope from time to time, perhaps correlating with some channel systems similarly aligned, and these events may indicate that progressive down-wastage of the ice mass continually allowed downslope drainage, directed by gravity, to replace that directed by the slope of the ice surface. How far down-hill such drainage was able to proceed is frequently difficult to determine, and it seems probable that much of it became redirected towards the south-east at lower levels, for no continuous system can be traced to the floors of the foothill basins.

Below 300 feet, however, the eskers and fluvioglacial terraces are conspicuously aligned in directions that vary between east, north-east and north, and, as such, illustrate a phase of meltwater drainage quite distinct from that indicated by the majority of deposits and channels above 300 feet. Since they consistently trend towards the Till valley and slope into it in a downstream direction, these formations represent the ultimate escape of fluvioglacial drainage from the north-east Cheviots.

This natural division of the deposits into two separate groups provides a convenient basis for a more detailed description and interpretation of them.

The Higher Group of Fluvioglacial Deposits

The higher group of deposits is by far the most extensive and may be



Harthope Valley
500
700
900
N
Old Middleton
South Middleton
Brands Hill
Lilburn Burn
500 yards
1
2
3
4
5
6
7
8
9
10
11
12
A
B
C
D
E
F
G
H
J
K
L
M
N
(i)
(ii)
(iii)
(iv)
v
vi

Figure 3.1

traced from Wooler to Percy's Cross (Maps 5 and 8). On hillsides above 400 feet, the eskers and terraces tend to occur as rather discontinuous systems, whereas between 400 and 300 feet they extend from Wooler to Percy's Cross with almost unbroken continuity, and are interrupted only by the floodplains of the Wooler Water, Lilburn Burn and Roddam Burn. The former connections between the truncated ends of ridges on either side of these floodplains can readily be appreciated from Map 5. The main reason for the discontinuity of systems above 400 feet is the presence, in places, of slopes that were too steep to accommodate eskers or to allow their subsequent preservation following deposition. For example, the steep and moderately steep slopes about Roseden Edge and Heddon Hill are relatively free from fluvioglacial deposits and tend to support that conclusion. The general absence of steep slopes below 400 feet allowed the formation of more complete esker systems. In the following discussion, the eskers are considered according to their altitude, beginning with the uppermost formations, for it is assumed that these were probably deposited earlier than those at lower levels. This theme also provides some measure of convenience for description and clarity of presentation.

A. The Upper Systems.

(1) The South Middleton System. On the slopes above South Middleton, a distinct series of eskers is aligned in a north-west to south-east direction, in which the forms are broadly parallel with one another and with the contours of the hillside on which they have been deposited (Figure 3.1). They occur in association with kame terraces and meltwater channels that are similarly orientated. The eskers in this system vary in form from quite small ridges merely 10 feet high, to more massive features rising as broad ridges over 60 feet above adjacent depressions. Although the majority are straight, minor

Photograph 3.b The ablation till that overlies fluvioglacial deposits near Old Middleton.

sinuousities occur in some, for example, at M and N, Figure 3.1. Downslope drainage of water at a later date has fragmented several ridges, but their original form is readily apparent in the field and from Figure 3.1. For the most part, the eskers are separated laterally by long marshy depressions and/or shallow meltwater channels. Occasionally, they divide and unite in a more confused manner, and sometimes they merge with vague, irregular spreads of similar deposits. Kettle holes are not numerous, but exist as well-defined hollows in places. Superficial exposures on several ridges reveal that the upper layers are predominantly sandy and that a thin veneer of water-worn gravel, in which the stones are generally from $\frac{1}{4}$ inch to 2 inches in size, is occasionally present near the surface; but such exposures are too infrequent to generalise on this point. The most lucrative information on the nature of drift deposits in this area is yielded by numerous small, occasionally indistinct sections along banks undercut by the small stream that runs down towards Old Middleton. The majority of these, however, are in that part of the hillside on which eskers and other landforms of fluvioglacial deposition are absent, and the exposed material is red, stoney till. Two sections occur where the stream has cut at right-angles through the eskers and the following are revealed. At section v, (Figure 3.1), 6 feet of deposit is exposed, and consists of bedded sands and gravels that vary in nature from layers of extremely fine powder to cobbles, up to $1\frac{1}{2}$ to 2 feet in girth. Slumped and cross-bedding are common characteristics. At section vi (Figure 3.1), $1\frac{1}{2}$ to 2 feet of reddish till (Photograph 3.b) rests on top of 8 to 10 feet of fine sand; this, in turn, overlies red, tenacious till. This evidence suggests that the eskers were deposited on top of till, the full depth of which is unknown, although it is at least 30 feet in one place. Isolated boulders, seemingly unrelated to the underlying sand and fine gravel, infrequently rest on the slopes and crests of eskers, and, together with the thin layer of till observed in section vi, suggest that small quantities of ablation moraine are present in places. Thin layers of ablation till are unlikely to survive intact on eskers since solifluction and other slope processes, immediately following deglaciation, would tend to be active on their steep, ice-contact slopes. While larger blocks might remain as lag deposits, the finer constituents would, for the most part, be removed, and it is therefore only a tentative suggestion that the above materials may be the remnants of a former cover that was more continuous.

Situated between 775 and 875 feet, the upper members of the South Middleton esker system are clearly associated with four terraces to the north-west, and with the meltwater channel issuing from the col at Brands Hill. The latter features were briefly referred to in the previous chapter, and it was suggested that a continuous line of subglacial drainage is represented by channel 1, terraces (i) and (ii), and channel 2 (Figure 3.1).

Terrace (i) begins 350 yards beyond channel 1 and terminates immediately before the intake to channel 2. Although the terrace surface lies several feet above the floor of that channel, it is below the level of the latter's sides. Any further extension of terrace (i) towards channel 1, "upstream" from it, has presumably been removed by subsequent meltwater flow at a lower level along the same general line of drainage. Indeed, the frontal margin of terrace (i) looks water-trimmed, rather than an original ice-contact edge.

Terrace (ii) is a narrow feature of similar extent to the latter, and, lying only a few feet lower, it leads accordantly to the intake of channel 2. It is therefore apparent that fluvioglacial erosion in the Brands Hill col was contemporaneous with a phase of deposition on the gently sloping hillsides beyond, for as stated in the previous chapter, these terraces are depositional features rather than the products of erosion. Presumably, the meltwater river issuing from channel 1 would have been heavily charged with debris eroded from the col, and so deposition was enforced as the gradient slackened on the gentle hillsides at A. As the meltwaters continued further south, a new phase of erosion was initiated when the gradient gradually steepened again down towards the Lilburn valley, leading to the inception of channel 2.

Terrace (iii) is the most extensive of the four terraces. Lying approximately 12 feet below terrace (ii), it extends for 700 yards along the hillside and reaches a maximum width of nearly 100 yards. The terrace surface stands about 40 feet above a continuation of channel 1 beyond its rock-cut section through the col. A good exposure in the terrace edge at X reveals coarse deposits consisting of gravels and boulders in a sand and grit matrix. The exposure is not sufficiently large to show whether or not the material is sorted and bedded, but its calibre, together with the sub-rounded nature of the particles suggest that this material has not been transported very far; it was probably derived from the excavation of channel 1. The distal end of terrace (iii) does not lead into channel 2, but terminates where an esker rises approximately 15 feet above its surface. This rather broad ridge feeds into a small group of eskers, below which lies channel 2 on its western margin. There are several lines of evidence suggesting that the eskers pre-date channel 2:

(1) The channel contains no fluvioglacial deposits.

(2) It is incised below the eskers.

(3) It partially truncates the side of an adjacent esker.

(4) Esker C, which forms a southern extension of the system has been isolated from the group by channel 2.

Since terrace (iii) clearly post-dates channel 2, it must also post-date the eskers. Furthermore, the proximal end of the esker, against which terrace (iii) abuts, seems to have been partially trimmed by the stream responsible for the terrace.

This upper group of eskers thus represents the earliest phase of deposition on the slopes above South Middleton and was probably associated with the initial stages of channel 1.

Terrace (iii) merges into a vague linear mound, approximately 10 feet high, that is orientated obliquely downslope as though it had been diverted by the adjacent group of eskers. This feature fades into an area of shallow undulations that may represent an irregular spread of fluvioglacial deposits. The orientation of eskers 450 yards to the south-east certainly continues the oblique, downslope trend, and probably represents the same line of drainage; this finally assumes a broad, shallow, channel-like form and terminates on the side of Lilburn valley.

The surface of terrace (iv) stands 20 feet above the floor of channel 1 where the latter's conspicuous channel form terminates as it becomes a shallow marshy depression. This terrace can be traced along the hillside for 430 yards until it leads into a prominent ridge 45 feet high. The latter is aligned rather obliquely for a short distance downslope. The final trend assumed by this drainage is roughly parallel with the broad, marshy continuation of channel 1, in which low gravel ridges occur at two points. The esker-like continuation of terrace (iv) points directly towards the esker at K. The low esker at L, that curves round parallel with channel 12 as it begins to run directly downslope, illustrates the same phase of drainage. These fluvioglacial forms clearly represent the later stage of meltwater drainage during which free downslope movement became established under glacier ice in an advanced state of decay, as described in the previous chapter.

There is therefore a clear sequence of meltwater drainage development represented by the landforms on slopes above 775 feet near South Middleton; this can be summarised as follows:

1. The first stage involved meltwater drainage flowing predominantly in a south-easterly direction. It has already been demonstrated that channels 1 and 2 were probably cut by a superimposed englacial stream, and, as such, represent subglacial drainage. Terraces (i) and (ii) are closely connected with these channels, for their location strongly suggests that their deposition was effected by the meltwater river that flowed through both channels. The pronounced change in gradient encountered by meltwater emerging from the Brands Hill col was almost certainly responsible for the depositional phase illustrated by these terraces. Similarly, a renewed steepening in gradient caused the resumption of erosion as meltwaters flowed down into the Lilburn valley. Terraces (i) and (ii) may thus have formed subglacially. It has been suggested that channel 2 post-dates the eskers at B and C. The alignment of these eskers is in harmony with that of channel 1, and they probably represent a depositional phase of meltwater drainage from that channel. Since they are obviously the deposits of several streams, it is suggested that the river of meltwater from channel 1 initially became disseminated into numerous branching segments on this flatter hillside. Ultimately, when a route had become established in the depression leading to the Lilburn valley, these meltwaters became concentrated into one large tunnel in which terraces (i) and (ii) and channel 2 were formed.

2. The second stage involved the deposition of terrace (iii). The surface of this terrace lies slightly below the intake level of channel 2, and was probably deposited immediately after the latter had become abandoned.

The eskers at B appear to have been partially trimmed by this new phase of drainage and clearly pre-date it, but those lying at a lower level on the hillside, about point G, could possibly have been formed at this stage. Terrace (iii) is plainly linked with channel 1, and was almost certainly deposited by the meltwater stream issuing from that channel.

3. The third stage is represented chiefly by the deposition of terrace (iv), and still involves meltwater from channel 1. The short ridge that terminates this terrace appears to be orientated more obliquely downslope than the previous formations. Its alignment with esker K is striking, and it may well illustrate a stage in the major change of meltwater drainage from an ice-directed flow south-eastwards, to a free flow directly downslope. Esker L apparently belongs to this stage also.

4. The last stage seems to have followed soon after and probably involved the final drainage of meltwater through the Brands Hill col. The rather indistinct, marsh-filled continuation of channel 1 belongs to this period. Together with a stream of meltwater in channel 11, this river partly trimmed the base of esker K and certainly represents the establishment of free downslope drainage under extensively decayed glacier ice.

Throughout the remarkable development of these fluvioglacial landforms, channel 1 remained in operation. The various courses assumed by meltwater beyond this feature seem to have been largely determined by the establishment of drainage towards the south-east at first, and later, by relaxation of the glaciological control forcing water to flow in that direction, so that free downslope movement could occur. There is no evidence to suggest that channel 1 was ever utilised subaerially by meltwater (although this may well have occurred in later stages of downwastage if the ice tunnel had collapsed), but it is difficult to imagine the existence of an ice tunnel of sufficient dimension to contain the entire suite of fluvioglacial terraces associated with it. For example, a distance of 270 yards separates the back edge of terrace (i) from the front edge of terrace (iv) at one point. It seems clear that terraces (i) and (ii) were formed subglacially because of their relationship with the subglacial line of drainage responsible for channels 1 and 2, but the wider extent of terraces (iii) and (iv) is more difficult to account for with this hypothesis, particularly if it is assumed that they were deposited in the same tunnel. Consequently, it may be argued that they are kame terraces that were deposited at the glacier margin. Nevertheless, it is difficult to accept them as true marginal formations for the following reason. The melt-water stream responsible for the deposition of their sands and gravels clearly emerged from channel 1. Glacier ice must therefore have been present up to at least 975 feet in the Harthope valley before meltwater was able to discharge through this channel. Assuming that the ice margin would have curved round the east side of Brands Hill, terrace (iii) formed approximately 1,500 yards down-glacier from the intake of channel 1. Since the surface of this terrace lies at 850 feet, the ice margin must have sloped with a gradient of 147 feet per mile if these formations are strictly marginal in origin. This gradient is considered too steep to represent the slope of the Tweed glacier, for if it is projected up-glacier, it implies 3,500 feet of ice over the Kelso area and 9,000 feet over the Tweedsmuir Hills; this is clearly untenable. An alter-native suggestion, to avoid the difficulty of a vast subglacial chamber is that terrace (iii) formed subglacially following the collapse of the tunnel in which terraces (i) and (ii) were deposited; similarly, terrace (iv) may have formed when the ice tunnel occupied by terrace (iii) had caved in, otherwise, it was deposited in the same tunnel which had become up to 200 yards in width by this time.

The main group of eskers comprising the South Middleton system lies below 800 feet. It runs parallel with the hillside and with channels 3-7, some of which occupy depressions between ridges. Few eskers in the system are obviously connected with meltwater channels and must represent the former courses of quite separate streams. Furthermore, meltwater channels parallel with the ridges do not appear to truncate them, except for the two notable exceptions at 11 and 12. These observations suggest approximate contemporaneity for the majority of fluvioglacial features aligned in this fashion. Initially, channel 6/7 probably continued to the Lilburn valley and channel 10 may represent its former extension. The formation of this esker system therefore coincides with the period of ice-directed meltwater drainage during which a south-easterly alignment was predominant. At a subsequent period during deglaciation, the system was truncated by channels 11 and 12, eskers K and L, and an extension of channel 1, as meltwaters escaped directly downslope approximately at right-angles to the earlier system. Since the later group of features formed subglacially, as suggested above, then the main esker system that it truncates at a lower level must also be explained by the subglacial hypothesis.

(2) The Dod Hill System. Approximately one mile after leaving the Threestone Burn embayment, the Lilburn Burn turns sharply at right-angles from its easterly course and flows north (Maps 5 and 6). The precipitous western flank of Heddon Hill rises on the east bank to over 900 feet, and on the west, a remarkable formation of slumps and terracettes has resulted where the stream has severely undercut over 100 feet of drift. These massive drift deposits may extend quite thickly as far west as the steep slopes that rise to Dod Hill (1,125 feet), and, as such, may represent an extensive plug infilling a deep pre-existing valley. Although sections are generally poor or absent, the

drift is almost certainly thickest where it is incised by the Lilburn Burn and probably thins out towards the slopes of Dod Hill (Map 5). Numerous rabbit burrows in slumped terracettes reveal the nature of upper layers in the deposits showing them to consist of sand and gravel, but whether or not they overlie till is impossible to determine on the basis of present exposures. The upper surface of this drift infill, lying between 700 and 775 feet, is occasionally terrace-like and in places is surmounted by a small group of eskers and kettle holes. Elsewhere, only gentle and vague slopes are present. An intricate system of eskers between 10 and 15 feet high occurs at A. Together with some broader ridges, one of which becomes 35 feet high, these continue the south-easterly trend of meltwaters observed near South Middleton. Hummocks and small ridges at B, and the 20-foot rock-cut channel on Dod Hill apparently indicate a more easterly-directed flow and were probably formed by locally-derived meltwaters. Superficial exposures at several points on these eskers and terraces reveal rounded gravels, fragments of which elsewhere sporadically litter the ground surface when visible through the carpet of heather. Boulders up to 3 feet in size were seen overlying the gravels in places, but the nature of much of this vast infill of drift remains obscure. The eskers are not clearly truncated by the terraces and appear to overlie them, implying that they post-date terrace formation. Since the eskers seem to continue the lines of subglacial meltwater drainage observed near South Middleton, it would seem that the underlying terrace formations are subglacial in origin as well. The lack of exposures, of course, makes it difficult to determine accurately that every plane surface is a true depositional terrace of fluvioglacial deposits; several gently sloping expanses may simply be the upper surface of a till deposit, thinly veneered in places with patches of sands and gravels. An alternative explanation is that the phase of terrace deposition occurred later

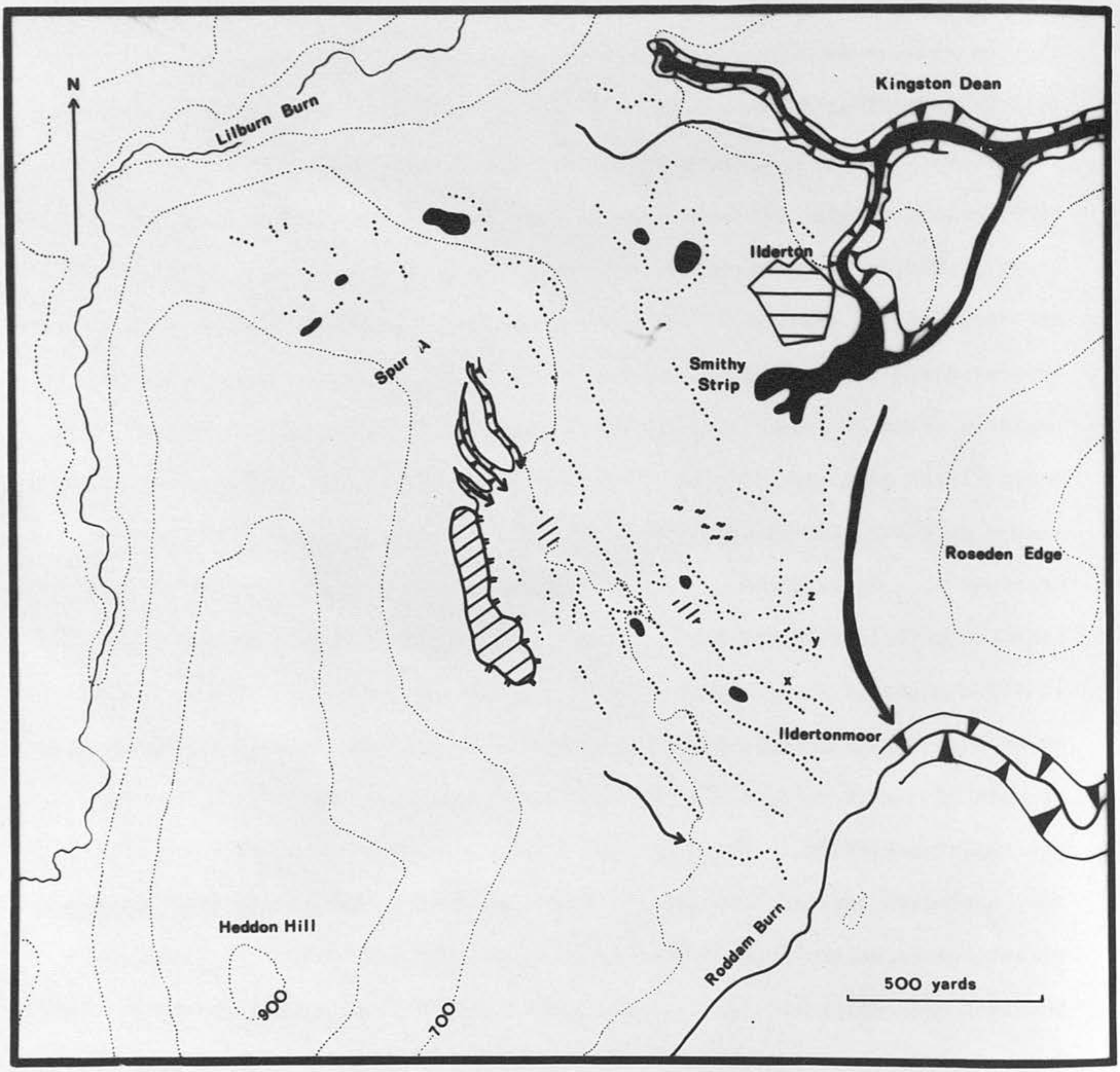

N
Lilburn Burn
Kingston Dean
Ilderton
Spur A
Smithy
Strip
Roseden Edge
z
y
x
Ildertonmoor
Heddon Hill
900
700
Roddam Burn
500 yards

Figure 3.2

and subaerially, so that they are banked up against the eskers, whose original form may be partly buried. Perhaps the major significance of these formations is that they continue the south-easterly alignment of meltwater drainage illustrated by the eskers and channels further north, near South Middleton.

South of Lilburn Burn, the low col that leads through to Roddam Burn is occupied by a wide, shallow meltwater channel. The drainage responsible for fluvioglacial formations in the vicinity of Dod Hill, together with some of that from the South Middleton system, must have converged on this col. It is surprising that the channel is so shallow, for the most part less than 15 feet deep, but this is perhaps counterbalanced by its width, which reaches 150 yards at one point.

(3) The Ilderton System. A group of conspicuous eskers is located on the lower slopes of Heddon Hill, above Ilderton village. Lying mainly between 525 and 650 feet, this system is bounded to the north and south by the valleys of the Lilburn and Roddam Burns respectively. Heddon Hill is a massive ridge of volcanic rock rising to 910 feet at its southern end. It is almost $1\frac{1}{2}$ miles long and $\frac{3}{4}$ mile broad. On its northern extremity, a broad spur (Spur A, Figure 3.2) leads off to the north-east and forms a watershed between Lilburn Burn and Roddam Burn. North of this spur, between it and the Lilburn valley, only vague mounds of fluvioglacial deposits exist, although more definite formations certainly appear round the lower end of the spur. The spur crest is devoid of meltwater features, but immediately beyond, on its south-eastern slope, three short meltwater channels are sharply defined. One of them is incised by over 15 feet into bedrock. On the moderately sloping hillside beyond these channels, lies the Ilderton esker system.

The highest fluvioglacial formations on the hillside, at 650 feet, comprise a line of former meltwater drainage that is represented by three separate landforms. Emerging from the spur side slightly above the outlet of the uppermost channel, a kame terrace slopes gently south-eastwards for approximately 400 yards. It is over 100 yards wide at its maximum development. Only one, rather meagre exposure reveals sand and gravel, but the root system of upturned trees and other minor exposures testify to the fluvioglacial nature of this formation. A gap of approximately 250 yards separates it from the 10-foot deep meltwater channel continuing the drainage line. This, in turn, merges into a small esker only 5 feet high, but which is nevertheless a distinct feature that winds down the valley-side and terminates 25 feet above Roddam Burn. Thus the uppermost line of drainage, although interrupted by a short gap, is represented in a most interesting manner by three different landforms. A short distance downslope from these, several low, linear mounds lead towards a broad esker consisting of coarse sands and gravels and this succession of features also continues the south-easterly trend of fluvioglacial formations. But it is on the same hillside, immediately below, that a system of extremely well-defined eskers forms the most distinctive group of landforms in this area. Rather vague ridges curve round the Heddon Hill spur from the Lilburn Burn valley and lead towards these eskers. The three rock-cut channels are also aligned in their direction, and it appears as though a considerable amount of meltwater drainage converged on the hillside south-east of the spur to produce the Ilderton esker system. Separated by depressions, kettle holes and short gaps, the eskers form an intricate network of dividing and uniting ridges. Cross profiles tend to be asymmetric, the lower sides facing uphill, and common heights reached are 10 feet on one side and 20 feet on the other. This presumably results from deposition on the moderately steep slope. Ridge crests are predominantly narrow, but occasionally broaden into flat-topped sections. Underlying material is exposed where farm tracks have been worn down through the deposit, and at a small, badly slumped gravel pit. These poor exposures suggest that the eskers are composed predominantly of small to medium-sized gravel ($\frac{1}{2}$ to 3 inches), contained in a matrix of coarse sand. The very sinuous esker, located 100 yards west of Ilderton village, develops into an extremely broad ridge at Smithy Strip, and, although not directly connected to the main group, it represents a similar alignment of meltwater drainage south-eastwards through the col between Heddon Hill and Roseden Edge.

In detail, the orientation of individual ridges in the lower part of this system is extremely interesting, for it illustrates a sequence of meltwater drainage development similar to that already described near South Middleton and to that displayed by some meltwater channels discussed in the previous chapter. The esker on which Ildertonmoor farm is situated, and eskers upslope from it trend almost parallel with the hillside in a south-easterly direction and terminate at about 550 feet, over 50 feet above Roddam Burn. The next esker in sequence (x, Figure 3.2), below Ildertonmoor, shows a distinct break from the general parallelism of the system as a whole, and is aligned more obliquely downslope in an east-south-east direction; it terminates at 550 feet. The adjacent esker several yards downslope (y, Figure 3.2) further develops this trend, and runs almost due east down the hillside at right-angles to the contours; it terminates as a double ridge at 525 feet. One segment of the next esker below (z, Figure 3.2) illustrates a similar orientation, but two distributary branches lead northwards to an area of low ridges that continue the northerly trend. Where these terminate, the broad depression feeding two tributaries of the Kingston Dean meltwater channel begins, thereby completing the drainage link.

The foregoing description illustrates a sequence of meltwater drainage development that seems to be characteristic of the north-east Cheviots. Early stages of the system, in the Ildertonmoor area, are represented by channels, a kame terrace and eskers orientated predominantly in a south-easterly direction. Meltwater responsible for these formations appear to have initially flowed across the site of the present Roddam valley previous to any drainage down it. The next stage is illustrated by esker x; the orientation of this ridge suggests that drainage down the Roddam channel had become established, to which meltwater responsible for esker x was tributary. This implies the inception of a relaxation in glaciological conditions that had directed meltwater drainage chiefly towards the south-east. In succeeding stages, eskers y and z demonstrate the progressive swing of fluvioglacial drainage in this area round to the north (although it was still flowing south-eastwards along hillsides immediately to the north-west); its outlet in this direction was into Kingston Dean, which had become established as a major channel evacuating meltwater directly downslope. It can only be supposed that either extensive collapse of ice tunnels aided this swing of drainage away from the Roddam channel, or else the latter was downcutting so rapidly that it caused the high-level abandonment of drainage tributary to it, thereby encouraging its diversion towards Kingston Dean. This later phase of drainage was almost in the opposite direction to that followed by earlier meltwaters.

(4) South of Roddam Dean. South of Roddam Dean, numerous eskers and kame terraces situated between 400 and 600 feet continue the line of former meltwater drainage from Ildertonmoor in a south to south-east direction. Pre-existing relief appears to have strongly influenced the path followed by fluvioglacial streams in this area. Immediately south of the col between the elongated masses of Heddon Hill and Roseden Edge, a third ridge, rising to 675 feet, forms a stalk to this Y-like assemblage of ridges. It will be referred to in this thesis as Fore Rigg (Map 8). Meltwater drainage from

Ildertonmoor appears to have flowed chiefly to the east of Fore Rigg and the few features indicating earlier flow west of that ridge will be described later in this chapter. The eskers at C, the four kame terraces at D, and eskers by E plainly illustrate a south to south-east alignment of meltwater drainage. Channel 27 shares a similar trend and, undoubtedly, the deep canyon of Brandon Dean served as a major outlet for many of the streams responsible for these landforms. Orientated roughly at right-angles to this group of fluvioglacial formations, another distinct series of channels and eskers demonstrates that a subsequent flow of meltwaters escaped more directly downslope towards the north-east. Their later age is established by the fact that they truncate the former group. Channel 28, together with its tributaries and adjacent fluvio-glacial ridges, probably represents an initial dislocation to the south-easterly directed drainage system in this area; it presumably caused the abandonment of Brandon Dean. The route taken by channel 28 and associated features goes down the northern flanks of Nova Scotia, with the main channel occupying a pre-existing embayment. Eskers and small channels mapped on the north-east slopes of Brandon Hill nearby probably depict the same phase of meltwater drainage. The exact mechanism by which Brandon Dean was abandoned is difficult to recognise, but the collapse of glacier ice that was becoming increasingly decayed during this later stage of downwastage, may perhaps have blocked the route to Brandon Dean (which is quite shallow in its upper reaches); this suggestion can only be conjectural in the light of present evidence.

Channel 26 turns abruptly downslope and isolates its former extension south-eastwards past Roddamrigg. It seems likely that much of the meltwater flowing towards channel 27 would have become diverted down channel 26, so that the former became almost totally forsaken. A group of sharp-crested eskers, some of which are 30 feet high, and several minor channels, occur immediately east of channel 26, and appear to illustrate a similar drainage alignment leading towards the wide depression west of Wooperton that was earlier interpreted as a dead-ice hollow. Meltwater streams that converged in this area presumably flowed englacially across the site of the hollow, for no trace of them has been left on the ground. Their combined waters finally escaped north-eastwards through the channel at Wooperton village. The final stage of drainage diversion in the vicinity of Roddam, during which Roddam Dean became established as a major meltwater channel truncating the drainage line between Ildertonmoor and Roddam, has already been described. Indeterminate ridges of sand and gravel above the north wall of Roddam Dean probably represent an early stage of fluvioglacial drainage down this pre-existing depression, prior to the deep incision that formed the channel.

(5) The Reaveley Terrace. Between Reaveley Hill and Fore Rigg a wide, embayment like valley opens south towards the Breamish from a low col at just over 600 feet near the Roddam Burn (Map 8). The floor of this col and the upper reaches of the embayment are plugged by an extensive terrace composed of fluvioglacial sands and gravels. Numerous small kettle holes pit the terrace surface; ice-contact slopes and dead-ice hollows form its margins and it seems to have been partly dissected on the east by a meltwater channel. The terrace slopes visibly, falling from 667 feet (O.S. spot height) in the north to 600 feet in the south. At this point the terrace form of the deposits ends and four eskers continue for a further hundred to two hundred yards down valley. The maximum dimensions of the terrace are approximately 1,000 by 1,500 yards. The proximal end of the terrace has been severely truncated by the Roddam Burn which now flows almost 60 feet below the terrace surface at one point. The terrace fragments that occur on the lower slopes of Heddon Hill below 700 feet on the other side of the Roddam Burn may therefore be remnants of the initial

expanse of the Reaveley terrace.

The slope of this terrace indicates fluvioglacial drainage predominantly in a south-easterly direction and conforms with the general pattern already established on the basis of channels, eskers and kame terraces in adjacent areas. Considerable volumes of meltwater appear to have entered the Roddam valley immediately west of the terrace: for example; (a) the broad meltwater channel in the col west of Heddon Hill descends from the north and terminates in the Roddam valley at 675-700 feet; (b) the impressive meltwater channel (31, Map 7) in the col west of Reaveley Hill brought considerable volumes of water from the south. Terminating at about 750 feet, the latter was almost certainly associated with an ice mass occupying the Breamish valley. It is difficult to determine whether or not these two channels functioned contemporaneously, but both probably contributed much of the sands and gravels that compose the Reaveley terrace.

During this period of drainage a route down the lower Roddam valley was not available, but the wide, pre-existing valley between Reaveley Hill and Fore Rigg afforded a low pass down which meltwaters could escape towards the Breamish valley. Although the rivers of meltwater that entered the Reaveley-Fore Rigg col appear to have issued from subglacial courses, it is unlikely that the broad terrace was deposited beneath glacier ice. Gjessing (1960, 1965) mentioned the possibility that under certain topographic circumstances, a subglacial sheet-like drainage could occur "when the water pressure (the pressure in the pore water, or the ground water, as well as in the free water) at the bottom of the ice counterbalanced the pressure of the ice so that the ice was lifted". He also observed that "Where tributary rivers enter the valleys, large subglacial fans are found. Contiguous, often braided eskers and other deposits indicate the further courses of the river systems through ice tunnels along the valley floors." The Reaveley terrace and its four distributary eskers may be thought analogous to the above landforms, described from southern Norway, and it may be thought that they have a similar origin. The validity of Gjessing's interpretation may be questioned, however, for it is by no means clear why meltwater rivers confined to definite subglacial tunnels should suddenly merge and form an extensive fan or terrace and then resume individual courses through subglacial tunnels. Such a theory is very difficult to apply to the Reaveley terrace and associated deposits. The abundant ice-contact edges, dead-ice hollows and kettle holes points to deposition in contact with stagnant and broken glacier ice. The four distributary eskers presumably represent former subglacial courses, but the main terrace mass was most likely built up chiefly in a sub-aerial environment, the deposits being banked against walls of stagnant ice and burying smaller, detached remnants that subsequently melted out to form kettle holes and dead-ice hollows.

Since all signs of fluvioglacial activity in the Reaveley valley terminate just below 600 feet, it may be deduced that the meltwater streams thereafter flowed along englacial tunnels through stagnant ice lying in the Breamish valley. Several cols whose crests are below 600 feet lead south from the Breamish valley at this point across the Breamish/Aln watershed, and the englacial flow of meltwater was probably maintained through these, thereby leaving no trace on the ground. The uppermost signs of meltwater erosion through this watershed are 425 feet in the Crawley Dean col farther east, but these are unlikely to be associated with this earlier and higher level of meltwater drainage.

The meltwater channel that partly cuts through the Reaveley terrace is interpreted as the course taken by water from blocks of stagnant ice marginal to the terrace as they slowly melted out.

Subsequent to the phase of drainage through the Reaveley valley, meltwater appears to have flowed down the Roddam valley as a route opened up in that direction. The northern edge of the Reaveley terrace is clearly truncated and adjacent ridges of fluvioglacial materials are aligned down-valley. Two kame terraces have been built around these ridges at a lower elevation and seem to be connected with the same level of deposition as that indicated by the uppermost kame terrace south of Roddam (Maps 6 and 8). Assuming that the latter are connected, this again indicates a swing of meltwaters towards the south-east and suggests that Roddam Dean had not yet been formed. Ultimately, a route directly downslope towards the east became available and Roddam Dean functioned as a major outlet for fluvioglacial streams and drainage from ice-free hillsides to the west. The Roddam Dean channel is an impressive canyon in which precipitous walls, rising over 100 feet high, have been carved out of conglomerate bedrock. Its narrow, gorge-like form begins immediately below 500 feet, so that it truncates the adjacent system of eskers and terraces that is aligned south-eastwards at this elevation. Channel form dies away at 300 feet, but water issuing from this valley appears to have truncated the great belt of eskers that lies along the hill-foot zone from Wooler to Percy's Cross at this height. This truncation is unlikely to have occurred until a subsequent phase of drainage, however, and meltwater from Roddam Dean probably joined the south-easterly directed drainage along the hill-foot at this time.

(6) The Brandon Terraces (Map 8). A wide embayment forms a pronounced indentation in the south-facing side of the Breamish valley a short distance north of Brandon hamlet; it is most probably a pre-glacial element of the topography. The eastern side of this embayment has been deeply dissected by a meltwater channel leading south from beyond the watershed and exhibiting a most sinuous

course in its upper reaches. The cross profile of the lower part of this channel is strongly asymmetrical. Rising almost 150 feet high, the east wall is cut chiefly in bedrock, whereas the opposite wall is much lower and has been cut through sand and gravel. Clearly, a phase of fluvioglacial deposition preceded erosion. These deposits lie between the channel and the central axis of the embayment between 375 and 475 feet and are arranged in the form of four narrow terraces, usually no more than 100 yards wide. The most extensive terrace of this suite is fringed by a very crenellate ice-contact margin. Brandon Dean is part of the extensive network of subglacial meltwater channels, aligned in a south-easterly direction, that was referred to in the previous chapter, and since it cuts through the terraces, it seems valid to interpret the latter as subglacial formations of an earlier date. In view of the relatively narrow extent of these terraces, it is possible that they were deposited in subglacial caverns within stagnant glacier ice occupying the Breamish valley. The source of the meltwater responsible for this deposition is more uncertain, however, and there appear to be three alternatives.

(a) All the meltwater came down the Breamish valley, flowing at between 375 feet and 475 feet, and not from Brandon Dean.

(b) The terraces were deposited entirely by meltwater that flowed through the upper reaches of Brandon Dean prior to the main erosive phase that excavated the lower 75 feet of the channel.

(c) The terraces represent a combination of (a) and (b).

A certain measure of combination between the first two alternatives is suggested on two accounts. Firstly, the main expanse of the widest terrace and its surface slope from north to south, are best explained by drainage from the west into the Brandon embayment, for it is difficult to envisage meltwater from the Brandon channel curving sharply westwards and then northwards in an uphill direction. On the other hand, a short distance north of the Brandon Dean intake, there are four kame terraces arranged in step-like fashion, one above the other, similar to those at Brandon. The two sets are perhaps too far removed from each other for accurate correlation, but the fact that four levels of meltwater deposition are represented by two similar sets of landforms within only $1\frac{1}{4}$ miles of each other, is perhaps more than coincidence. The uppermost terrace at Roddam lies approximately at 550 feet, that at Brandon, at 460 feet. If the two terrace groups are connected, then the Roddamrigg-Brandon channel provided the link. The Brandon terraces are thus possibly explained by the confluence of meltwaters from two different sources. Ultimately, the last drop in base level controlling fluvioglacial erosion and deposition in this area effected the main excavation of Brandon Dean and the partial truncation of the terrace sequence.

The Upper Systems - Conclusion. The various esker systems, kame terraces and associated meltwater channels that occur liberally on the flanks of the north-east Cheviots above 400 feet, have been resolved into two main groups. The earlier streams of meltwater were directed chiefly towards the south-east, trending roughly parallel with the hillsides. This system is closely related to the major phase of meltwater drainage in the north-east Cheviots, during which numerous channels became incised into cols by the superimposition of en-glacial streams. Terraces and eskers above South Middleton are plainly connected with the Brands Hill channel which originated in this fashion. Small patches of possible ablation moraine, frequently in the form of large sub-angular blocks dispersed over fluvioglacial deposits beneath glacier ice. Their relationships to subglacial meltwater channels perhaps endorse this suggestion more firmly. Accordingly, they demonstrate extensive deposition of sands and gravels in an englacial or subglacial environment. Recent work by

Price (1964) suggested that eskers can become superimposed to the ground from former englacial channels, and while some of those in the north-east Cheviots may have had such an origin it is difficult to demonstrate this with reference to any particular group.

Eskers and channels comprising the second major group of fluvioglacial features are orientated roughly at right-angles to the previous group and run more directly downslope. Since they seem to truncate the earlier group in places, they therefore represent a later phase of drainage that began to function when ice-directed flow to the south and south-east became abandoned at the upper levels. This enabled a free downslope movement to become established and suggests that glacier ice, lying in the foothill basins and on lower slopes of the massif, had become extensively decayed and crevassed at least above 400 feet at this stage.

The small system of low eskers lying above 400 feet on the dip-slope of Weetwood Moor occurs somewhat in isolation from the main mass of fluvioglacial deposits in the north-east Cheviots. Apart from illustrating that some meltwater streams flowed south-eastwards in this area it is of little significance, and was probably associated with the first phase of drainage described above.

B. The Lower Systems.

(1) The Wooler Area. Peripheral slopes of the north-east Cheviot massif descend approximately to 400 feet where they generally terminate. Below this level lies the fringing sub-Cheviot depression, composed, for the most part, of the Wooler, Lilburn, Chatton and Hedgeley Basins. From Wooler to Percy's Cross esker systems extend with almost unbroken continuity along the western margins of these basins (Maps 5, 6 and 8). With few exceptions,

Photograph 3.c Eskers just south of Wooler.

Photograph 3.d Ice contact topography near Wooperton.

eskers within the systems located between 300 and 400 feet are generally aligned in a direction that varies from south to south-east, similar to those in the Higher Group. Esker systems below 300 feet are mostly orientated between east and north. The complex of esker systems, as a whole, begins on the flanks of Horsdon Hill immediately south of Wooler, where narrow kame terraces merge into steep-sided eskers, some of which are over 50 feet high. The broad ridges of sand and gravel on which Wooler is situated are probably linked with this system. Whereas the kame terraces slope southwards, almost parallel with the hillside contours, the eskers lead off abruptly at a relatively steep angle down the side of the Wooler Water valley (Photograph 3.c). They have obviously been truncated by the present stream and were formerly much more extensive. There are no meltwater channels directly connected with these particular features, but they appear to have been associated with fluvioglacial streams from the north-west that drained round the north-east spur of Horsdon Hill. A few hundred yards to the south, in the vicinity of Earle Mill, the eskers are aligned slightly more parallel with the present river, and this has allowed their preservation as a more continuous system; here, they are orientated towards the south-south-east. From this point southwards to Coldgate Water, meltwater rivers issuing from the Horsdon and Earle channels, in conjunction with those from the north, deposited a remarkable system of anastomosing eskers. Three eskers clearly emerge from the outlet of the Horsdon channel; similarly, a kame terrace and esker at Earle seem closely related to the Earle Whin channel. The majority are low features, not more than 20 or 30 feet high, with slopes that appear to have been so continually cultivated that much original form has been destroyed, and they are probably less steep now than initially. Farther south, towards Coldgate Water, the forms become more massive, and rise over 40 feet above adjacent hollows.

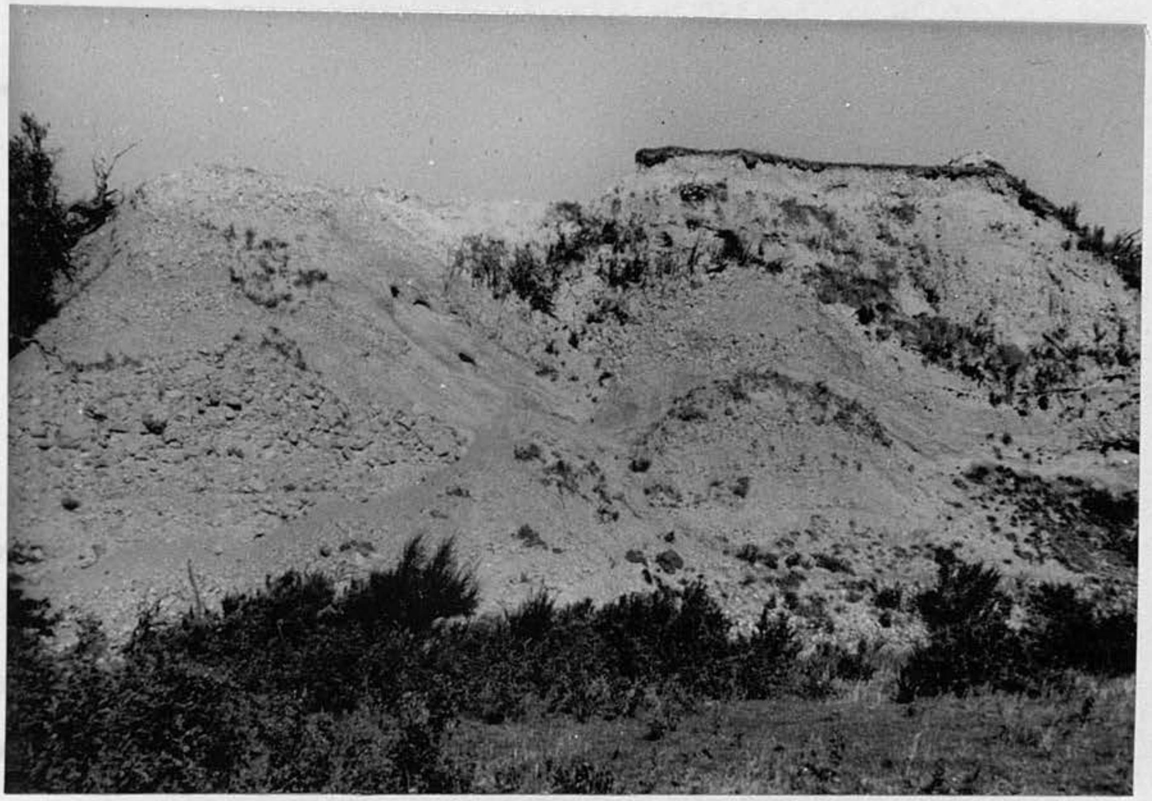
Undercut bankings along the Wooler Water, however, reveal up to 100 feet of fluvioglacial deposits. This suggests that the eskers are only relatively minor, superficial features overlying a much thicker deposit (the base of which has not been observed; the deposit may be much more than 100 feet). The esker system is liberally pitted with kettle holes, several of which contain water (Photograph 3.d) or marshy vegetation, while the majority of others stand out clearly in the landscape as uncultivated enclaves in which rough vegetation grows, or into which large blocks cleared from the fields have been dumped. The much higher density of kettle holes amongst these eskers contrasts notably with that of the Higher Group described above. Although the deposits are extensively undercut all along the left bank of the Wooler Water, their internal composition is exposed in only four sections, because considerable slumping and dense vegetation form thick screens along the greatest part of the bank. The material is roughly similar in all four sections, and, for the most part consists of bedded silts, sands, gravels and cobbles. In more detail, the sections are as follows:

Site 1 (Grid Ref. 3996/6266). The upper 30 feet is composed of dirty sand, gravel and cobbles, containing small contorted inclusions of clay and silt. Below a sharp unconformity, the lower 20 feet consists of bedded gravel, grit, sand, silt, and laminated clay; the latter is pink and grey/brown in colour. Faults, lenses and inclusions are characteristic. The final 10 to 15 feet down to river level is obscured by slumped debris.

Site 2 (Grid Ref. 3997/6264). The upper 25 feet consists of dirty gravel and cobbles contained in a coarse sand and grit matrix - similar to the upper part of section 1.

Photograph 3.e (i) The section in fluvioglacial deposits on the left bank of the Wooler Water, near Haugh Head.

Photograph 3.e (ii) Ablation till overlying the fluvioglacial deposits in Photograph 3.e (i). The fluvioglacial sediments begin at the hammer head.



Beneath, occurs, 6 inches–2 feet ..... clay, partly slumped into position.

1 foot 6 inches ..... laminated silt and clay: red/brown and green/brown in colour.

7 feet 6 inches ..... fine sand and silt: finely bedded; pink and grey/green in colour.

22 feet ..... obscured to river level by slumped debris.

Site 3 (Grid Ref. 4002/6258). The bank here is only 20 feet high. In this section a deposit of extremely coarse material is exposed from top to bottom. Water-worn cobbles and boulders up to 3 feet in diameter are crudely bedded with a southerly dip and are cemented together along with fine gravel. Sandstone boulders of various types are quite common along with igneous material, chiefly andesite, but much more significant is the considerable number of the former that are so rotted that they virtually disintegrate at a touch. It is extremely unlikely that these were transported in such a condition, unless frozen solid. If they were originally transported in a fresh condition, it is difficult to understand by what means and under what conditions they subsequently rotted to their present state. They occur at various levels within the deposit and no evidence of weathering horizons is apparent; surrounding stones remain quite fresh.

Site 4 (Grid Ref. 4002/6256). This section is no longer in the condition in which it was first observed by the writer in 1962, and by Derbyshire a few years previously. At that time, a true cross-section of the ridge in which it was exposed could be observed (Photograph 3.e). Recent sand and gravel workings have destroyed this face. The section revealed bedded fluvioglacial deposits exhibiting a remarkable range in grain size; water-worn boulders up to 5 feet across were interbedded with very fine sand. Above a 10-foot layer of boulders and cobbles in the lower part of the section, overlying beds were arranged in two great undulations, in which layers of sand could be readily observed at a distance by the rows of holes excavated by Sand-Martins. The layers of enormous boulders demonstrate that meltwater streams of immense power flowed in this area. The conspicuous variety in particle size suggests variable flow and the undulatory nature of the bedding is interpreted as a later slump structure. Derbyshire observed a tendency to easterly dip in this bedding and "its decline in altitude from the south-west to the north-east indicating flow from the Cheviot slopes". Since this section was almost at right-angles to the trend of the ridge in which it occurred, Derbyshire's suggestion is untenable. Had the latter mapped depositional landforms in detail, he might have reached a different conclusion, for the direction in which the water flowed was most likely to have been approximately parallel with the ridge crest, i.e. from north-west to south-east. The present section (1966) being worked by the gravel firm is aligned almost parallel with the ridge, and, as such, represents a longitudinal exposure of the material. The layer of large boulders is no longer to be seen, but the other layers are similar to those previously exposed. Sand, gravel and cobbles are all inter-bedded with one another in an extremely irregular manner, sand layers frequently giving way to beds of gravel and looking as if they have been pinched out. There is no predominant dip in the beds that suggests the direction in which the meltwater flowed, since some dip south-eastwards, others north-westwards, and much slumping seems to have followed the original deposition. Overlying these deposits is material quite different in nature. It was exposed more clearly at the bank top before economic exploitation of the gravels began, and a 3-foot layer extended quite continuously over the underlying deposits (Photograph 3.e). It consists of a dark, red-brown material composed of unsorted stones thinly scattered within a matrix of sandy silt. The stones vary in shape from angular to rounded and a slight clay content is occasionally present in the matrix. Quite distinct in texture and composition from the underlying sands and gravels, this deposit is interpreted as ablation till and is discussed in more detail in Chapter 9.

(2) The Coldgate Area. The first pronounced break in the northern part of the Wooler-Percy's Cross esker belt occurs where the Coldgate/Wooler Water emerges from the volcanic massif and truncates the esker belt at right-angles to its alignment, before bending sharply northwards to enter the Milfield Plain. Between Haugh Head and Wooler, a prominent gravel plain, 300 to 450 yards wide, has been developed by the lateral erosion and redistribution of these fluvioglacial deposits, accomplished by the Coldgate/Wooler Water.

South of Middleton Hall, an unknown depth of sand and gravel forms a vague, undulating spread lacking distinct ridge forms; it may represent the deposit of meltwaters that flowed down the line of the Coldgate valley. Bedrock is exposed at the bank foot south of Broom Crook (B.C. Map 5), but the extent to which this underlies the Broom Crook ridge remains uncertain. About Broom Crook, superficial sands and gravels assume definite esker form, and ridges up to 25 feet high indicate meltwater drainage east-south-eastwards out of the Coldgate valley, to join the main Wooler-Percy's Cross esker belt. The fan-like pattern assumed by these eskers can be clearly seen on Maps 5 and 6.

(3) Coldgate Water to Lilburn Burn. The line of meltwater drainage represented by the esker system to the north is continued south of the Coldgate/Wooler Water by equally distinct eskers. Kettle holes and dead-ice hollows are also associated with these ridges (Map 6). Just over a mile to the south-east of the Coldgate/Wooler Water, the valley of the Lilburn Burn similarly truncates the eskers roughly at right-angles to their alignment. Varying in width from

200 to 400 yards, the valley is incised to depths that range from 5 to 80 feet, reflecting the cross-sections of eskers and intervening depressions. Extensive slumping, a stabilised cover of trees, and undergrowth, obscure any sections that may have been revealed in the past. The present stream channel is located away from the valley sides, so that the Lilburn Burn does not actively undercut them at any point and no sections are currently exposed in the truncated eskers. Although the majority of these formations are aligned south-eastwards, they appear to have been intersected by a later set of forms that illustrate meltwater drainage curving eastwards round the southern flanks of the Fell Sandstone dip-slope, towards the Till valley. Kame terraces at U and V (Map 6), whose accordant surfaces presumably represent a contemporaneous level of deposition, clearly truncate several of the Coldgate eskers and two adjacent ridges which rise prominently over 35 feet above terrace level. The kettle holes in the surface of terrace V, the crenellate ice-contact margin of U, together with the large intervening kettle hole, confirm that the deposition was around and probably on top of large blocks of dead ice. The enormous kettle hole or dead-ice hollow at W appears to represent the site of a former mass of dead-ice that was not fully buried by fluvioglacial deposits. Numerous eskers curve eastwards towards it, distinct from the south-easterly trend of the main belt. Broad swells, composed of fluvioglacial deposits, east of kettle hole W, contrast notably with the more sharply-defined ridge form of eskers to the west. This is possibly explained by a more disseminated flow of meltwater issuing from dead ice in the kettle hole. The final stages of meltwater drainage in this area appear to be represented by the shallow channel that partly truncates terrace U, and also by the small outlet from kettle hole W to the Lilburn Burn, eroded, no doubt, by meltwater from the dissipating block of ice.

(4) South of Lilburn Burn. South of Lilburn Burn, the main esker belt attains its fullest development (Map 6). Individual ridges may be traced continuously for over a mile in some instances, but the whole system is predominantly characterised by complex groups of ridges that continually divide and unite round kettle holes. Elongated dead-ice hollows frequently separate eskers, several of which develop flat-topped crests in places, particularly at their highest parts. The feature at T is perhaps best termed a flat-topped mound, but its distinct linear alignment parallel with the esker belt clearly indicates its association with the same system of meltwater drainage. Eskers comprising this system are directed predominantly between south and south-east, and lie chiefly to the west of the minor road that runs south-west from East Lilburn to join the A 697 near Roseden Cottage (R.C. Map 6). On either side of Roseden Cottage, the eskers are breached by the lower reaches of the Kingston Dean meltwater channel. The latter therefore post-dates the south-easterly directed phase of fluvioglacial drainage. To the south, and lying mainly west of the A 697, eskers and kame terraces continue the extensive belt of fluvioglacial deposits in a southerly direction towards the Roddam Burn. Along the foot of Roseden Edge, these features continue to the very outlet of the Roddam Dean meltwater channel which appears to have been cut through them at a later date. Indeed, the severed extensions of the eskers are clearly seen south of Roddam Burn, where they continue towards Wooperton. At two places on the south bank of Roddam Burn, immediately beyond the gorge, and $\frac{1}{4}$ mile farther east, reasonably clear sections in the fluvioglacial deposits have been exposed by the recent undercutting action of the stream. The following are revealed:

Site 1 (Grid Ref. 4033/6209).

15 feet-18 feet	largely obscured by rain-wash, but sufficiently clear to identify beds of poorly-sorted, water-worn gravel and cobbles alternating with bedded layers of fine sand.
6 feet	unsorted, unbedded admixture of grit, gravel and cobbles (up to 12 inches), many of which are well-rounded; several are fragments of sandstone.
4 inches	bedded, fine sand.
1 foot 6 inches	grit, gravel and cobbles - some of which are well-rounded.
4 feet-5 feet	obscured by slumped debris to stream level.

Site 2 (Grid Ref. 4036/6207).

3 feet	bedded sand and medium gravel - partly weathered and disturbed by soil development.
4 feet	coarser gravel, unsorted but well rounded; some pieces of igneous rock are rounded like eggs.
10 feet	partly obscured by rain-wash and slump, but well-bedded fine sand and layers of small gravel could be distinguished.
3 feet-3 feet 6 inches	grit and small gravel.
2 feet	sub-rounded gravel, 2 inches-3 inches in size.
2 feet	bedded fine sand.
2 feet	obscured to stream level by slumped debris.

Both sections show characteristics of internal composition typically associated with eskers, namely, irregularly bedded sands, gravels and cobbles, predominantly well-worn by water action. Perhaps the most interesting information yielded by exposures in this area is from the bank top several yards downstream from section 2. Here, 1 foot 3 inches of till rests on top of the fluvioglacial deposits. It consists, chiefly of red clay that contains a scattering of stones, some of which are sub-rounded. As such, it is extremely similar in its composition and location to the material overlying fluvioglacial deposits at Haugh Head, described previously in this chapter. Its only difference is the rather more tenacious consistency. This till is probably similar in origin to that at Haugh Head and is interpreted as ablation moraine.

From the outlet of the Roddam Dean channel there radiates a fan-like pattern of eskers partly truncating and partly confused with the south-easterly directed system. Ridges comprising this group run chiefly towards the north-east and clearly post-date the main belt. They will be discussed at greater length in a subsequent section of this chapter.

South-east of Wooperton, between the lower flanks of Nova Scotia and the A 697, a remarkable group of eskers extends over a distance of half a mile (Maps 6 and 8). The most continuous ridges form a central artery to the system and clearly run towards the south-east, but the unusual characteristic of the system is the orientation of numerous branch eskers apparently feeding the central stem. These join at right-angles from the north-east and south-west, thereby creating an almost trellis system of eskers. The group, as a whole, contains perhaps the most sharply defined ridges in the north-east Cheviots. Steep, ice-contact slopes rise to crests over 70 feet high, the highest points being frequently located where ridges converge. Similarly, intervening kettle holes are deep and extremely well-preserved. It is considered that the central stem represents the main south-easterly alignment of meltwater drainage and continues the esker belt from the north-west. The ridges at right-angles to it, however, are probably not associated with that phase of drainage, and are believed to represent a later stage during which deposits were spread across the former system at right-angles to it.

Towards Percy's Cross, eskers that continue the south-easterly directed belt become less sharp and more massive and are margined by large dead-ice hollows elongated south-eastwards. The entire system dies away in vague undulating topography near Percy's Cross, beyond which no further esker formations exist.

The Lower Systems - Conclusion. Apart from the relatively narrow transverse gaps in these systems that were cut by later drainage, the complex belt of anastomosing eskers, kame terraces, kettle holes and dead-ice hollows may be traced in a general south to south-east direction for a distance of approximately $6\frac{3}{4}$ miles from Wooler to Percy's Cross. Situated for the most part between 300 and 400 feet, it represents a phase of meltwater drainage aligned along the foothill fringe of the north-east Cheviots at its junction with the adjacent basins of Wooler, Lilburn and Hedgeley. Many of the meltwater streams feeding the system appear to have come from various sources, that include - (a) the great systems of meltwater drainage flowing round the north-east Cheviots; (b) some of the adjacent channel systems, for example, the Horsdon and Earle channels; (c) drainage down the Coldgate valley and other outlets for drainage from the interior of the Cheviot massif; (d) minor sources on Weetwood Moor.

Although some of the kame terrace formations in the system may have been deposited subaerially, much of the great esker belt may be subglacial in origin. In support of this interpretation is the following evidence: (a) the clear relationship of some eskers to the subglacial channel systems; (b) the subsequent truncation of many eskers by meltwater channels, also of subglacial origin (Chapter 2); (c) up to 3 feet of ablation till overlying fluvioglacial deposits that compose the eskers - observed at two sections. A more detailed discussion concerning the origin of the eskers is deferred until Chapter 9.

A great number of well-defined kettle holes, many of which contain water or marshy vegetation, are liberally strewn throughout the system. Several are of considerable dimensions, reaching up to 400 yards across, but the majority are much smaller. The high density of these features and of the numerous dead-ice hollows, genetically similar, demonstrates that fluvioglacial deposition responsible for this vast system of landforms was associated with glacier ice in an advanced state of decay. Indeed, it must have been stagnant.

Although this system of landforms fades out on the northern fringe of the Hedgeley Basin, $1\frac{3}{4}$ miles from the intake of Crawley Dean, there is little doubt that this large channel controlled the evacuation of meltwaters that flooded into the Hedgeley Basin from the north. Accordingly, it determined the upper level of deposition at any one time. Since this control could have operated only through the ice itself, the ice must have been extensively decayed with water flowing freely through it, as suggested previously. Since the crest in the floor profile of Crawley Dean lies at between 300 and 325 feet, it is to be expected that the higher parts of eskers and terraces directed towards it would be built up to a level not much lower than that elevation. This is normally the case, and the few exceptions below 300 feet generally lie above 275 feet, which is not inconsistent with this line of drainage since the intake to Crawley Dean also begins at that height.

The absence of eskers and associated formations between Percy's Cross and Crawley Dean is thought to be related to subsequent developments in the fluvioglacial drainage system that may have destroyed any former extension of the system in this area; these are discussed below.

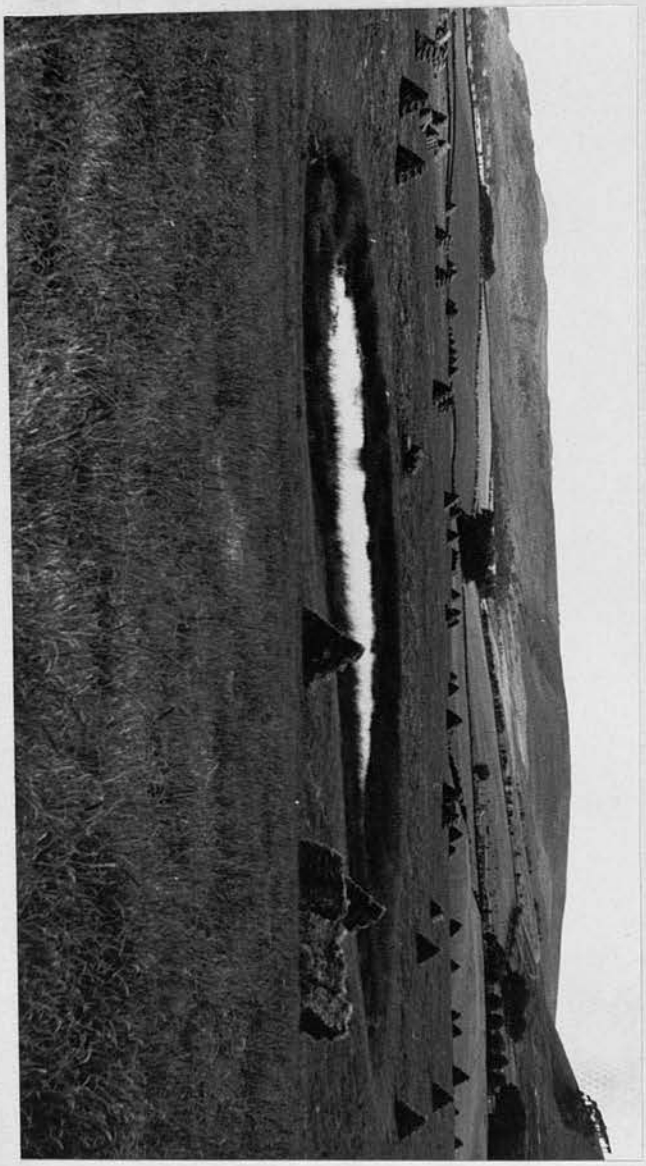
C. The Hedgeley Glacial Lake.

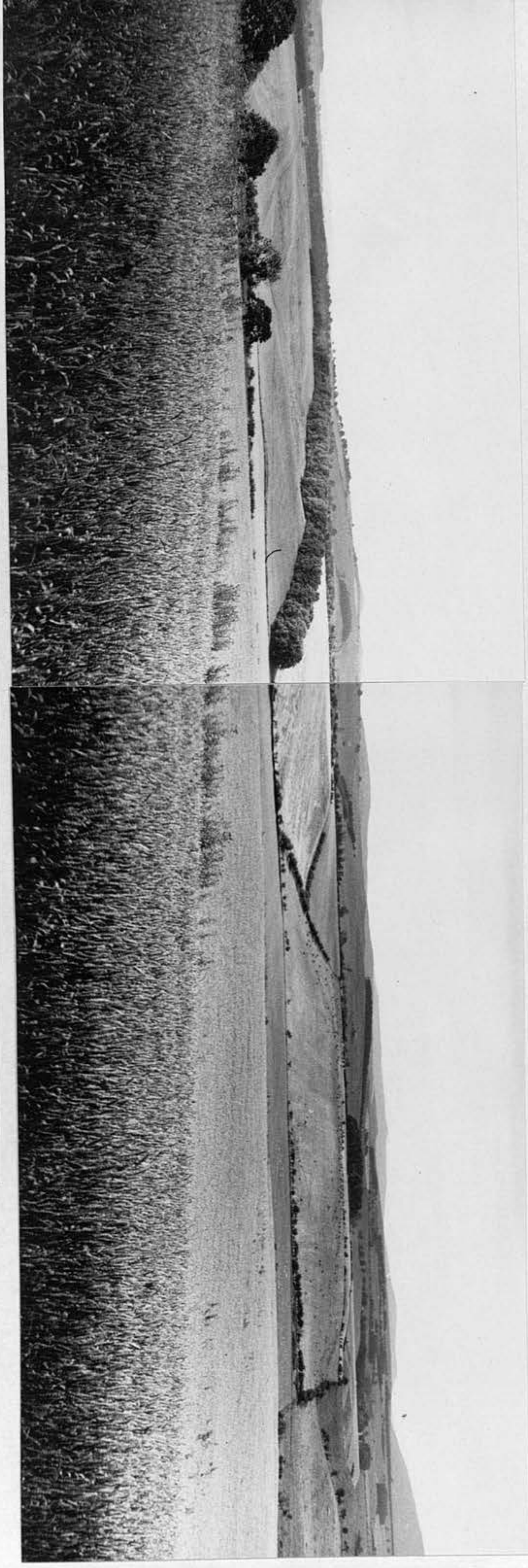
It has already been demonstrated in the previous chapter that the phase of south-easterly directed meltwater drainage was followed by a period during which fluvioglacial streams and water from ice-free areas within the Cheviot massif flowed more directly downslope towards the Till valley. The former drainage entered the Hedgeley Basin chiefly in englacial and subglacial tunnels and cut an outlet through the watershed to the south-east subglacially, forming the Crawley Dean channel, until the level of meltwater flow lowered to approximately between 300 and 325 feet. The abandonment of Crawley Dean as a subglacial channel appears to have coincided with the establishment of freely draining meltwater routes downslope towards the Till valley and evacuation to the north, presumably through and beneath stagnant glacier ice. There is, however, considerable evidence that suggests an intermediate stage in this major realignment of the drainage characterised by the formation of a pro-glacial ice-dammed lake in the Hedgeley Basin. During this intermediate stage it seems likely that the lake extended northwards into the ice, forming an englacial water-table. Many of the eskers aligned south-eastwards below 400 feet were probably formed at such a time and the englacial water-table determined the upper level of deposition. The surface level of the lake was controlled by Crawley Dean, which functioned as an overflow until the drainage route northwards had opened up and the lake disappeared.

Since prominent landforms and deposits were created in association

Photograph 3.f The Hedgeley outwash delta viewed from the north-north-east.

Photograph 3.g Kettle hole in the Hedgeley outwash delta.





with the lake, it is necessary to devote the following section of this chapter to their description and interpretation, before proceeding to deal with the lower group of esker systems in which the eskers occur chiefly below 300 feet and are not orientated towards the south-east. The latter were deposited perhaps partly contemporaneously with, but chiefly subsequent to, the glacial lake period.

(1) The Wooperton Delta. Between Wooperton and New Bewick stretches one of the most remarkable landforms of fluvioglacial deposition in the east Cheviot area. It is a plateau-like feature, the maximum dimensions of which are approximately 2,100 yards from west to east, and 1,500 yards from north to south (Maps 6 and 8). The surface of this plateau slopes quite gently predominantly from west to east (Photograph 3.f), but gentle inclinations descend also towards the north-east and south-east, so that the whole feature slopes, fan-like, from a western apex. Rising up to 80 feet above fringing depressions, the plateau is heavily pitted by at least 25 kettle holes (c.f. Photograph 3.g) and dead-ice hollows that impart considerable irregularity to its surface. The long channel-like depression in the plateau may be partly of dead-ice origin and partly the result of meltwater erosion. The plateau margins on the north and south are extensively fretted by dead-ice hollows, so much so, that between some of the larger hollows, finger-like ridges descend from the plateau surface. In the south-west corner, kettle holes and dead-ice hollows have even isolated an outlier from the main body of the plateau. Eastwards, the plateau surface slopes gently and continuously as far as New Bewick, where a distinct break of slope marks its termination, but there is no ice-contact margin at this extremity. To the west, an ice-contact margin is plainly developed in places, but of even greater significance are the three prominent ridges that lead into the plateau, apparently feeding

it. The greatest expanse of the plateau surface lies just above 300 feet. Although there are no sections to reveal the nature of underlying materials, several small exposures show that it is predominantly composed of sand and water-worn gravel.

An explanation for this plateau feature is readily apparent. The numerous kettle holes and dead-ice hollows clearly indicate deposition in close association with stagnant and highly fragmented glacier ice. The deposits were evidently contained to the north and south by irregular masses of dead ice - as suggested by the crenellate outline of the plateau on these margins. Ice was also present on the west, where three esker ridges emerge from the complex system about Wooperton and feed into the plateau. Eastwards, however, the plateau surface gradually slopes to its termination in an area devoid of ice-contact slopes, an area that can most logically be interpreted as a proglacial environment. A short distance beyond this distal extremity of the plateau rises the steep scarp face of Bewick Hill and the river Till presently flows out of the Hedgeley Basin through the narrow intervening gap. This plateau of sand and gravel was evidently deposited by a system of melt-water streams issuing from either ice tunnels or open ice-walled channels, in which eskers were formed. Their outlets were into a proglacial environment in which deposition continued around and upon extensively decayed glacier ice that lay mainly to the west, north and south but was absent on the east. The eastern extremity is characterised by a break of slope that resembles the frontal slope of a delta.

On the foregoing evidence, it is suggested that the plateau-like mass of sand and gravel is an outwash delta, constructed upon and between large blocks of stagnant glacier ice. It is believed that the upper level of deposition was controlled by the surface level of a proglacial lake that was

Photograph 3.h Laminated clay and silt on the right bank of the river Till, north-west of Beanley.

Photograph 3.i Ice rafted boulder from the laminated clay and silt.

impounded within the Hedgeley Basin by the decaying front of the Tweed glacier. This level was approximately 300 feet, the height at which water could achieve outflow through the Crawley Dean channel.

(2) The Lake Floor Deposits. In addition to the remarkable outwash delta described above, there is other evidence in the Hedgeley Basin pointing to the former presence of a glacial lake.

There are numerous sections exposed by recent river erosion, along the eastern bank of the Till, in which laminated clays and silts are revealed (Photograph 3.h); similar deposits have been proved by hand-augering where exposures are lacking. An account of these will illustrate the nature of this material and the extent of the lake in which they accumulated.

At the outset, it is significant that laminated clays and silts have not been observed above 300 feet, for this evidence is in accordance with the surface level of the outwash delta. The sediments are most clearly exposed in river-side sections a few hundred yards north-west of Beanley village (along the river bank between Grid Refs. 4075/6188 and 4080/6197). The sections vary in height from 7 feet to approximately 40 feet, and although badly slumped in places, a sufficient depth of free face is exposed in which to observe the material. (The clarity of these sections tends to change with each large flood.) Comprehensive details of all the observed sections are listed in the appendix, but a general account of the clays and silts is necessary at this stage in order to support the contention that a glacial lake formerly occupied much of the Hedgeley Basin. The laminated clays and silts overlie till at the tops of the four sections in which they are exposed and are at least 10 feet thick. The clays are quite varied in colour, predominantly brown with blue/grey and red layers frequently interbedded. Micaceous silt, yellow to light-brown in colour, generally separates the clay layers and normally occurs as thicker layers. Beds of sand occur sporadically and in two sections, a layer of fine gravel is present in which the particle size ranges from $^1/10$ inch to $\frac{1}{2}$ inch.

These deposits clearly indicate:

(a) deposition of generally fine sediment in a body of still or almost still water.

(b) conditions of variable flow, during which layers of clay-size particles became interbedded with layers of silt and occasionally coarser particles.

These points strongly suggest that this deposition took place in a lake in which conditions of deposition continually fluctuated. Periods that favoured the settling out of clay-size particles alternated with periods during which silt-size particles were laid down, and these were sporadically interrupted by times when even coarser deposits were brought in. In an area dominated by landforms of fluvioglacial activity, lake floor deposits of this nature are generally regarded as evidence of a former glacial lake. This conclusion is supported by the occurrence within the laminated sediments of a large sub-angular boulder (Photograph 3.i) that was probably rafted into the lake on an iceberg.

Although the laminated clays and silts are exposed only in the sections mentioned, their presence has also been proved in an upstream direction as far as 300 feet O.D. For example, south of Low Hedgeley farm laminated blue and brown clays have been dredged up from beneath 6 feet of gravels by the machine of a sand and gravel company that is presently exploiting the gravels. Another company operates on the same gravel plain about 1,000 yards upstream and has also dredged up laminated clay from beneath the coarser deposits. The clay apparently peters out in an upstream direction where it begins to occur as strips beneath 18 feet of gravel and cobbles just before the 300-foot level is reached.

Downstream from the river sections, the present flood-plain of Till constricts considerably as the river flows through the narrow gap between the distal end of the Wooperton outwash delta and the scarp slope of Bewick Hill. A short terrace slopes gently riverward from the edge of the delta and has been undercut by the river at some time in the past; it stands about 20 feet above the flood-plain. A series of auger holes was made in the surface of this feature to determine the nature of underlying materials since no sections are available. The results are given in the appendix. In brief, underlying deposits consist of laminated clays and silts similar in colour and texture to those previously described. Immediately downstream, this terrace merges into the marginal ice-contact slopes of the outwash delta and there have been no recorded observations of laminated sediments similar in nature to those just described from any point downstream. It is therefore concluded that the lake did not extend north of New Bewick.

South of New Bewick an interesting succession of deposits was revealed by information procured from auger bores and recently dug drainage trenches along a line of section that runs roughly from the former lake floor up the frontal slope of the outwash delta on to its surface. With reference to Maps 6 and 8, the following information is briefly tabulated to illustrate the succession.

Site 6	- on the lake floor ...	3 feet of clay.
Site 7	- on lower slopes of the delta ...	6 inches of clayey soil.
		4 inches of silt and clay.
		14 inches of laminated clay.

Site 8	- well up the delta front	... 7 inches of sandy soil.
		15 inches of fine silt.
		5 inches of silt and clay.
		9 inches of laminated clay.
Site 9	- top of the delta level	... 21 inches of sand.
		12 inches of silt.
		3 inches of silt and clay.
Site 10	- on main delta surface	... 36 inches of fairly fine sand. (surface very gravelly).
Site 11	- the floor of an adjacent kettle hole	... 20 inches of yellow/brown silt.
		14 inches of light grey silt.
Trench A	- floor of meltwater channel	... 18 inches of peat.
		36 inches of grey and red silt.
Trench B	- lower slopes of channel side	... 18 inches of soliflucted debris.
		24 inches of grey and red silt.
Trench C	- centre of channel floor	... large angular blocks of sandstone dug out and piled up nearby.

The main points emerging from this information include the way in which laminated clays are overlain and interbedded with more and more silt in an upslope direction until they are completely replaced by it. The silt, in turn, gives way to sand, which becomes more widespread towards the main delta surface, where more gravel appears to litter the ground. This is the progression of deposits to be expected from lake floor level up the frontal slope of a delta and on to its surface near the distal edge. The kettle hole is floored with later river alluvium, suggesting that a block of ice occupied it while lake floor deposits were being laid down in adjacent areas, for the kettle floor lies below the level of the highest outcrops of laminated clay and silt. The large sandstone blocks indicate that bedrock may lie immediately below the delta sediments.

From all the evidence presented above, it is concluded that the surface of a glacier-dammed lake at one time extended approximately from Brandon hamlet in the Breamish valley as far as New Bewick in the Hedgeley Basin. Lying slightly above 300 feet, it controlled the level of fluvioglacial deposition that built into it an extensive outwash delta extending from Wooperton to New Bewick. At least 10 feet of laminated clays and silts accumulated on the floor of the lake, but since they do not occur north of the ice-contact slopes fringing the delta at New Bewick, it is believed that lake waters were ponded at this point by a stagnant extension of the Tweed glacier. The coincidence in height between the Wooperton delta and the crest in the floor profile of Crawley Dean strongly implies that the lake could not drain northwards at that time. For this reason, the lake surface lay between 300 and 325 feet as long as the water was forced to escape south-eastwards. The water ultimately drained away northwards when a route in that direction became available. Since there are no lower terraces fringing the Wooperton delta, it is believed that the evacuation of lake waters was relatively rapid, with no major interruptions to its progress. The present river Till probably follows the drainage route initiated at that time.

The Lower Group of Fluvioglacial Deposits

A smaller series of eskers and terraces truncates the south-easterly aligned systems of deposits and probably represents the ultimate movement of meltwaters from the north-east Cheviots. The former lie mainly below 300 feet and comprise the Lower Group of fluvioglacial deposits.

It has already been described how several systems of meltwater drainage established channels directly downslope in a north-easterly direction from about the 600-foot level, and these intersect the earlier deposits at lower levels. The channels normally terminate at about 300 feet, but extensions of the same lines of drainage can be traced to lower levels by contiguous esker systems. The latter also intersect the Higher Group of eskers. Frequently in the form of fan-like assemblages of ridges radiating from prominent meltwater channels or pre-existing valleys, the Lower Group of fluvioglacial deposits may be described by reference to the distinct systems that comprise it.

(a) The Roddam System. By far the most impressive series of formations in this group are the eskers that begin near the outlet of Roddam Dean (24, Map 6). Meltwater drainage from the Wooperton area probably joined that fluvioglacial drainage system. A prominent group of eskers fans out north-eastwards from Roddam Dean (Map 6), partly smothering, but mainly deposited in a breach cut through the earlier, north-west to south-east system. Indeed, much of their constituent material might have been derived from the redistribution of these earlier sands and gravels. The Roddam system consists predominantly of eskers, some of which are over 60 feet high, and the entire mass is pitted with kettle holes and dead-ice hollows. The system is bounded on the east by a sharp ice-contact slope, at the foot of which lies an alluvial flat presently drained by the Roddam Burn. The deeply indented nature of this eastern margin indicates that these slopes are not undercut stream banks, but that its crenellate outline is the result of contact with an irregular mass of dead ice. Beyond the main road (A 697), the western boundary of the system is aligned approximately with the minor road running north to East Lilburn. This road follows a linear dead-ice hollow which imparts a prominent ice-contact margin to the system on the west. A few short tributary eskers emerge from the outlet of the Kingston Dean channel lying to the west, and the system of massive eskers continues northwards to East Lilburn where it fades away at about 200 feet. Perhaps the most remarkable esker within the system is that adjacent to, and partly truncated by, the present Roddam Burn. It forms the most easterly component of the system and, lying between 200 and 225 feet, its crest represents the lowest level of deposition in the system. Although interrupted by gaps, it continues as a sinuous ridge for at least 2,000 yards. Along the greater part of its length it rises sharply to a narrow crest that has steep ice-contact slopes on either side. The maximum height of approximately 60 feet is reached at a bend in the esker near its northern extremity. Along the middle part of its course, the esker merges into an extensive terrace. The maximum dimensions of the latter are approximately 500 yards by 300 yards and since its surface lies at a similar elevation to the crest of the esker, it probably represents the same phase of deposition, and formed subaerially between ice-walls. On its western margin the esker is clearly connected with other ice-contact landforms that comprise the Roddam system, for it is bounded by dead-ice hollows and is partly continuous with some of the ridges. Between the esker and the river Till lies a broad belt of fluvioglacial sands and gravels that rises above 250 feet. A prominent flat-topped mass occurs at one point, but for the most part, these deposits consist of gentle undulations devoid of distinct linear trend. There are numerous shallow kettle holes and a vague north-south alignment characterises the few faint ridges that can be distinguished. In view of the approximate coincidence in elevation of the surface of this belt of deposits with the general level of the main Roddam esker system, it possibly belongs to the same phase of deposition.

A distinct pattern of fluvioglacial deposits has been traced in a north to north-east direction leading away from the outlet of Roddam Dean. It is suggested that water issuing chiefly from this channel, perhaps augmented by water from the Wooperton Dean and Kingston Dean systems (28 and 23, Maps 8 and 6), were responsible for the truncation of earlier deposits which they partly reworked to form the later group of eskers and terraces. The system as a whole declines in altitude from 300 feet near Roddam to just above 200 feet at East Lilburn where it terminates. It is also considered that the considerable expanse of low ground on either side of Roddam Burn north of Wooperton is in no way related to the present stream, for the distinct ice-contact slopes rising from it on all sides plainly indicate its origin as a dead-ice hollow. The relatively feeble erosive power of the Roddam Burn since the deglaciation of this area is demonstrated by the narrow breach that it has made through the esker.

(b) The Lilburn System. On emerging from the volcanic massif the narrow valley of the Lilburn Burn widens abruptly as it cuts through the belt of fluvioglacial deposits fringing the Cheviot foothill zone. Varying in width from 250 to 750 yards, this valley truncates the Wooler-Percy's Cross esker belt almost at right-angles to its trend and curves north-east, east and south-east to meet the Till valley at East Lilburn (Map 6). Between Lilburn Tower and Lilburn Grange the margins of the apparent flood-plain of the stream are most unusual, assuming the shape of an irregular embayment. This form is inconsistent with the more regular outlines of the valley elsewhere and with lateral migrations that are to be reasonably expected of the present stream (or even of an enlarged stream during past periods of moister climatic conditions). Furthermore, an alluvial fan built out by the small stream north-east of Lilburn Tower remains unmodified. If the full extent of this alluvial plain is a result of lateral meanderings of the present stream, then some degree of modification should be present where it impinged on the Newtown esker group. Along one short section, called Compass Bank (C.B. Map 6), this is certainly the case, and the truncated ends of eskers rise up to 80 feet above the valley floor, but the valley width at this point is approximately 450 yards, quite in keeping with the "normal" sections elsewhere. This is the only point at which the esker group has been truncated. On the eastern margin of the irregular embayment, the eskers emerge from it and appear not to have been undercut at any time. Furthermore, the embayment lies adjacent to an exceptionally large kettle hole and an area deeply pitted by similar features. For these reasons it is suggested that this peculiar enclave, apparently part of the present flood-plain of the Lilburn Burn, is more logically interpreted as a large dead-ice hollow or kettle hole that has been partially breached.

In the vicinity of Newtown the Fell Sandstone dip slope from Weetwood Moor is clothed with deposits of fluvioglacial sands and gravels. For the most part, these deposits are in the form of large eskers that rise over 70 feet high in places. They are margined and pitted by large kettle holes and dead-ice hollows, from which their ice-contact slopes rise steeply. At one point two broad ridges join to form a flat-topped plateau, from which two distributary ridges lead off north-eastwards. This esker system as a whole, is in direct alignment with the drainage down the Lilburn valley that truncated the earlier esker belt at right-angles. It fans out in a general north-east to east direction, in a similar fashion to the Roddam system emerging from Roddam Dean. The conspicuous meltwater channel aligned parallel with the Newtown esker system on its northern margin presumably belongs to the same phase of meltwater drainage, and drainage responsible for the channels, kame terraces and eskers south-east of the Coldgate/Wooler Water valley have already been described as tributary to this trend. The highest crests amongst these deposits do not exceed 275 feet, and, fading away as indistinct forms on the west side of the Till valley, the eskers do not continue below 225 feet.

There is, therefore, clear evidence that the ultimate flow of glacial meltwater in this area was in a north-easterly direction towards the Till valley. The Wooler-Percy's Cross esker belt was widely breached by this drainage system in which sands and gravels were deposited in the form of massive eskers near Newton. The upper level of deposition was conspicuously lower than that displayed by the former esker system and did not exceed 275 feet. The number and dimensions of associated kettle holes and dead-ice hollows testify to the highly decayed state of the glacier ice at this stage.

(c) The Chatton System. North of the point where the Lilburn system of deposits terminates, the Till valley is devoid of fluvioglacial landforms for about a mile. Beyond this gap they resume as a conspicuous assemblage of features, broadly orientated from south to north, around which the river bends from a north-easterly course to flow westwards through its water-gap in the Fell Sandstone ridge (Map 6); it then enters the broad lowland of Milfield plain. The account of glacial deposits in the Chatton-Wooler Basin written by Burnett (1927), does not describe the landforms in detail, but analogy was made between these "kettled gravels" and phenomena accompanying the wastage of piedmont glaciers in Alaska. Consequently, Burnett concluded, "it appears that in the melting of the dead ice-sheet the surface becomes covered with irregular spreads of dirty gravel and stones originally included in the glacier itself: these form a protective coat to the remaining ice beneath, so that in the further differential melting a hummocky surface results. ..... So the process goes on until the whole ice-sheet is melted, leaving finally our present day kettles or hollows which mark the position of the last ice cones." The one-inch drift map published by the Geological Survey for this area shows fluvioglacial sands and gravels over the broad spur of land that forms the core of the incised meander of the Till north of Chatton village. Rising over 100 feet above the river, this spur is devoid of sharp ice-contact slopes that characterise the adjacent deposits to the west. It is unlikely that the entire ridge is a massive deposit of sand and gravel in view of the size and form of all other fluvioglacial landforms in the east Cheviots. Furthermore, the only available exposure (of a superficial nature, occurring at the northern end of the ridge near its base) reveals a brown-coloured till in which limestone boulders (up to 12 inches in size) are beautifully polished and striated. It is therefore suggested that this ridge is not composed of fluvioglacial sands and gravels, but chiefly of bedrock that is covered with a thin veneer of till.

Fluvioglacial deposits with distinct topographic expression in this area lie mainly west of Chatton and may be resolved into two groups, each orientated differently. One group trends from south to north parallel with the base of the Fowberry Moor dip-slope; the other turns abruptly from the former and heads more directly downslope towards the valley bottom. However, the two groups appear closely connected and cannot readily be treated separately. The system begins as a series of conspicuous eskers up to 40 feet high, the uppermost member of which has its crest approximately at 325 feet. One esker leads upslope from 200 feet and feeds into the main group that is situated, for the most part, at about 250 feet. Numerous kettle holes, the largest of which contain water, are located in depressions bordering the ridges that splay out in a fan-shaped pattern. The front edge of this fan-shaped formation of eskers is extremely interesting. It may be traced roughly from west to east for $1\frac{1}{4}$ miles as a prominent ice-contact slope. The steep face is deeply indented by dead-ice embayments which impart a highly crenellate aspect to its outline and it rises 50 feet high in places. North of the large dead-ice hollow at S an extensive triangular-shaped terrace slopes northwards and eastwards from its apex on the western margin of the Chatton system of deposits and continues the alignment of meltwater drainage from the eskers adjacent to the south. At its distal margins the terrace merges into a series of massive ridges leading off from it radially. They rise 50 feet in some instances above broad and sometimes rather amorphous dead-ice hollows. The steep-sided meltwater channel at R partly dissects the terrace and deepens to about 50 feet before it outlets into the Till valley at between 150 and 175 feet.

There is only one clear section within the entire complex of deposits. At this point 20 feet of fluvioglacial materials consist chiefly of stratified sand, containing occasional lenses of fine gravel, overlain by irregularly bedded gravel and cobbles. Elsewhere, the nature of underlying material is revealed only by minor exposures and ploughed fields, but these are sufficient to indicate the widespread occurrence of sand and water-worn gravel composing the Chatton system of deposits.

The alignment of landforms in this system clearly illustrates that the dominant direction of meltwater drainage was generally northwards along the western side of the Till valley. The fluvioglacial deposits fan out from southern sources and decline in height towards the north and north-east. The small feeding esker that slopes uphill implies either a measure of hydrostatic pressure during its formation, and hence subglacial drainage, or else superimposition from an englacial or a supraglacial position. It is generally accepted that eskers form in tunnels within or below glaciers, and so the fluvioglacial deposition in the Chatton area was probably through or under highly decayed ice. That this glacier mass was almost certainly stagnant is demonstrated by the widespread occurrence of kettle holes, large dead-ice hollows and the prominent ice-contact slope that extends west from Chatton village. The alignment of landforms also points to a dissemination of meltwaters from courses marginal to the west side of the Till valley and out into large masses of stagnant ice occupying the basin north of the Chatton ridge or meander core. Judging from the quite massive form of many ridges, this drainage was contained by vast, tunnel-like chambers that led downslope beneath the ice. Although Gjessing (1965) suggested that extensive delta and terrace formations in southern Norway were deposited subglacially when hydrostatic pressure had lifted up the overlying ice, it seems that the glacier ice occupying the Chatton basin had become too highly crevassed and fragmented to allow the development of a similar situation, and the large kame terrace is more reasonably interpreted as a subaerial formation. It was probably deposited when the stagnant ice masses had shrunk considerably in size towards the lowest lying ground, but the presence of large eskers leading off from the terrace indicates that the ice mass was sufficiently large to accommodate vast tunnels and an extensive drainage network. Some of the larger eskers probably formed in open ice-walled canyons. A later drop in the base level that controlled fluvioglacial deposition is suggested by the meltwater channel cut through the terrace. Its outlet lies approximately at 150 feet and there are no depositional landforms below 175 feet. It may therefore be inferred that meltwaters drained freely out of the Chatton basin when the fluvioglacial water-table had dropped below 150-175 feet.

(d) The Hetton System. The Hetton valley is a broad strike vale that enters the Chatton basin from the north about $1\frac{1}{2}$ miles east of the Weetwood water gap (Map 3). On the north side of the Till valley, immediately west of the Hetton Burn, several conspicuous ridges of fluvioglacial materials lie at a similar elevation to those comprising the Chatton system, and were probably formed contemporaneously with them (Map 3). The alignment of these massive eskers indicates meltwater drainage down the Hetton valley and through stagnant glacier ice that occupied the Chatton-Hetton low ground to a height of at least 250 feet. There are no kettle holes in the Hetton area, but two prominent dead-ice hollows indicate the decayed nature of the ice. Two sections are exposed where the river Till has undercut these formations and an extremely coarse deposit of gritty sand and gravel is revealed. Many of the stones are over 18 inches in size and show evidence of having been worn by water action. The coarseness of this deposit contrasts with the calibre of material exposed in the only section available within the Chatton system, and although only tentative conclusions can be drawn, this suggests a different source of meltwater - presumably the Hetton valley.

The extent of fluvioglacial sands and gravels shown on the one-inch drift map of the Geological Survey is much greater than that mapped by the writer. The evidence on which the Survey's mapping is based is not clearly known, for few drift exposures can presently be observed with which to establish accurate limits. One striking fact is that wherever an exposure does exist, bedrock is revealed. These are shown on the Geological Survey's drift map as isolated, drift-free enclaves, yet they are the only reliable exposures of sub-soil materials. The exposed bedrock is an orange/brown sandstone that weathers to a sandy detritus, several inches of which frequently overlies the solid rock. This may have been misinterpreted by Burnett as fluvioglacial sand, for there are certainly no landforms or exposures that substantiate the extent of glacial sand and gravel appearing on the one-inch drift map.

Since the Hetton eskers swing round towards the Weetwood gap in their lower reaches, it is considered that they belong essentially to the same late phase of fluvioglacial drainage in the east Cheviot area as the previously described systems, namely, when meltwater escaped down the Till valley.

The Lower Group of Fluvioglacial Deposits - Conclusion. Considerable evidence has been presented in support of the contention that following a phase of south-easterly directed meltwater drainage, during which a great esker belt was formed predominantly above 300 feet between Wooler and Percy's Cross, an abrupt change in the drainage direction occurred. Meltwater then flowed roughly at right-angles to its former alignment, truncated and partly reworked the previous deposits and constructed separate systems of eskers and terraces that lead generally towards the Till valley. For the most part, these later systems are closely associated with major valleys leading from the Cheviot massif and with the larger meltwater channel systems orientated directly downslope towards the sub-Cheviot depression. Significantly, the eskers normally fan out from the exits of these valleys and the points at which erosional forms merge into those built up by deposition, most frequently occur approximately at 300 feet. The deposits decline in height down-valley to 200 feet in the Chatton-Hetton basin, from which meltwaters escaped beneath and through stagnant ice by way of the Weetwood gap, into the Milfield Basin.

This striking change in the direction of fluvioglacial drainage from the north-east Cheviots undoubtedly corresponds with the abandonment of the Crawley Dean meltwater channel and the draining of the Hedgeley glacial lake. This channel functioned throughout the phase of south-easterly directed drainage, during which the crest in its floor profile at any one time controlled the base level of erosion and the upper level of deposition in areas to the north. Subsequent to its abandonment, as the mass of glacier ice progressively stagnated and downwasted in the sub-Cheviot depression, free downslope drainage penetrated through and beneath the decaying ice and flowed north-east and north towards and down the Till valley. It can therefore be assumed that the Tweed glacier at this time must have reached an advanced stage of decay and fragmentation, allowing rivers of meltwater and drainage from ice-free hillslopes to flow quite freely through and beneath it to find their ultimate outlet presumably somewhere in the North Sea basin.

CHAPTER 4.

MELTWATER CHANNELS IN THE SOUTH-EAST CHEVIOTS

Introduction

The south-eastern flanks of the Cheviot massif are extensively scarred in places by glacial meltwater channels (Maps 7, 8 and 10). Although less numerous than those in the north-east Cheviots, they occasionally attain dimensions and complexities that are equally impressive. Nevertheless, they have received much less attention than their northern counterparts and although Anderson (1932) mapped and described some of them in considerable detail, Smythe (1912) referred only to a small number of them within the general framework of "forsaken watercourses" in Northumberland.

The features vary conspicuously in form and dimension, for while Northfieldhead Hill is serrated by shallow depressions, frequently no more than 8 feet deep, the channel at Fawdon is a rock-walled canyon over 100 feet in depth. The majority lie between these extremes of dimension. For the most part, the channels are cut through a superficial veneer of drift and into underlying bedrock, and whereas several appear to be incised entirely within bedrock, few have been mapped that are wholly in drift. The floors are frequently streamless and may be occupied by marshy vegetation, such as coarse grasses, reeds and peat; larger channels sometimes contain small, misfit streams. The floors and lower slopes of channel walls are normally covered and partly obscured by various deposits, including alluvial fans and other fluvial deposits, scree and solifluction formations. Sections and exposures are generally poor, but no till was observed in the channel floors at any point and none has ever been reported in such locations in the south-east Cheviots. Seldom do the meltwater channels occur as isolated features, and, like those in the north-east Cheviots, the majority are grouped in composite systems in which anastomosing patterns are common. Furthermore, the most complex and impressive networks and individual features are located in cols and valley heads.

Previous Work

The observations made by Kendall and Muff in 1901 on meltwater channels in the north-east Cheviots were not extended south of the Breamish, although these authors were aware that a considerable mass of glacier ice had encroached upon the south-eastern flanks of the Cheviot massif. A general description and interpretation of the remarkable channels south of the Breamish did not appear in the literature until Smythe reviewed the glacial geology of Northumberland in 1912. Smythe accepted without question Kendall's principles of meltwater channel formation, which had become widely established in the glacial literature of the time, and firmly believed that streams of water overflowing from lakelets ponded up in valley heads by glacier ice were responsible for the breaches through cols and over spur crests. A marginal origin was suggested for features aligned more parallel with the hillsides. Anderson (1932) discussed these channels in greater detail but made only "trifling modification" to Smythe's work and fully endorsed the latter's views. Anderson mapped the channels on a six-inch base map, however, and this enabled him to present an excellent map of the features along with his account in the Cheviot memoir. Common (1953) also recognised the presence of meltwater channels in this area during his investigation of the geomorphology of the Cheviot massif, but subsequently referred briefly to only three of them (1957). Sissons (1960) criticised Anderson's interpretation that they were cut by streams overflowing from an ice-dammed lake and stated, "the detailed evidence given by Anderson" suggests that "These small channels appear to be of marginal, submarginal and related origins formed by relatively small volumes of water locally derived." The most recent reference to meltwater channels in the south-east Cheviots was made by Derbyshire (1961), who dismissed them in the following words, "Apart from the north Lambden channel (49) only one col gully runs northwards. This is the Reaveley Hill channel (72), which together with the huge subglacial col gullies south of the Breamish marks the limit of the late-glacial dominance of the thick ice to the south of the Cheviots."

The general opinion of some previous writers that these channels demonstrate the former presence of a southern mass of glacier ice in the south-east Cheviots is undeniable, but adequate descriptions and interpretations of the complex systems are unusually absent from the literature - especially in view of the copious attention received by similar features in the north-east. Detailed mapping of these channel systems suggests that the earlier theories put forward by Smythe and Anderson are untenable and that the significance of these channels as indicators of former positions of the retreating ice margin may be questioned.

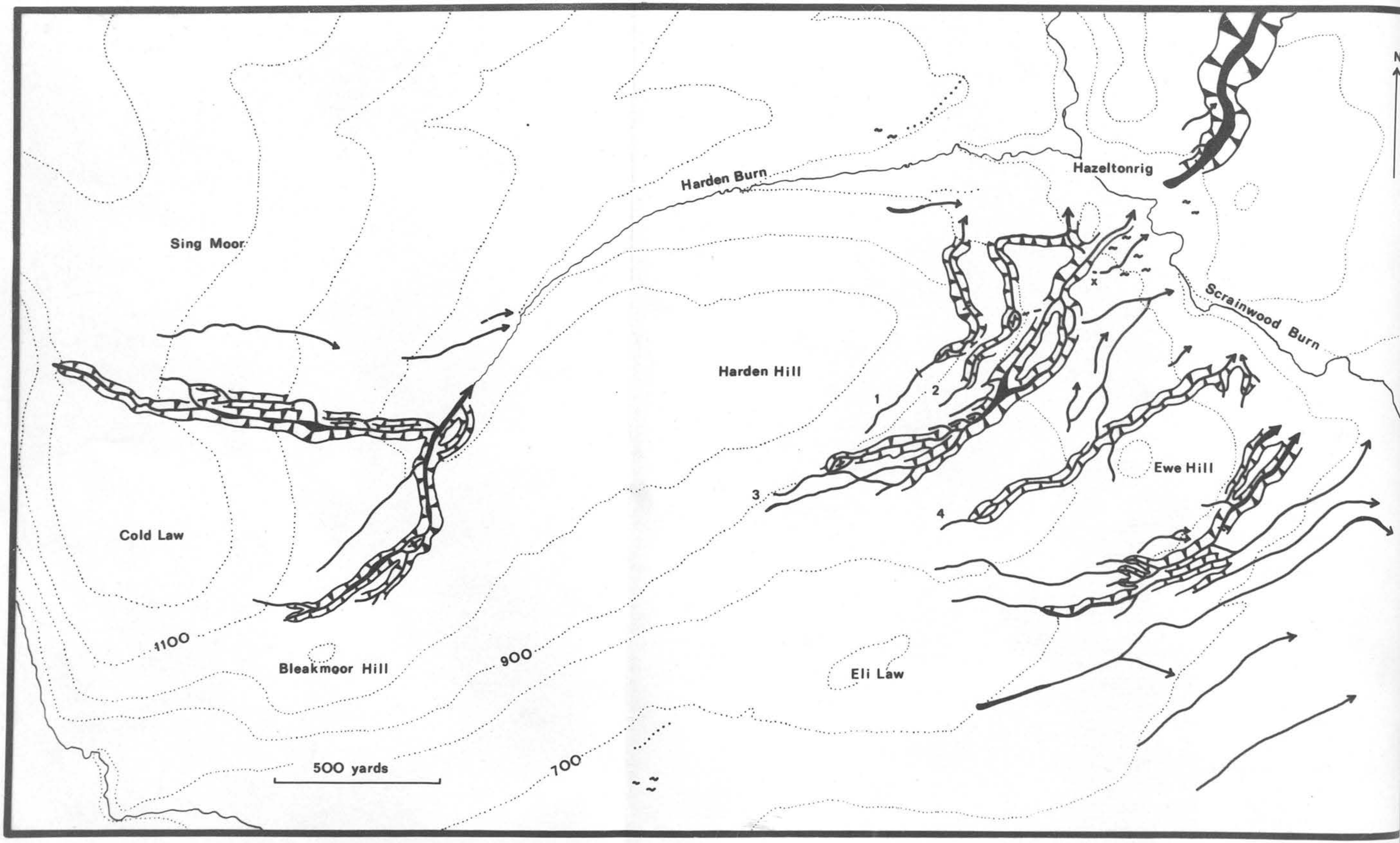
Channel Types

It was observed in Chapter 2 that meltwater channels may be classified genetically into six main categories: proglacial, marginal, open ice-walled, direct cuts, lake overflows and subglacial.

Proglacial: Since the majority of channel systems in the south-east Cheviots are aligned approximately at right angles to the present stream valleys only glacial diversion of water can satisfactorily explain their location. Consequently, they cannot have formed proglacially, beyond the ice-front, although it is possible, in view of the alignment of the Fawdon Dean channel, that some proglacial drainage flowed through it during a late and transitory stage during the downwastage of the ice. For the most part, however, many channel systems and parts of individual channels contain complexities in form that can only be explained if glacier ice existed in their immediate vicinity during their formation.

Marginal: Whenever meltwater channels are observed to trend parallel with or at small angle to the contours of a slope, it is necessary to consider the possibility that they were formed by subaerial drainage along the ice margin. Indeed, such an origin was considered by both Smythe (1912) and Anderson (1932) for the intricate system on the slopes of Harden Hill (Map 10), and the latter writer commented that "a certain amount of oscillation" of the ice edge was "expressed by an anastomosing series of shallow feeders on the east face of Harden Hill". Since it has already been indicated (Chapter 2) how Sissons (1960) has indicated the difficulty of interpreting any meltwater channel in Britain as marginal, it will be sufficient at this point to demonstrate that the views of Smythe and Anderson for the Harden Hill channels, and others that they considered marginal in origin, are untenable.

The Harden Hill Channel System (51, Map 10): The ground between Bleakmoor Hill and Harden Hill is in the form of a broad, rather level-topped ridge, the crest of which lies mostly above 1,025 feet and is aligned from south-west to north-east. Steep slopes descend on either side of the ridge, but a pronounced break of slope interrupts the continuity of the eastern flank and a flattish bench extends approximately 400 yards in width before the edge of the volcanic massif again descends quite steeply to the more gentle topography underlain by the fringing Cementstone rocks. The channel system that furrows the eastern flank of Harden Hill, and its spur-like extremity to the north, is perhaps the most intricate of the east Cheviot area. Four major channels,



N
Sing Moor
Harden Burn
Hazeltonrig
Scrainwood Burn
Harden Hill
Cold Law
Ewe Hill
Bleakmoor Hill
Eli Law
1100
900
700
500 yards
1
2
3
4
x

Figure 4.1

together with many tributaries, comprise this system and for ease of description, they have been numbered 1 to 4 (Figure 4.1).

Channel 1 begins as a narrow bench on the eastern side of Harden Hill and slopes gently uphill at a small angle to the contours for the first 240 yards of its course. A few yards beyond the crest in the floor profile, which lies no more than 10 to 15 feet above intake level, the channel becomes contained between two rock walls and curves obliquely downhill for about 40 yards. Following a sharply incised chute section only 30 yards long, it resumes a course almost parallel with the contours for a further 80 yards. Along the latter section, its downslope wall is absent, except for one small rock outcrop. Instead of continuing this course along and gradually down the eastern flank of Harden Hill, channel 1 bends abruptly at right-angles to its original alignment and climbs steeply uphill over the spur crest in a most unusual manner. A second crest in its floor profile occurs 20 feet above its level at the bend, and the channel depth at this point, on the spur crest, is 15 feet. Beyond the spur crest, channel 1 slopes obliquely (almost at 45 degrees) down the western side of Harden Hill, but before terminating, it turns sharply from this alignment and plunges at right-angles across the contours of the slope as a prominent chute. The channel finally terminates at 750 feet beside a ridge of sand and gravel, lying above the present Harden Burn.

Channel 2 is no less remarkable and displays similar characteristics. Beginning as a bench on the eastern side of Harden Hill, within 100 yards, it becomes contained between rock walls, then curves obliquely uphill until its floor is 20 feet above intake level. The channel is 15 feet deep at this point. Immediately beyond this crest in its floor profile, channel 2 begins to curve gently back downslope to resume a course almost parallel with the contours and, like channel 1, it lacks a wall on the downslope side along this section. Similar to channel 1, channel 2 also turns abruptly at right-angles from this alignment and slices through the spur crest, but this is accomplished without the development of a second uphill gradient. Almost precisely at the same part of the hillside where channel 1 plunges as a chute, channel 2 bends through 120 degrees and slopes steeply and obliquely as a one-sided feature down the western flank of Harden Hill. On reaching the 700-foot level, it turns sharply through 90 degrees and assumes a course more parallel with the hillside for 130 yards before terminating at 675 feet on the bank-top above the present Scrainwood Burn.

Channels 1 and 2 are incised entirely in bedrock, apart from the last few yards at their outlet ends which may be cut partly through fluvio-glacial deposits.

Channel 3 is a much more composite feature than the previous channels and two major intakes unite to form the main feature. The intakes themselves are rather complicated, for loops and dividing-uniting patterns are quite characteristic. Along the last 60 yards of its length, the inter-channel divide progressively narrows until it is in the form of a knife-edge ridge of bedrock. Precisely at this confluence point, further complications appear on the left side of channel 3. These take the form of small, but distinct, chute-like features furrowing the channel wall so that inter-chute divides protrude as rock bastions. Totalling 5 in number, the first three of them appear to represent off-shoots from a shallow, high-level channel that runs parallel with the main channel 20 feet above it. The fourth occurs where a shallow tributary plunges into the main channel, and the fifth is a short chute that enters at the point where channel 3 is over 50 feet deep.

From the confluence of its two intake branches, the floor of channel 3 progressively broadens until it becomes 45 yards across at the point where it splits into two sections that diverge round a broad ridge of bedrock, over 50 feet high. The narrow left branch is asymmetric in cross profile, the higher, upslope wall rising 70 feet above the floor. The floor of the wider, shallower right branch seems to lie at a slightly higher level than that of the previous channel. An added complexity within this right branch is the low ridge, around which the floor bifurcates and re-unites before its confluence with the left branch. Channel 3 continues beyond this point with a steep gradient as it slopes down the northern end of the Harden Hill spur. At the 700-foot level, an earlier extension of channel 3 turns sharply left through 90 degrees to join a similarly-aligned section of channel 2, but there is evidence that this connecting route was abandoned while the main route continued to function, because the former hangs several feet above the main outlet. The latter continues directly downslope towards the Scrainwood Burn, above which it terminates at 650 feet. This final outlet from channel 3 appears to have been cut through drift, possibly sand and gravel, and the shallow depression leading off at x may represent another outlet that functioned for only a short period.

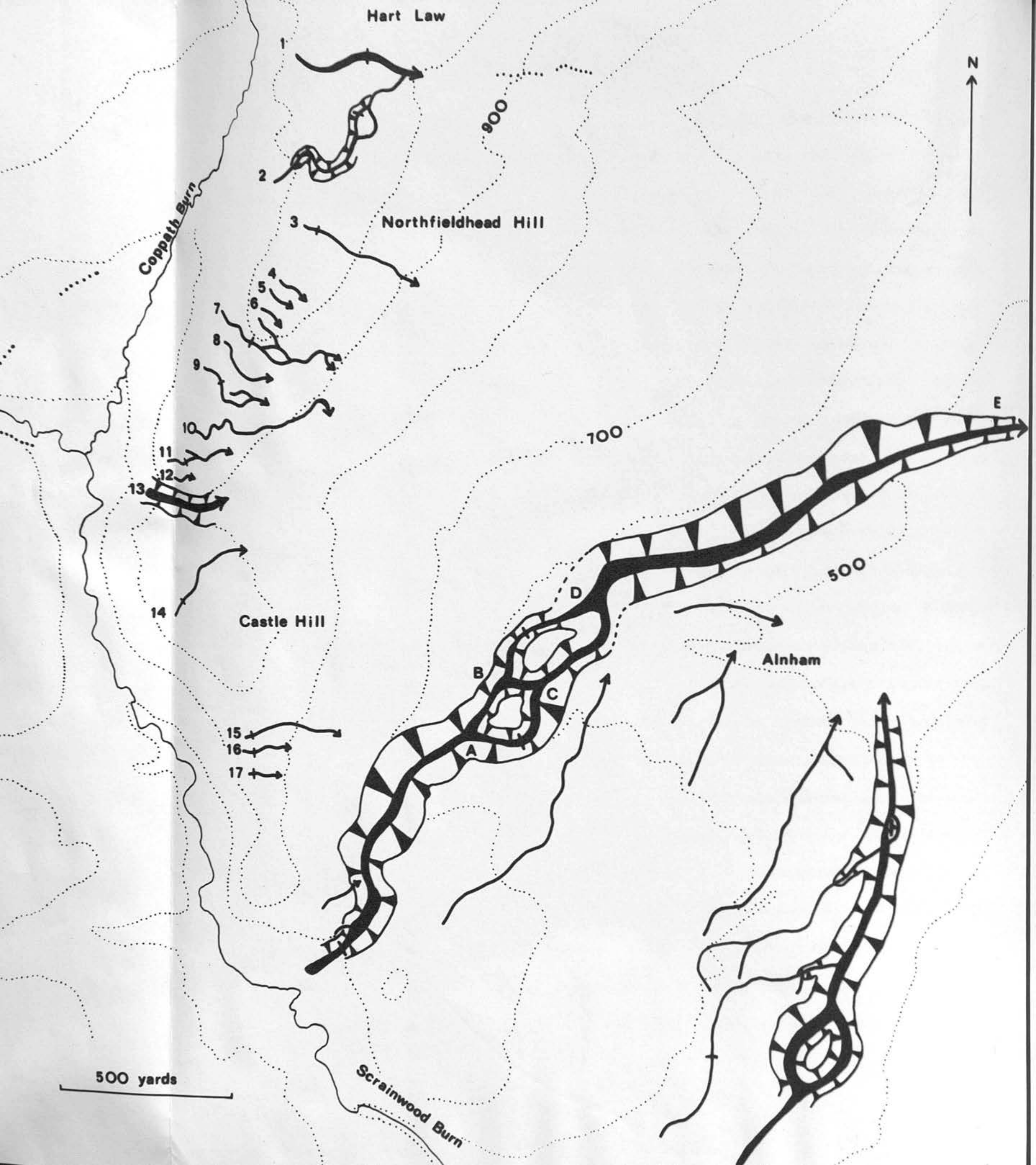
Channel 4 is the shallowest of the four channels and attains a maximum depth of only 12 feet, but is, nevertheless, well-defined throughout its length. Beginning at 825 feet on the bench feature below Harden Hill, it runs as a shallow grassy depression for 90 yards before more conspicuous walls rise above its floor. The course assumed by channel 4 is relatively straight and uniform as it leaves the bench surface and is aligned directly downslope towards the Scrainwood valley, but there are numerous associated channels adding to the complexity of the arrangement. For example, at one point the left wall of channel 4 is breached by a shallow distributary that leads off obliquely downslope and connects with another small channel beginning independently on the hillside. This feature ultimately joins what appears to have been a former distributary of channel 3 before terminating at 650 feet. A short tributary, issuing from the shallow col west of Ewe Hill, enters channel 4 on the right, but no further complexities appear until near the outlet, where two deep chutes enter on the right wall. Channel 4 terminates at 625 feet. Two other small channels occur on the hillside between channels 3 and 4.

Previous work concerning these channels is as follows. Channels 1 to 4 appear in simplified form of necessity on Smythe's rather small-scale map of such features in north Northumberland, but if he was aware of the numerous complexities just described, he made no reference to them in his text. Describing "the great series ..... on Harden and Ewe Hills", he said, "Many of the last are marginal trenches, and indicate that the ice-edge lay parallel to the direction of the Harden Hill ridge." Channel 1 (E 10 on his map) attracted his attention more than the others, for he continued, "E 10 is marginal in the western part of its course, then it crosses the water-parting, making a very conspicuous gap in the sky-line, and falls abruptly on the north side of the hill (Photograph 7)."

Anderson mapped the channels in considerably more detail but treated them rather lightly in his accompanying discussion. Commenting on Smythe's observations, Anderson admitted that "During the recent revision only trifling modification of his work has been needed", and it is quite apparent that Smythe's interpretation of the channels was accepted. The work of previous investigations in this area therefore led to the conclusion that the meltwater channels on Harden Hill were formed marginal to glacier ice lying against the eastern flank of that hill, and that they fed into a proposed ice-dammed lake in the Spartly valley.

In addition to the doubtful validity of interpreting most meltwater channels by the marginal hypothesis, as outlined by Sissons (1960), there is other evidence which also refutes such an interpretation for those on Harden Hill. The distinct uphill sections present in channels 1 and 2 cannot be satisfactorily explained by the subaerial flow of water marginal to the ice, and are more reasonably interpreted as subglacial phenomena. Furthermore, the oblique and chute-like sections of the same two channels clearly do not represent the former slope of an ice margin, for the gradients are much too steep. The numerous complexities in channel 3 are unlikely to have been eroded by marginal drainage, since it is difficult to conceive that the glacier margin could have oscillated in a manner such as to produce this complicated network of interconnecting channels. Similar arguments apply to channel 4, which descends 200 feet within 1,000 yards and is associated with numerous connected branches. The orthodox marginal interpretation does not, therefore, satisfactorily account for the various characteristics observed in channels 1 to 4. The one-sided sections of 1, 2 and 3 indicate that glacier ice must have been present on the downslope side of these channels, but the uphill gradients along certain sections of the floors suggest that the meltwater streams were flowing under considerable hydrostatic pressure and must have been contained between two walls of ice in subglacial tunnels.

The Ewe Hill Channel System (52, Map 10): The slope below the bench-like facet (on the eastern flank of Harden Hill) of which Eli Law and Ewe Hill are partially isolated remnants, is furrowed by numerous meltwater channels (Figure 4.1). Although short sections of these features trend almost parallel with the hillside, much longer sections of them are aligned predominantly at an oblique angle to the slope. Short tributaries and interconnecting branches are commonly developed, the continual abandonment of higher routes for those at lower levels being particularly striking. The form and location of this



Hart Law
N
900
Northfieldhead Hill
Coppath Burn
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
700
E
500
D
Castle Hill
B
C
A
Alnham
500 yards
Scrainwood Burn

system are similar to those of features deduced as submarginal in origin by Sissons (1961a). They are unlikely to have formed marginally to the ice edge as suggested by Smythe and Anderson since their relatively steep gradients crossing the contours obliquely are excessive for an ice margin and their very complexity similarly precludes this interpretation.

The Northfieldhead Hill System (47, Map 7): The striking spur of land forming the eastern watershed of the Coppath and Spartly Burns extends from Hart Law in a gentle arc that curves from a south-south-west to a south-south-east direction. Over its length of approximately two miles, the spur descends from 1,100 feet to 825 feet above Hazeltonrig and although its crest declines quite uniformly for the most part, a pronounced col separates the main Northfieldhead ridge from a lower extension, called Castle Hill. No fewer than 17 meltwater channels were mapped on the crest of this spur. The majority are quite small features varying in depth from 5 to 10 feet, but one exceeds 15 feet and that in the col is almost 40 feet deep. Cut entirely in bedrock, their length is normally between 75 yards and 500 yards and only four of them extend far beyond the spur crest and down the eastern flank of Northfieldhead Hill. A brief description of several individual channels is considered relevant to the thesis since many possess characteristics of form that aid an interpretation of their genesis. From north to south, the channels have been numbered 1 to 17 (Figure 4.2).

Channel 1 begins west of the spur crest at about 1,025 feet and slopes gradually uphill along the first 250 yards of its course as a broad, shallow depression. The downhill section to its outlet at 1,000 feet is only 140 yards long.

Channel 2 is perhaps the most intriguing feature on the spur, and joins channel 1 immediately before the latter's outlet. If the water which eroded this channel flowed in the same direction as the regional meltwater drainage (observed from all other channels in the area), then the feature climbs over 50 feet from its intake on the western flank of Northfieldhead Hill to its crest on the spur summit. Reaching a maximal depth of 15 feet, it is a steep-sided feature that describes a sinuous course with undercut bends, up the hillside. Before the crest is reached, a meander loop branches off to the right, and rejoins the channel beyond the summit. The downslope section of channel 2 is only 150 yards long and is much more poorly developed than the uphill section, for it lacks conspicuous walls.

Channel 3 has a short uphill section, rising only a few feet before cutting across the spur crest to plunge steeply down the eastern flank of Northfieldhead Hill at right-angles to the contours, terminating at approximately 880 feet.

Channels 4, 5 and 6 are all short and only a few feet deep, 4 and 5 being one-sided features.

Channel 7 is a more complex channel that climbs uphill from its intake with a gently sinuous course for a distance of 150 yards. The crest in the floor profile is about 15 feet above intake level and the channel has become 10 feet deep by this point. Beyond the spur crest, the channel bifurcates and a short tributary slopes obliquely down from the ridge summit to join the northern loop which then rejoins the southern. As the channel begins to descend the eastern side of the spur, it turns abruptly, almost parallel with the hillside for several yards before again plunging steeply downslope to its outlet at 900 feet.

Channel 8 begins west of the spur crest but avoids developing an uphill section by curving round the sloping summit on a level course 8 feet deep.

Channel 9 also intakes west of the spur and curves upwards for 10 feet along the first 75 yards. It becomes 15 feet deep between steep walls, before sloping down towards the east. A tributary branch, the intake of which begins on the spur summit 8 feet above the main channel, joins accordantly before the latter's outlet.

Channel 10 describes a most peculiar course, meandering sinuously across the spur crest with a double bend and then sloping almost parallel with the contours on the eastern side of the hill. Its upslope wall rises over 25 feet in height prior to the final sharp bend that occurs as the channel plunges directly down the hillside to terminate at 875 feet.

Channel 11 begins as a bench and climbs several feet uphill for about 50 yards. A short tributary enters on the left beyond the spur crest and the entire channel describes a gently winding course.

Channel 12 is a short feature similar to 4, 5 and 6.

Channel 13 is the most conspicuous feature on the spur, cutting through the col between Northfieldhead Hill and Castle Hill as a steep-sided, rock-cut gorge almost 40 feet deep. It is, however, quite short and within 260 yards of its intake, fades away only several feet lower in level.

Channel 14 is a small feature that begins with an uphill profile and then winds across the summit of Castle Hill to terminate at 900 feet on its western side.

Channels 15, 16 and 17 are similar features situated quite closely to one another across the narrowest portion of the entire spur. All exhibit short uphill intake sections towards the spur crest.

The first reference to any of these channels was that made by Smythe, and this was to only three of them. The glaciological situation he envisaged in this area was described in the following words, "The edge of the western ice sheet ..... lay parallel to the ridges and across the Spartly and Biddlestone Burns, the result being the production of a great number of marginal trenches along the ridge sides and direct cuts across the water-partings." The first one he mentioned is channel 2; this he considered "is the only one of the series which drains westward". The other two to which he referred appear from his map to be 11 and 13. To explain these he envisaged a slight recession of the ice margin so that "The upper part of the Spartly valley was now free from ice, and, being dammed lower down, overflowed by Coppath, cutting the swires E4 and E5" (channels 11 and 13). Anderson, using a larger-scale base map plotted 15 of these channels but rather surprisingly did not refer to the anomalous uphill profiles and tributary sections. Had he contemplated the significance of such anomalies, he might have reconsidered his theory of a "lake being drained by cuts over the long ridge east of the Spartly Burn". Anderson explained the channels not only as lake overflows, however, for he continued, "With one exception ..... these are small, no more than 10 feet deep at most, and they are a striking witness to the steady retreat of the ice-margin." Anderson's interpretation of these channels shows several inconsistencies. He clearly envisaged an ice-dammed lake in the Spartly embayment, the margins of which at first lay "about the 1,000-foot contour line", and as the ice surface downwasted the surface level of the lake dropped in harmony, so that each successive channel down the crest of Northfieldhead spur represented a temporary lake-overflow. In the first instance, the highest channel of the series lies at 1,025 feet, implying that the initial level of the lake postulated by Anderson must have lain at this height and not at 1,000 feet. For water to have been ponded at this level, the ice barrier must have lain at an equal elevation. Furthermore, if the overflowing water was guided marginally across the spur at 1,025 feet, then glacier ice must have been present on the spur at that point. If the latter situation developed, then glacier ice must also have filled the Spartly embayment to at least the 1,025-foot level. This being so, there would have been an extremely small area of ice-free hillside against which Anderson's "considerable lake" could have formed. Even allowing that a lake did form, the up and down long floor profile of channel 1 cannot be explained by subaerial drainage from such a lake. Furthermore, channel 2 is clearly confluent with channel 1, which it joins accordantly, and was most probably formed at the same time. Since it lies at a lower level on the spur, Anderson's interpretation of this channel as a lake overflow is untenable, for a lake cannot be drained at two different levels in close juxtaposition on the same spur. This criticism is strengthened by the pronounced uphill gradient of channel 2 throughout the greater part of its length. Eight of the remaining channels possess uphill gradients along their floors and the sinuous course of channel 10 also tends to preclude a marginal-lake overflow interpretation. Phenomena that are frequently associated with the former presence of an ice-dammed lake, such as shorelines, floor deposits and deltas, are absent in this area and it is thus unlikely that a lake was ever ponded in the Spartly embayment for a duration sufficient in length to allow the formation of distinct landforms. Since eleven of the Northfieldhead channels have up/down floor profiles and are most satisfactorily explained by the subglacial hypothesis, then it is reasonable to similarly interpret the remainder, for they lie in close juxtaposition on the same spur.

Applying similar arguments to other channels that Smythe and Anderson explained as lake overflows or marginal features, it can be concluded that clear evidence to support the contention that meltwater channels in the south-east Cheviots were formed in positions marginal to glacier ice, is distinctly lacking.

Lake Overflows: Although the meltwater channels described by Smythe and Anderson as lake overflows have been discounted as such and interpreted differently, the possibility that other channels in the south-east Cheviots might have formed in such a manner must be considered, for glacier-dammed lakes commonly occur in some areas presently glaciated. Before meltwater channels in regions now ice-free can be thought of as former lake overflows, certain requirements pertaining to these features ought to be met. For example, (1) the channel should be located in the lowest part of a col; (2) there should normally be only one channel in the col; (3) the longitudinal floor profile should fall continuously (except, perhaps, for minor irregularities) from intake to outlet, unless subsequently modified by extensive infill of scree and other slope debris; (4) the channel should not exhibit anomalies such as abandoned loops, tributary networks and isolated rock knolls in the floor and on the walls; (5) the channel should not contain ice-contact deposits or till if it is considered to have formed entirely in an extra-glacial environment. In the south-east Cheviots, there are no channels that satisfy all of these requirements, and only one feature exhibits characteristics that justify discussion of it as a possible lake overflow channel.

The Pigdon Channel (46, Map 7): The northern rim of the Spartly embayment is broken by a conspicuous col between Hogdon Law (1,797 feet) on the west and High Knowe (1,325 feet) on the east. Steep slopes sweep smoothly down from these hills to the peat-covered floor of the col, approximately 250 yards across. Two small valleys emerge from the morass and lead towards the Spartly Burn. The western valley is wide and shallow and appears to be a pre-existing feature, whereas the eastern is a deep, rock-bound canyon, largely the work of meltwater erosion and presently occupied by the streamlet called Pigdon's Sike. Although the channel begins on the col crest as a marshy depression only a foot or two deep, its actual depth may be greater, for the considerable accumulation of peat has probably obscured much of the original topography. Within 80 yards, the channel deepens to 25 feet as it becomes a narrow gorge with precipitous walls. At one point the channel widens slightly to accommodate complexities in the floor where bifurcation has occurred. A narrow ridge of rock 8 feet high forms the intervening divide, through which a short breach connects the main channel route on the right with the subsidiary branch on the left. A narrower gorge section continues from the confluence of these two segments as far as the outlet, where the channel again widens on account of an abandoned loop cut in the left wall 20 feet above the main floor; the loop is 30 feet deep. Signs of meltwater erosion terminate quite suddenly here, as the present stream plunges out of the channel as a waterfall and flows towards the Spartly Burn between shallow banks. Beginning in the col at 1,100 feet, the Pigdon channel terminates at 975 feet. Although the channel appears to begin at a point that is not the lowest position in the col, this can be taken only as tentative evidence against the lake overflow theory since peat development may well account for it. The complexities in form, however, are perhaps more difficult to account for by subaerial erosion, unless the presence of an adjacent intrusive dike of quartz-porphyry influenced channel development. Perhaps the most serious objection to the interpretation of this feature as a lake overflow channel is the complete absence of any other phenomena that might reasonably be expected to have formed in association with an ice-dammed lake sufficiently large and permanent to account for the Pigdon channel. Had a lake formed in the broad embayment north of the col, then some evidence of deltaic sediments ought to exist where numerous tributary valleys enter from the surrounding higher ground; such deposits are totally lacking.

There is, therefore, no conclusive evidence that former glacier-dammed lakes were responsible for the formation of any meltwater channels in the south-east Cheviots. Indeed, it may even be suggested that glaciological conditions pertaining in the Cheviot massif during the last glaciation did not allow the formation of ice-dammed lakes sufficiently permanent to leave records of their former presence.

The most significant aspect of the Pigdon channel is the way in which it demonstrates an alignment of meltwater drainage from north-west to south-east in this particular part of the south-east Cheviots. For this to have occurred, the surface of glacier ice in the Shank Burn embayment must have reached at least 1,100 feet and stood higher than ice occupying the Spartly embayment.

<u>Open Ice-Walled Channels</u>: Because of the considerable altitudinal range from intake to outlet (frequently over 300 feet) in a great number of meltwater channels in the north-east Cheviots, it was considered that such channels had not formed in open ice-walled courses. Indeed, it is very unlikely that open walls of ice 300 feet deep can ever develop (Glen 1954). Furthermore, the presence of isolated rock knolls and complex networks of narrow, inter-channel divides, was interpreted as additional evidence against the likelihood that the rivers of meltwater responsible for their formation flowed between highly fragmented open walls of ice. Such objections to this mode of channel development are similarly relevant in the south-east Cheviots, where only two systems possess characteristics of form that do not immediately preclude the open ice-walled theory for their formation.

The Biddlestone System (54, Map 10): This group of interconnected meltwater channels occupies the slopes and floor of the shallow through valley leading from the River Alwin valley to that of the Netherton Burn. The broad curve of this depression, at between 800 and 700 feet, approximately coincides with the edge of the Cementstone rocks overlapping unconformably upon the faulted margin of the volcanic massif. While parts of the system may be cut in bedrock, many of the shallower segments are incised predominantly in drift, and at no point is a depth of 25 feet exceeded. Outstanding characteristics of the system include the multiplicity of interconnecting branches, the considerable sinuousity displayed along certain sections and the consistent alignment towards the lowest route in the valley, where the greatest volume of water appears to have become concentrated. The only reference to this system in previous literature is that made by Anderson (1932), when he observed that, near Biddlestone, "Anastomosing channels ..... bear witness to oscillation of the ice margin .....", but there is such a dense network of features that this explanation is most improbable. The greatest range in altitude shown by channels in the system is slightly over 200 feet, but 125 feet is commonly exceeded. It is not improbable that open walls of ice could have been maintained with these depths, but perhaps more serious objections may be raised if it is implied that a network of deep crevasses, some of which wound sinuously downslope could ever have existed in such concentration. Although there is no conclusive evidence, it appears more reasonable to suggest that the greater part of the Biddlestone channel system was formed entirely beneath the ice in the submarginal zone.

The Trewhitt System (56, Map 10): Numerous meltwater channels occur in the vicinity of Trewhitt Moor, one to two miles east of Coquet Dale. The five channels entrenched on the face of the broad spur, forming the watershed between the Trewhitt and Foxton Burns, comprise part of the system. They are mostly less that 15 feet deep, but wind down the spur as distinctive dry gullies. The deepest components of the system are located in the floor of the Trewhitt valley head, where meltwater streams cut channels up to 25 feet in depth. The whole system could have been cut entirely by rivers of meltwater flowing between open walls of ice, for their characteristics of form and location present no obvious objection to this theory, but they might equally have been eroded entirely by subglacial drainage.

Direct Cuts: Previous literature on glacial meltwater channels in the east Cheviot area occasionally refers to "direct cuts", a concept of channel formation that must be considered for the south-east Cheviots. Conclusions reached on the subject of "direct cuts" in Chapter 2, however, apply with equal relevance to this area, and on that basis, the majority of channels in the south-east Cheviots may be excluded from such a potential mode of origin. Perhaps the most notable exception is Fawdon Dean, which possibly functioned as a "direct cut" at some stage during its genesis. Further discussion of this channel will illustrate the possibility.

Fawdon Dean (44, Map 8): This enormous channel is entrenched in the floor of a deep pre-existing valley, located precisely where a great fault has thrown a variety of rock types into close juxtaposition, including Cementstone, andesite, ash and agglomerate and minor inliers of Silurian material. Active denudation along this line of geological weakness appears to have opened up a high-level through-valley across the Aln/Breamish watershed, into which the Fawdon Dean channel was subsequently incised by meltwater draining north-eastwards. Precipitous walls, covered throughout by various slope debris and vegetation, rise to a maximum height of over 100 feet above a flat marshy floor that is obscured by peat and reeds; a very small stream eventually emerges from this tangle of morass and presently flows through an artificial ditch to join the Breamish. Preliminary signs of meltwater erosion associated with Fawdon Dean, which is cut almost entirely in bedrock, appear at 825 feet, and the last vestige fades away at approximately 425 feet. The main channel begins after a complex system of seven short feeders have united. Four of these intake segments are cut into the lower slopes of Cochrane Pike and have resulted in a remarkable series of pot-hole-like features, called the Bowl Holes on the 6-inch and $2\frac{1}{2}$-inch maps of the Ordnance Survey. The four chutes descend almost 25 feet into a central plunge-pool, which is thus encircled by smooth-sided re-entrants that resemble four half bowls. A short distance downslope, meltwaters from the Bowl Holes were joined by another short tributary, and their combined volume cut a one-sided channel that curves sharply round into the main gorge. Other small tributaries, between which rise sharp-crested ridges of bedrock, are located in the broad valley leading gently uphill from the south-west towards Fawdon Dean. The entire network of intakes focusses on the col crest, and beyond, vast undercut bends lead into the main channel. The impressive form of the channel is maintained quite uniformly for over a mile, until it curves sharply round a flat-topped spur and emerges from the narrow confines of the through-valley into a somewhat wider valley-head. Almost immediately the channel form changes; the walls lose their definition and although the left side exhibits some degree of oversteepening by meltwater, the opposite side is simply the flank of the Fawdon spur descending uniformly and steeply, bearing no evidence of fluvioglacial modification, to the flat floor of the channel. There seems to be little doubt, however, that torrents of meltwater sweeping down Fawdon Dean were chiefly responsible for the broad valley-within-valley form of lower Fawdon Dean. The intricate network of feeders strongly suggests that glacier ice was present to at least the 825-foot level when meltwaters first began to discharge through the Fawdon valley. During this stage in the downwastage, it is quite possible that the thicker ice masses in the Aln and Breamish valleys were still connected by a remnant limb of glacier ice occupying the Fawdon valley; Cochrane Pike and Fawdon Hill presumably rose as nunataks at this time. For these reasons, it seems quite likely that Fawdon Dean owes its inception to subglacial meltwaters. However, as the general ice cover progressively downwasted, there must have been a time when remnant ice within the narrow confines of the Fawdon valley became either so thin and fragmented that it quickly disintegrated, or else wasted down towards the Aln valley as a distinct glacier tongue. In either case, it is apparent that considerable volumes of meltwater would have been able to discharge directly from the margin of the major ice mass in the Aln valley by way of the convenient route available through the col into the Breamish ice mass. The consideration that this drainage through Fawdon Dean was subaerial would avoid the necessity of a vast subglacial tunnel, sufficiently permanent to allow over 100 feet of incision beneath it. Yet it must be conceded that erosion was perhaps effected quite rapidly along this line of weakened bedrock, and since many examples of even larger subglacial channels exist in Britain and elsewhere in Europe the subglacial concept must be considered as a possible alternative with which to explain the Fawdon channel.

Concerning the evidence for channel types other than subglacial, in the south-east Cheviots, it may be concluded that the Trewhitt system might have been formed chiefly in open ice-walled crevasses, and that much of Fawdon Dean was possibly cut subaerially. Alternatively, these examples could equally well be explained as subglacial forms.

<u>Subglacial</u>: The various characteristics of form that convincingly point to a subglacial origin for meltwater channels were exhaustively discussed in Chapter 2, and there is no need to repeat the arguments here. In brief, however, it may again be emphasised that channels possessing true up/down longitudinal floor profiles are generally considered to have been formed subglacially by meltwater flowing under hydrostatic pressure. Some of the channels and channel

systems falling into this category in the south-east Cheviots have already been mentioned, and in addition, the Scrainwood system and the Silverton Hill channels (49 and 53, Maps 7 and 10) may also be included. While the majority of channel systems in the area can therefore be accounted for in this way, there remain three groups in which other inherent characteristics point strongly to a subglacial origin.

The Leafield System (42, Map 7): The main artery of this system is a rock-cut gorge 30 feet in depth, significantly called Dry Dean. It begins as a flat, marshy-floored channel leading into the hillside high up on the slopes of a small valley tributary to the Cobden Burn (a tributary of the Breamish), and within 50 yards, it turns abruptly at right-angles to slice through a col in the Leafield Edge ridge. The initial 130 yards have an uphill gradient and climb over 10 feet above intake level, but this relatively small crest in profile could be a result of subsequent modification by infill. A remarkably straight course is maintained through the relatively narrow confines of the pre-existing col and three small chute-like tributaries gash its right wall. Immediately beyond the ridge, where the col broadens out on to the eastern flanks of Leafield Edge, complexities in channel form begin. Benches and loops, apparently abandoned at higher levels above the main channel floor, isolate rocky knolls on both sides. Farther down a second major channel appears; parallel with and tributary to Dry Dean, it lies about 10 feet higher up. The narrow intervening divide is breached at two points by channel segments that seem to have originally connected Dry Dean with its tributary. Their intake ends hang 8 feet above Dry Dean and must have been forsaken some time before the latter ceased to function as a meltwater channel. Lower down two minor furrows barely three feet deep nick the interfluve in a similar manner. The lower section of the present Leafield Burn valley appears to have been largely cut by meltwater and possibly connects with loop x. A system of interconnecting channels thus appears to have at one time functioned on the gentle slopes of this pre-existing valley. Eventually, the Dry Dean branch assumed dominance with the consequent abandonment of other branches. The various complexities of this system can only be accounted for if glacier ice was present in their immediate vicinity. It is unlikely that meltwater drainage was subaerial, for this would entail narrow wedges of ice perched precariously on inter-channel divides and rising 100 to 200 feet high, (the system descends from 950 feet to 700 feet). Furthermore, the three right-wall tributaries are features distinctly similar to subglacial chutes described by Mannerfelt (1945), Sissons (1961a) and others. For these reasons, it is considered unlikely that this braided system of meltwater channels was formed in any manner other than subglacial.

The Middle Dean System (43, Map 8): The broad col that almost isolated the Ingram spur from Cochrane Pike was probably lowered on account of headward development by the two wide valley heads on either side. Within this col lies the Middle Dean channel, descending north-eastwards until it is firmly entrenched in the floor of the pre-existing valley tributary to the Breamish at Ingram. The diminutive ribbon of water called the Middledean Burn does not appear until almost half-way down the length of Middle Dean, so that the upper reaches are either dry or covered with marshy vegetation and slope debris. Although intake level lies at 850 feet, it is clear from the upper rims of the channel walls that meltwater incision began at approximately 900 feet. From an initial width of 30 yards, the peat-floored intake soon develops into a narrower, steep-sided canyon that maintains this form, 70 feet deep in bedrock, through the col. Previous workers in the Cheviots commented upon the peculiar irregularities in the floor profile along this section of the channel. Four broad, basin-like depressions, of which two are rather marshy (water-filled following prolonged wet weather) and the other two dry, alternate with narrow connecting passages that impart a somewhat beaded character to this part of the channel. Smythe (1912) explained these pool-like phenomena by scree damming, but Common considered that they "..... are all separated by low rock bars". It is most likely that a combination of an original irregular floor profile and scree debris accounts for the features. The crest of one inter-pool divide lies distinctly above the intake level of Middle Dean, and it is quite possible that the channel possesses an original up/down longitudinal floor profile of sufficient relief to require a subglacial explanation. But since there is some doubt concerning the quantity of subsequent in-fill, additional evidence in support of such an interpretation must be presented. In this connection, reference is made not only to three abandoned loops isolating small rock knolls, but also to five gullies cut into the right wall of Middle Dean. The latter are mostly streamless, less than 10 feet in depth, and terminate up to 15 feet above the floor of the channel. Furthermore, the absence of debris cones below their outlets contrasts with the pronounced alluvial fan built out by Corbie Cleugh, a small stream incised into the same slope. These characteristics strongly suggest that the five gullies are subglacial chutes, and, as such, must have been cut into the wall of a channel that had already been formed (or was forming) subglacially. The only apparent alternative to this interpretation is that Middle Dean was cut during a previous glacial phase possibly in a subaerial fashion, but there is no evidence either to support or refute this suggestion. The depth of penetration by meltwater indicated by the Middle Dean channel is in excess of 425 feet, considerably more than that observed in the majority of meltwater drainage systems elsewhere in the Cheviots and in Scotland.

The Singmoor System (50, Map 10): Channels comprising this system begin in two deep cols in the ridge of country forming a watershed between the Biddlestone and Harden Burns. The highest members of the series are located at approximately 1,200 feet in the col between Singmoor and Cold Law, where a 20-foot gorge heads on the brink of the Biddlestone valley (Figure 4.1). Adjacent signs of meltwater erosion are relatively insignificant compared with this feature. Beyond the confines of the col, the Singmoor channel steeply descends the hillside in a series of plunge-pool sections, the deepest of which is over 50 feet, and then grades fairly accordantly into the floor of the channel emerging from the Cold Law/Bleakmoor col.

Along this section of the Singmoor channel a complicated network of interconnected channel segments occurs on the left wall. They all lie at a higher level than the main channel and narrow ridges of bedrock form inter-channel divides.

Between the steep, south-east slopes of Cold Law and its lower appendage, Bleakmoor Hill, lies a distinct col forming an extension to the head of the Harden Burn valley. The col is dominated by a meltwater channel which begins with a double intake and quickly becomes 25 feet deep. The higher intake segment runs obliquely across the contours of the slope to join accordantly the lower segment, heading on the col crest. Immediately beyond the relatively narrow confines of the col additional complexities appear in the channel as they do at a similar stage in the Singmoor system. In this instance a double-headed tributary enters on the right bank and joins the right loop of the channel floor, which has bifurcated round a small rock knoll here. A further characteristic is the conspicuous loop, over 15 feet deep, that hangs about 10 feet above the channel floor near its confluence with the Singmoor system.

Previous reference to these channels has been somewhat fragmentary, for Smythe (1912) only mentioned their relevance in feeding "Biddlestone waters" towards the large meltwater cuts at Alnham and Scrainwood. Anderson's map portrays the channel pattern with more accuracy than that of Smythe, but he alluded to them with equal vagueness. Discussing the proposed lake dammed up in the Spartley valley, he stated, "The earliest feeders of this lake came in from the west, on either side of Cold Law, where the northern one had a double intake at 1,120 and 1,090 feet." The exact way in which Anderson believed the channels were formed is not clear from the above description, but they were probably part of his "direct flow" category. The former presence of an ice-dammed lake in the Spartley valley has already been discounted in this chapter, and the validity of interpreting such complex channel networks as "direct flows", questioned. The steep, frequently oblique, downslope trend of the channels and the presence of narrow inter-channel divides, are characteristics that require the immediate presence of glacier ice during their formation. Since the maximal altitude through which the Singmoor channel descends is 250 feet (1,200-950 feet), the streams are unlikely to have flowed between open walls of ice. Accordingly, the entire system is interpreted as a subglacial formation.

The combined meltwaters of this system flowed into the pre-existing Harden valley. The present stream is clearly underfit in a valley floor that was presumably cut partly by these meltwaters, but it is extremely difficult to gauge the extent to which they deepened the valley and to determine how far down-valley meltwater erosion was effective. All that can be stated with confidence is that meltwaters were guided by this pre-existing route and flowed down it.

The Alnham Channel (48, Map 7): At the southern extremity of the Northfieldhead-Castle Hill spur, an outlying hill rising to 814 feet appears to have been originally part of the spur, to which it was connected by a low col at 725 feet. A deep meltwater channel subsequently became entrenched through the col so that the hill mass now stands isolated from the main spur. One of the pre-existing valley-heads presently supplying a small headstream of the river Aln opens out from this col. The Alnham channel is one of the most impressive fluvioglacial formations in the south-east Cheviots, extending over a distance of more than $1\frac{1}{2}$ miles with rock-cut walls that frequently rise over 50 feet high; a maximum depth of about 100 feet is reached at one point (Figure 4.2). The feature begins as a broad, marshy depression on the left bank of the Scrainwood Burn, west of the former col, and climbs several feet uphill over the initial 100 yards of its course. When it enters the col, precipitous rock walls quickly develop as the channel swings through in two great curves, the undercut wall of one rising over 100 feet above the floor. Two minor chute-like features furrow the left wall. Beyond the col, the channel form broadens considerably as the gorge continues down the line of the pre-existing valley. At point A, the channel widens abruptly in order to accommodate a deep meander loop that branches off to the right, around a prominent rock knoll. The main section of the channel continues in a straight line to point B. An uphill gradient characterises the first 130 yards of the meander loop, causing the floor to rise about 25 feet to its crest in the meander apex. Beyond that point the loop curves and slopes quite steeply down to rejoin the lower route accordantly at C. To the left of point B, another offshoot branch curves uphill round a massive ridge that represents part of the pre-existing valley-side, isolated as an erosion remnant. This second loop rises almost 30 feet to a crest in its floor profile before descending steeply to rejoin the lower route accordantly at D. At point D, the pre-existing valley bends eastwards from its initial alignment towards the north-east, but meltwaters were apparently unable to follow this route, for the channel continues north-eastwards beyond D, past Alnham, where it is incised into the lower slopes of a broad spur from Northfieldhead Hill. The channel lacks distinct walls at D, where the broad pre-existing valley leads off towards the east, and only a wide marshy flat connects the two conspicuous gorge sections of the main Alnham channel. The section from D to E on the flanks of Northfieldhead Hill is distinctly asymmetric in cross-profile; the upslope wall rises 60 feet, the opposite wall, 45 feet. Towards its outlet, the channel becomes progressively smaller and narrower until it ultimately fades away at 425 feet on the western side of the Aln valley.

Smythe (1912) observed the Alnham channel and suggested that it had functioned as an overflow draining a former ice-dammed lake in the Spartley valley, an interpretation later endorsed by Anderson (1932). On the basis of Anderson's map and description, Sissons (1961a) concluded that "The deep gorge ..... was formed by a large subglacial river that was supplied with water from a large area." The former existence of an ice-dammed lake in the Spartley valley has already been discounted, so that the impressive Alnham channel cannot have been excavated by overflowing lake waters. Indeed, it contains several characteristics that necessitate the close proximity of glacier ice during formation. For example, at D, a wall of ice must surely have guided meltwater from that section of the gorge south-west of Alnham into the section on the flanks of Northfieldhead Hill, preventing the escape of water eastwards down the pre-existing valley. Furthermore, that part of the channel between D and E, cut along the lower slopes of Northfieldhead Hill can only be accounted for if an ice barrier hindered meltwater from flowing directly downslope towards the base of the slope. Loops A-C and B-D cannot be explained by subaerial flow of meltwater. Both are continuous erosional features cut by streams flowing in a general north-easterly direction, and the uphill sections are believed to be entirely original. There is no evidence that suggests they are abandoned high-level loops, the uphill segments of which are actually postglacial formations. On the contrary, the absence of any debris fans from the extremities of the loops, and their accordant junctions with the central route within the channel, suggests contemporaneity for the whole of this intricate network. This implies a subglacial origin for the channel because hydrostatic head of pressure would have been necessary to force the meltwater streams uphill round the loop sections. This reasoning also applies to the channel's intake, which rises several feet before entering the main gorge. The actual mechanism by which a large subglacial river became braided to form this figure-8 pattern is difficult to establish, however. It is perhaps possible that sections A-C-D and A-B-D were the two original branches and that the breach B-C was a later formation, but the initial split to form the two original branches would still remain unexplained. Since the entire system appears to have functioned simultaneously, it may be considered that water from the vast subglacial river flowing through the narrow col under considerable hydrostatic pressure forced its way laterally out of the uniform channel as it entered the more spacious environment of the pre-existing valley at A, resulting in a braided pattern of flow. Alternatively, the superimposition of a braided englacial stream network could equally have resulted in this way.

The Harden Burn valley, down which flowed considerable volumes of meltwater, is in conspicuous alignment with the Alnham channel, as indeed are channels on the flanks of Harden Hill. It therefore seems unrealistic to avoid connecting these lines of meltwater drainage, as Sissons (1961a) chose to do on the basis of Anderson's description. Consequently, it is suggested that the enormous volumes of water required to explain this large channel came predominantly from the combination of subglacial meltwater drainage systems to the south-west, from the Harden valley and the flanks of Harden Hill. These supplies may have been augmented by drainage down the Spartley Burn valley. It was the concentration of this water that breached the low col through the Northfieldhead-Castle Hill spur and eroded the remarkable channel as far as the upper reaches of the Aln valley, where the meltwater appears to have resumed an englacial course, for no further trace is to be seen on the ground.

Conclusion on Subglacial Drainage

From the foregoing discussion, it emerges quite clearly that the form and location of most meltwater channels in the south-east Cheviots can be adequately explained only by the subglacial flow of meltwater. In addition to the presence of up/down longitudinal floor profiles, common to many channels, the occurrence of complex networks of interconnected segments and abandoned loops, with narrow ridges and knolls of bedrock between, also constitutes evidence in support of the subglacial interpretation. Apart from several channels on Northfieldhead spur and a few in the Harden Hill system, there is a remarkable concentration of these features in cols and valley-heads, a relationship that is equally prominent in the north-east Cheviots. Topography in the area under consideration consists predominantly of deep steep-sided valleys and embayments separated by interfluves, the crests of which are frequently indented by cols. As outlined in Chapter 2, the presence of subglacial meltwater channels in an area of such broken relief, strongly implies that the rivers of meltwater reached subglacial positions by being superimposed on to underlying topography from englacial courses within the downwasting ice. It is therefore concluded that the majority of meltwater channels in the south-east Cheviots were formed by the superimposition of englacial river systems onto the ground beneath as the ground intersected their courses consequent upon the downwastage of the glacier ice in which they were contained.

The System as a Whole: Conclusion

Meltwater channels in the south-east Cheviot area, between the rivers Breamish and Coquet, have now been described in considerable detail. The rather generalised and simplified accounts written by Smythe (1912) and Anderson (1932) for some of the channels have been criticised in the light of current theories on the formation of meltwater channels, but their pioneer work is much appreciated and Anderson's map is of outstanding quality for that time. A reappraisal of the various channel systems has revealed important complexities in form that were previously ignored or dismissed as irrelevant. In many instances, these apparently minor intricacies in channel development have betrayed the subglacial nature of the systems. These, together with other characteristics inherent in the channels, were fully discussed before reaching the conclusion that they represent the incisions made by streams superimposed from former englacial positions.

Several important and intriguing points emerge from a general appreciation of the channel network as a whole, and these may be of considerable significance in an interpretation of glacial events in northern Northumberland.

(1) Since the direction of the major meltwater channels in an area is usually parallel with the former direction of ice-movement, it seems reasonable to infer that the alignment of most channel systems, in the south-east Cheviots, clearly indicates their association with a mass of glacier ice that invaded this area from the west, spreading north-eastwards round the south-east flanks of the massif.

(2) The uppermost channel in the area is located at just over 1,200 feet,

and although this indicates nothing more than the minimum level of glacier incursion, the absence of channels on hillsides above this elevation strongly contrasts with the remarkable concentration of such features below.

(3) The altitude of the uppermost channels on flanks of the massif declines from south-west to north-east.

(4) Approximate boundaries to the incursion of this western ice mass into the south-east Cheviot area may be tentatively established in the following places:

(a) Above Biddlestone it encroached to at least 1,300-1,400 feet. From this level on Cold Law, an approximate line running north-eastwards through Castle Hill and Cochrane Pike may effectively define its penetration into the massif, for north-west of that line, channels appear to be associated with a separate ice mass and will be discussed in a subsequent chapter.

(b) To the north, it seems to have penetrated no farther than the Breamish valley, for the trend of the channels north of the river implies that they are not associated with the southern ice mass.

(c) The channels beneath the scarp face at Lorbottle (57, Map 11) are the easternmost that can be ascribed to the western ice lobe. Farther east, meltwater channels are aligned from north to south and were probably linked with the northern ice mass.

(d) Coquet Dale and the area to the south-east lie outside the scope of this thesis, but meltwater channels in these areas have been observed on aerial photographs, and, together with those shown on Smythe's map, indicate fluvioglacial drainage towards the south-east. For example, the channel known as Selby's Cove, cuts through the Simonside Hills, and that by High Carrick at the head of the Elsdon Burn also leads south-eastwards.

(e) South of the Simonside Hills, the majority of meltwater channels that have been mapped by Smythe trend predominantly in an easterly direction and clearly relate to the western ice mass.

The overall pattern of meltwater channels thus suggests that the marginal zone of the western ice mass was channelled into low-lying ground between the flanks of the volcanic massif and the great escarpments of Thrunton and Simonside. Although the latter hills rise almost to 1,450 feet, the presence of Selby's Cove indicates that they were probably submerged beneath the ice. The extension north-eastwards of this ice mass reached the vicinity of Glanton, where it came into conflict with northern ice from the Tweed valley. The zone of confluence (or conflict) appears to have extended generally southwards from the Glanton-Thrunton area, east of which southward moving ice led to the formation of meltwater channels aligned towards the south and south-south-east.

(5) When the meltwater drainage systems were flowing at the higher levels (e.g. between 1,200 and 1,000 feet), the water presumably escaped from the area englacially in an easterly direction, over the low ground between the volcanic massif and the sandstone escarpments. The ultimate escape of meltwaters from the south-east Cheviots was by way of the valleys of the Breamish, Aln and Coquet. While drainage from some of the channels entered the Breamish catchment, it is almost certain that this water was diverted south-eastwards into the Aln valley, for it was demonstrated in Chapter 2 that there was no outlet available down the Breamish-Till depression until a very late stage in the deglaciation. Consequently, it appears that the Aln valley received virtually the combined volumes of meltwater liberated by the southern and northern ice masses that flowed round the flanks of the east Cheviots - especially during

the major period of deglaciation and as long as Shawdon Dean functioned as an outlet for meltwater from the Hedgeley Basin. Since there are no low outlets towards the south from the Aln valley, it must be concluded that enormous volumes of meltwater followed the alignment of this valley eastwards to the North Sea basin. At a later period, the Coquet valley was utilised as a convenient outlet for fluvioglacial drainage - as indicated by the Trewhitt channel system - but the volumes of water involved appear to have been much smaller than in the Aln valley,

CHAPTER 5.

FLUVIOGLACIAL DEPOSITS IN THE SOUTH-EAST CHEVIOTS

Introduction

In the north-east Cheviot area, a most intriguing complex of fluvioglacial deposits forms an extensive blanket of varying thickness over the lower slopes of the volcanic massif and the fringing basins of Cementstone. For the most part, these deposits consist of eskers and kame terraces and are liberally strewn with kettle holes and dead-ice hollows. It was suggested in Chapter 3, that the esker systems in particular appear to represent the same drainage alignment as that indicated by the great channel systems on adjacent hillsides, and the two sets of phenomena were probably formed in close association with each other. Since a comparable series of meltwater channel systems is located in the south-east Cheviots, the margins of which are also flanked by broad, low-lying depressions, some evidence of fluvioglacial deposition similar in nature and extent to that associated with the northern channels, is to be expected. Yet one of the most astonishing aspects of the landscape in this area is the almost complete absence of ice-contact fluvioglacial deposits. In order to unravel this apparent dichotomy of landform association, each of the low-lying areas in which extensive deposition might have been expected, will be discussed.

The Lower Breamish Valley

One of the most remarkable parts of the Breamish valley is that which extends from Ingram to Low Hedgeley, a distance of approximately $3\frac{1}{2}$ miles (Map 8). West of Ingram, the valley is relatively narrow, with precipitous slopes rising sharply above the floor, which is from 150 to 250 yards wide in this vicinity. East of Ingram, the valley sides open out abruptly and rise much more gently from the floor, which now becomes considerably wider as it broadens from 400 yards at Ingram to 1,250 yards at Brandon. The configuration of this section may well have developed chiefly in the late Tertiary through differential denudation on varying lithology, but this is not the most significant aspect in the context of this chapter. Equally intriguing is the vast plain of coarse gravels and cobbles extending from side to side of the valley from Ingram to Low Hedgeley. The sharp angle of intersection between the valley sides and the gravel plain implies a considerable depth of infill. Furthermore, at no other place in the Breamish/Till valley do gravels and cobbles of this quantity and calibre occur. They begin suddenly at Ingram and terminate equally abruptly at Low Hedgeley, and their outcrop is sufficiently conspicuous to merit a distinct symbol on the 1:25,000 and 1:10,560 map sheets of the Ordnance Survey. Opposite Low Hedgeley, the gravels have been proved to be approximately 6 feet in depth and from this place they thin rapidly downstream until within less than 100 yards, they merge into a floodplain composed of alluvial silt and fine gravel. Upstream, borings made by the local sand and gravel company indicate that the gravels and cobbles thicken progressively, until 20 feet occur near Brandon. It is also possible that a considerable depth of these materials fills the Breamish valley in the vicinity of Ingram.

From the Breamish headwaters to Ingram there appear to be few sources that could supply the present streams with such quantities of debris. The incisions made into the sporadic deposits of drift by tributary streams within the Breamish catchment area are relatively slight and the calibre and quantity of alluvial materials that seem to be currently moving down the upper Breamish valley appear inadequate to satisfactorily account for the Ingram plain in terms of present processes. Indeed, the river normally does no more than flow over a shallow bed in the gravels, although it may achieve a certain amount of redistributory work in times of flood, and by Low Hedgeley the Breamish flows in a channel cut 10 feet below the surface level of these materials. Accordingly, it is suggested that the Ingram gravel plain may not be related to the normal regime of the Breamish and that a more logical explanation involves the last period of deglaciation in the east Cheviot area.

A most unexpected characteristic of the wide, lower Breamish valley is the notable absence of ice-contact fluvioglacial deposits. Yet two very large meltwater channels, over 80 feet deep in their upper reaches, enter the valley at Ingram, and others join further upstream. The vast, subglacial rivers of meltwater that discharged down the steep gradients of these channels presumably carried considerable volumes of debris, and it is reasonable to expect much of this debris to have been immediately deposited as the velocity of the meltwater streams became arrested when they encountered the low-lying floor of the Breamish valley. One might envisage the broad valley as an environment admirably suited for the lingering presence of stagnant glacier ice and the concomitant deposition of impressive esker systems by the subglacial meltwaters that are known to have flowed in this area. However, the only relief formed by fluvioglacial sands and gravels occurs immediately east of Branton village, where a large, amorphous mass, bounded on the south-west and south by a large kettle/dead-ice hollow, slopes gently towards the south-east (Map 8). It rises approximately 60 feet above the Ingram gravel plain and has evidently been undercut by the waters responsible for that deposit. Since it is in direct line with the outlet of Brandon Dean and slopes with the same alignment, the two features are probably connected. The four kame terraces at Brandon have already been discussed (Chapter 3), and, although these deposits may in part have been derived from Breamish meltwaters, they are of relatively small extent and are unlikely to represent the entire volume of deposition in this area. Similarly, the rather indistinct esker-like masses east of the Brandon Dean outlet are of little significance.

The anomalous absence and restricted extent of ice-contact fluvio-flacial deposits in the lower Breamish valley may be explained either by -

(1) no original deposition of ice-contact forms; this implies the unimpeded flow of water and the continual transport of large quantities of debris through and under stagnant glacier ice; or,

(2) considerable deposition of ice-contact forms that were removed almost completely by subsequent processes.

Considering these alternatives, the kame terraces at Brandon and the amorphous mass of sands and gravels at Branton, indicate that the environment in which meltwater drainage occurred readily encouraged ice-contact deposition. For this reason, and in view of the extensive body of fluvioglacial deposits in adjacent areas to the north and east, the first suggestion is considered unlikely. On the other hand, the kame terraces at Brandon, the deposits near Branton, and the terrace-esker complex in the Reaveley valley, tend to support the second alternative. In this respect, their location is of extreme importance, for each mass of deposit is outwith the main alignment of the valley floor and shelters in the lee of a pre-existing embayment. The kame terraces and associated eskers are located well beyond the reach of any erosive activity that may have occurred in the Breamish valley subsequent to their deposition, except, perhaps, for the lowest terrace at Brandon. The mound of fluvioglacial deposits near Branton exhibits a distinctly undercut face fronting onto the Ingram gravel spread, and was evidently much more extensive formerly. It therefore seems that a solution to the anomalous absence of ice-contact forms of sand and gravel in the lower Breamish valley is indicated by these facts. A substantial part of the Branton gravel mass appears to have been removed, but it was truncated only until it had retreated back to a position in which it was protected by the spurs of bedrock at East Hill and Cow Hill. The kame terraces at Reaveley and Brandon were already in sheltered localities and remained relatively unmodified.

It is therefore concluded that ice-contact deposits may well have occupied much of the lower Breamish valley following the stagnation of glacier ice, but subsequent events almost obliterated them completely. The present stream provides little indication that it could ever have accomplished such extensive modification over this wide expanse of valley, and it seems more logical to recall a period during which larger volumes of water flowed down the Breamish valley. At such a time, any eskers and kame terraces that may have been located on the valley floor were severely modified (in the same way that such formations near Wooler are presently being degraded), and their material redistributed to form the extensive gravel plain between Ingram and Low Hedgeley. This hypothesis accounts for the abnormal concentration and wide expanse of cobbles and gravels, when no present source is readily apparent. Since the surface of this gravel plain terminates slightly below 300 feet, it is perhaps significant to recount the former presence, in the Hedgeley Basin, of an ice-dammed lake whose surface rose to a similar level (Chapter 3). In this respect, the gravel plain represents the deltaic counterpart, in the Breamish valley, of the Wooperton outwash delta further north. It will be suggested in Chapter 8 that the mid-Breamish valley probably contained part of the Cheviot ice cap, the glacier mass having moved down-valley in this area. At the stage of deglaciation that gave rise to the Hedgeley glacial lake, Breamish ice was presumably in a state of active recession. It therefore seems reasonable to suggest that large volumes of meltwater surged proglacially down the Breamish valley at this time, truncated and redistributed any ice-contact fluvioglacial deposits between Ingram and Brandon and built the Ingram gravel plain out into the Hedgeley lake.

The Upper Aln Valley and Adjacent Areas to the South

The upper Aln valley is relatively broad and is bordered by gently sloping hillsides culminating in elevations of approximately 900 feet to the north and 700 feet to the south (Map 8). The smooth sides and floor of the valley are totally devoid of fluvioglacial deposits, so much so, that it would seem questionable whether or not any meltwater drainage had passed this way at all. Yet, as described in Chapter 4, impressive systems of meltwater channels lead directly into the upper reaches of the Aln valley and indicate that immense volumes of subglacial drainage did, in fact, flow in this vicinity. A short distance to the south, the Biddlestone group of channels illustrates how a good deal of fluvioglacial drainage was conducted into the wide valley presently drained by the Netherton Burn (Map 10), but, similar to the Aln valley, there is hardly a vestige of meltwater deposition to be observed. In the same way, the neighbouring Foxton and Trewhitt valleys contain evidence of meltwater erosion in their upper reaches, but depositional counterparts are entirely absent from the floors and lower basins.

Indeed, the only fluvioglacial deposits recorded by the Geological Survey and the writer, in this part of the south-east Cheviots, occur as very small patches in four places (Map 7). These are as follows:

(1) Grid Ref. 3968/5086. Fluvioglacial cobbles and gravel, contained in a sandy-grit matrix, compose a short ridge (5-6 feet high) and rather amorphous undulations on the gentle slopes at the foot of Harden Hill, east of Biddlestone. They are probably part of the meltwater drainage system that eroded channels at Biddlestone and on the Eli Law-Ewe Hill ridge. There is no evidence to suggest that these deposits were ever more extensive than their present outcrop, and

they probably represent localised deposition over a small area.

(2) Grid Ref. 3976/5103. Another patch of fluvioglacial debris occurs on the crest of the interfluve north of the Harden Burn. It is in the form of a narrow ridge, about 280 yards long, that seldom exceeds 10 feet in height, although at its lower end, one side rises to over 20 feet. There are no sections to reveal the true nature of its internal composition, but minor exposures and surface litter show pebbles that are apparently water-worn. The ridge is possibly an esker. The small extent of this deposit is hardly commensurate with the composite system of deep channels that scores the hillsides about Sing Moor and Cold Law and it probably represents only a minor, localised phase of deposition from that drainage system.

(3) Grid Ref. 3976/5114. In the vicinity of the Coppath Burn-Spartly Burn confluence, three elongated mounds of sand and gravel form a discontinuous esker, between 10 and 15 feet high. This feature seems to mark a former line of meltwater drainage aligned obliquely down the hillside west of the Coppath Burn before turning to flow down the Spartly Burn valley. As such, it probably represents the route that meltwaters were obliged to assume subsequent to the abandonment of drainage across the upper part of the Northfieldhead spur.

(4) Grid Ref. 3988/5124. The remaining deposit of note to be included in this section occurs on the steep slopes of Hart Law, approximately 200 yards below the outlet of channel 1 in the Northfieldhead system. It is a narrow-crested ridge varying in height from 5 to 20 feet and running obliquely downslope. Clear sections are not available, but a sheep-hollow reveals subangular - sub-rounded gravel in a coarse sandy matrix, and blocks up to 2 feet in diameter lie on and near the surface. It is a clear example of a subglacially engorged esker, the type first described by Mannerfelt (1945).

The constituent materials were probably derived from channels 1 and 2 on Northfieldhead spur, for the sub-angular nature of many stones suggests that they have not been transported far, and hence experienced only minimal attrition. The esker is thus simply a local deposit of no major significance to the general pattern of deglaciation in this area.

Apart from these four localised patches of fluvioglacial materials, there are no deposits of sand and gravel in the south-east Cheviots to compare with the remarkable esker complexes north of the Breamish, despite the fact that impressive systems of meltwater channels furrow the hillsides. At first, the contrast seems rather incredible and appears to require an explanation for which there is apparently little guiding evidence. However, a closer inspection of the meltwater channel systems south of the Breamish may point towards a reasonable solution in the following manner.

(1) Three of the largest channel groups in the south-east Cheviots are those that lead into the Breamish valley, and it has already been demonstrated how any associated deposits contributed to the great gravel spread between Ingram and Low Hedgeley.

(2) The seventeen channels on the Northfieldhead-Castle Hill spur are very small features, except for channel 13, and because the associated meltwater streams resumed englacial courses on the lee side of the spur, between 400 and 500 feet above the level of the valley floor, the absence of correlative deposits is not surprising.

(3) The channels by Biddlestone, Newton and Trewhitt are somewhat larger, and each intricate system is composed of numerous branches. They all terminate on open, low-lying ground, and it would seem reasonable to expect some quantity of debris eroded from these channels to have been deposited in such a seemingly favourable environment; yet such deposits are totally lacking.

The most reasonable explanation for this anomaly almost certainly involves the accordant alignment of these channel systems with the pre-existing valley-heads into which they are cut. This situation would have allowed unimpeded flow of meltwater down the valleys, thereby enabling the relatively small quantities of debris eroded from the channels to be transported outwith the area and into the Coquet valley.

(4) Similar topographic relationships pertain in the upper Aln valley, into which grade the massive channels from Hazeltonrig, Alnham and Scrainwood. The volume of debris excavated from these gorges, however, is quite considerable, and even unimpeded flow down the Aln valley must have encountered serious difficulties in transporting such huge quantities of sand and gravel completely outwith the area. Nevertheless, even although the upper valley is broad, with a flat floor and moderately sloping sides, there are no ice-contact ridges, mounds and kettle holes. Further down-valley, however, especially east of the road that runs from Little Ryle to Great Ryle, it soon becomes apparent that the wide flat in which the present stream is entrenched is not the alluvial floodplain of the Aln. Broad terraces of fluvioglacial sands and gravels extend as conspicuous landforms as far down-valley as Low Barton, and it is considered that these formations represent the deposits of the meltwater drainage that was directed into the Aln valley. It seems that glaciological conditions during the recession of ice in this area did not promote the development of fluvioglacial landforms often associated with deposition in fragmented, stagnant ice, an environment in which braided meltwater streams commonly deposit esker complexes. Indeed, the deposition of sands and gravels in the Aln valley occurred where little or no dead-ice remained, for only uncertain evidence of ice-contact slopes can be observed in association with the terraces. Since a fuller discussion of the deposits is given in a subsequent section of this chapter, it is perhaps sufficient to conclude, at this stage, that the

volume and gradient of meltwater drainage entering the upper Aln valley was adequate to transport concomitant debris a considerable distance beyond the channel mouths, until deposition was achieved in an environment relatively free of stagnant glacier ice. The accordant alignment of the ice-directed meltwater drainage with the Aln valley, may have been an important contributing factor to this situation.

In view of the foregoing observations, it may be concluded that although the multitude of fluvioglacial channels in the south-east Cheviots illustrates that the marginal zone of the southern ice lobe contained many systems of meltwater streams, the lower ground does not hold commensurate evidence of such drainage, in the way of eskers, kame terraces and kettle holes, in direct contrast to low ground north of the Breamish.

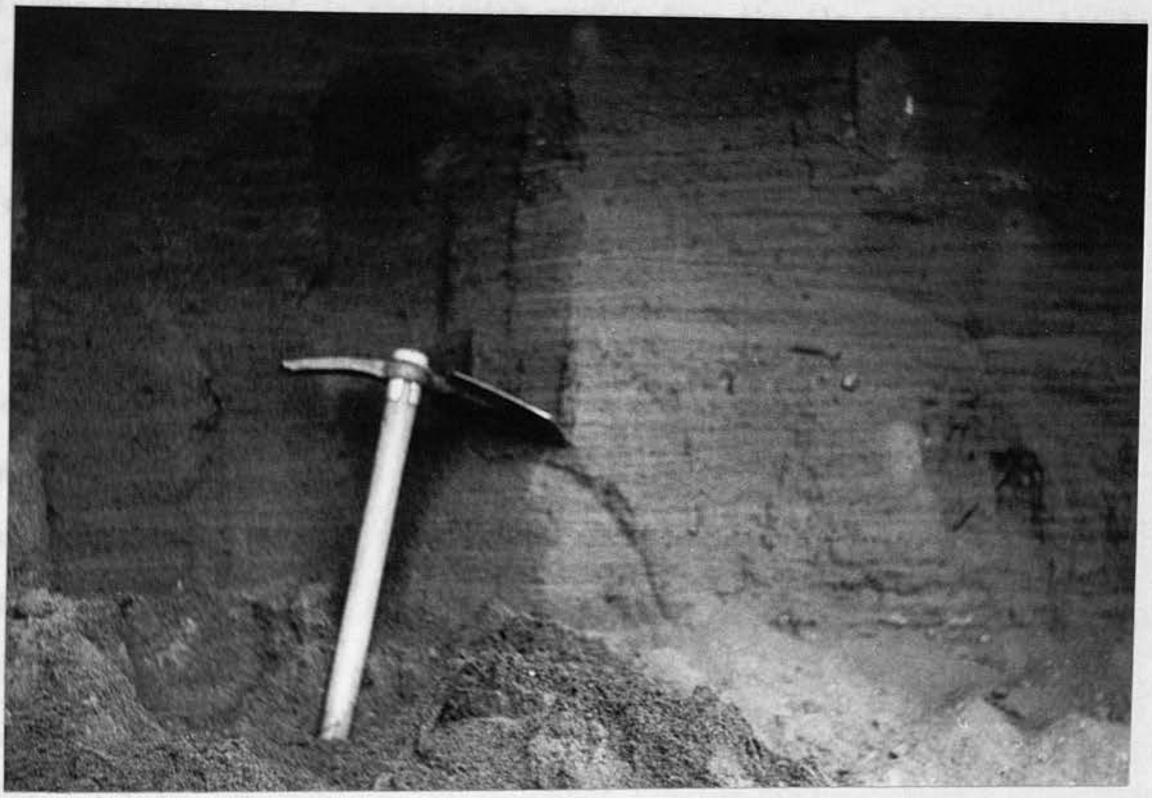
The Mid Aln Valley

Downstream from Whittingham, the Aln valley begins to open out considerably, particularly where the wide tributary valleys of the Coe and Shawdon Burns enter from the south and north-west respectively (Maps 8 and 9). In this locality, the mid Aln valley is perhaps more properly referred to as a basin, since the valley floor here attains its greatest width of approximately 1,200 yards. Relatively gentle and moderate slopes rise above the rather featureless and gently sloping plain. Deposits laid down by glacial meltwater certainly occur in this area, but their composition and topographic expression are almost totally devoid of characteristics normally associated with an ice-contact environment. They may be discussed under the following headings.

(1) The Whittingham Terraces. From Alnham to Little Ryle, the river Aln presently flows over an infill of till covering the floor and lower slopes of the valley. All that the stream has accomplished post-glacially is the exca-

Photograph 5.a The Whittingham delta surface (Aln valley) looking up-valley from the east-north-east.

Photograph 5.b Finely bedded coarse sand in the Whittingham delta, left bank of the Aln downstream from Whittingham.



vation of a narrow channel no more than 6 to 8 feet deep, and generally less; it presently meanders from side to side between shallow walls. The catchment area for this upper part of the Aln valley provides tributary streams no larger than mere ditches, and the present river channel seems proportional to the volume of water it contains. Downstream from Little Ryle, a distinct terrace borders the river on either side, instead of the gentle till slopes so prominent farther upstream, and east of Ryle Mill, the terrace surface becomes increasingly higher above the floodplain of the river. Upstream from East Lodge (E.L. Map 8), poor exposures in three places reveal sand and water-worn gravel of a coarse calibre. Although the terrace surface is over 15 feet above the floodplain at East Lodge, it is mainly downstream from this point that terrace forms become more prominent in the landscape (Photograph 5.a). Two separate levels can be distinguished and perhaps their most dominant characteristic is their fragmented nature.

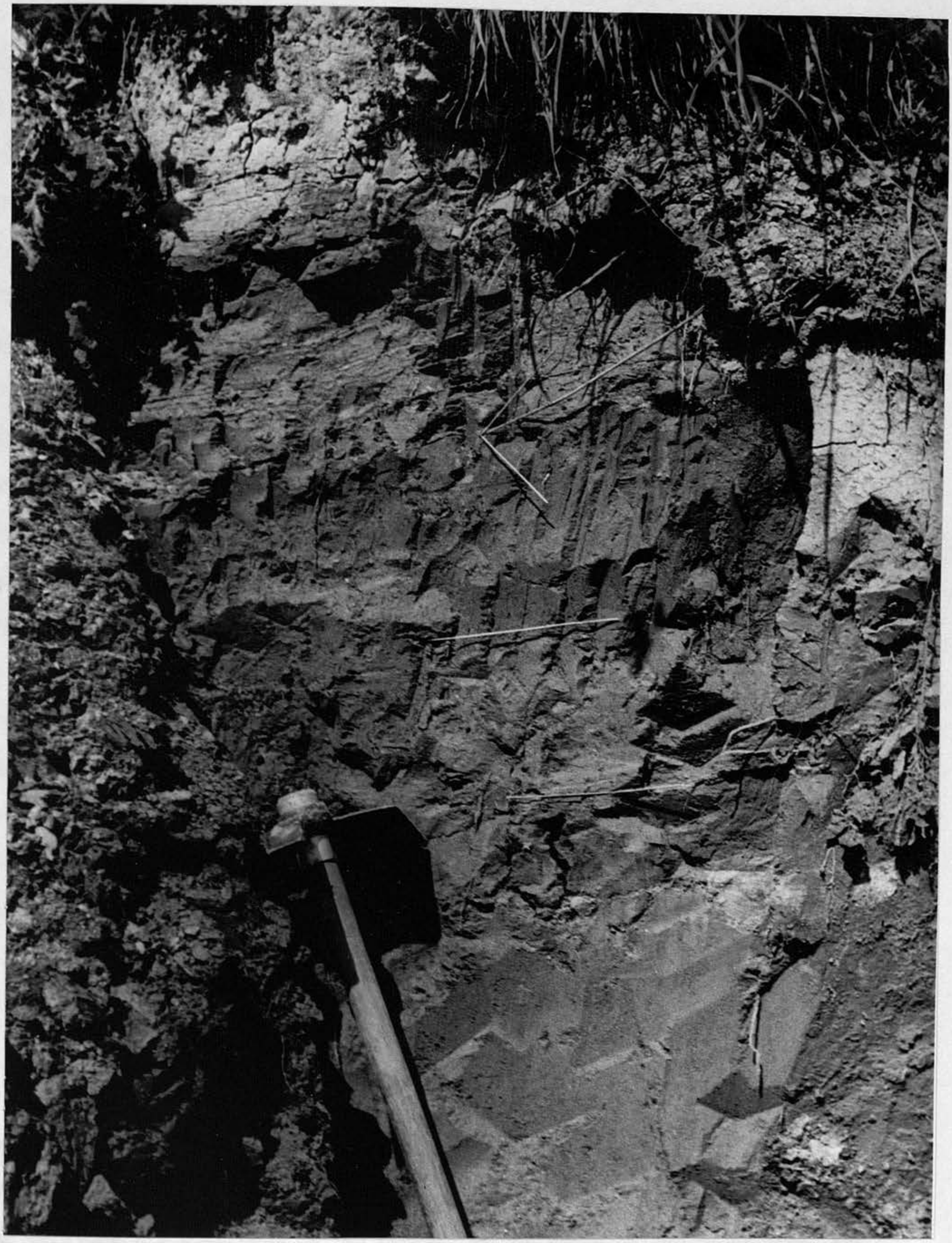
(1) The higher level is that which grades down the Aln valley from Little Ryle to areas (A) and (B), where it terminates. Beyond, only two more fragments occur at a similar level; these are at (C) and (D), and possibly represent remnants of what may originally have been a much more extensive formation. For the most part, this upper level does not descend below 250 feet.

(2) The lower level is much more fragmented and was formed either contemporaneously with, or subsequently to, erosion of the upper terrace. This is suggested not only by the occurrence of possible out-lying fragments of the upper terrace rising above the lower formation, but also by the distinct, channel-like depression along the inner margin of the lower terrace at area (A). Between the various fragments of this lower terrace occur numerous linear depressions that resemble old stream courses; for example, at points (i), (ii), (iii), (iv) and (v). Numerous small sections in these terrace fragments

reveal that they are predominantly of a sandy nature, although water-worn pebbles sporadically litter the surface. One section (Grid Ref. 4076/5122) shows at least 3 to 4 feet of coarse sand (Photograph 5.b), and at another place, sand occurs to the base of the undercut bank, which is approximately 25 feet high. The northern limit to the extent of this deposit is difficult to establish by eye, for parts of the terrace surface are slightly undulating and it is not always flat; but by means of augering, a reasonably accurate junction with the gentle till-covered slopes adjacent was established. The lower terrace is everywhere below 250 feet and descends from slightly below that height to approximately 200 feet at its furthest extension down-valley. That the materials composing this level were not wholly derived from the upper reaches of the Aln is clearly demonstrated by the position of certain terrace fragments. For example, fragments (a) and (b) slope gently out of the Callaly Burn valley, and the gap between them was probably cut at a later date by the stream responsible for their deposition. The "post-glacial" stream has been forced to bend sharply at right-angles to its main alignment by these terrace deposits, and joins the Aln much further downstream than it would otherwise have done. Fragments (c) and (d) are two sloping terraces at a similar level emerging from the Swine Burn valley. There are no sections to reveal the nature of underlying materials, but augering has established that they are composed predominantly of fine sand and silt. Similarly, fragment (e) is probably related to the Coe valley.

Evidence that suggests these deposits are probably of deltaic origin includes the terrace-like nature of their surface, the down-valley slope of this surface and the foreset bedding revealed in one section. The distinct linear depressions must also be accounted for. While the majority are probably stream-cut courses, the possibility that (i) and (iv) are dead-ice hollows

Photograph 5.c   Section on left bank of Aln downstream from Whittingham, showing, from top to bottom, laminated clay, silty clay, clay, sand.



should not be excluded, for they are somewhat larger and more irregular than the other depressions. It is reasonably clear that considerable volumes of water at one time flowed down the Aln valley and deposited these sands and gravels in a terrace form that was also partly eroded by distributary streams. It has already been suggested, in this chapter that meltwater from receding glacier ice was responsible for these formations in an environment that was largely ice-free; but the possibility that small remnant blocks of dead-ice persisted for some time after the onset of deposition may also be considered. Implicit in the suggestion that these deposits are of deltaic origin, is the former presence of a lake in the mid Aln valley. In addition to the delta terraces, direct evidence in support of the former existence of such a lake is the presence of laminated clays and silts. These represent the true floor deposits and may be observed at a number of places, the best sections occurring on the undercut left bank of the river Aln, approximately 900 yards downstream from Whittingham; these are as follows:

Section 1 (Grid Ref. 4078/6122) shows 15 feet of red and grey laminated clay, containing thin layers of micaceous silt, and overlain by 4 feet of sand with some small gravel.

Section 2 (Grid Ref. 4081/6123) shows the following sequence (Photograph 5.c):

12 inches ..... laminated clay and silt.

$6\frac{1}{2}$ inches ..... silt.

6 inches ..... silt, laminated clay, silt.

15 inches ..... fine sand - to base of the section.

Exposure 3 (Grid Ref. 4082/6124) is simply an outcrop of laminated clay near the top of a sloping, degraded bank.

Good sections are also exposed, in places, on the banks of the Swine Burn, for example:

Section 4 (Grid Ref. 4087/6119) shows:

9 inches ..... soil and gravelly alluvium.

12 inches ..... weathered grey clay (probably alluvium).

43 inches ..... laminated clay; brown, grey and red.

Section 5 (Grid Ref. 4087/6118) shows over 6 feet of laminated clay.

Section 6 (Grid Ref. 4086/6117) shows:

33 inches ..... stream alluvium (silt and gravel layers).

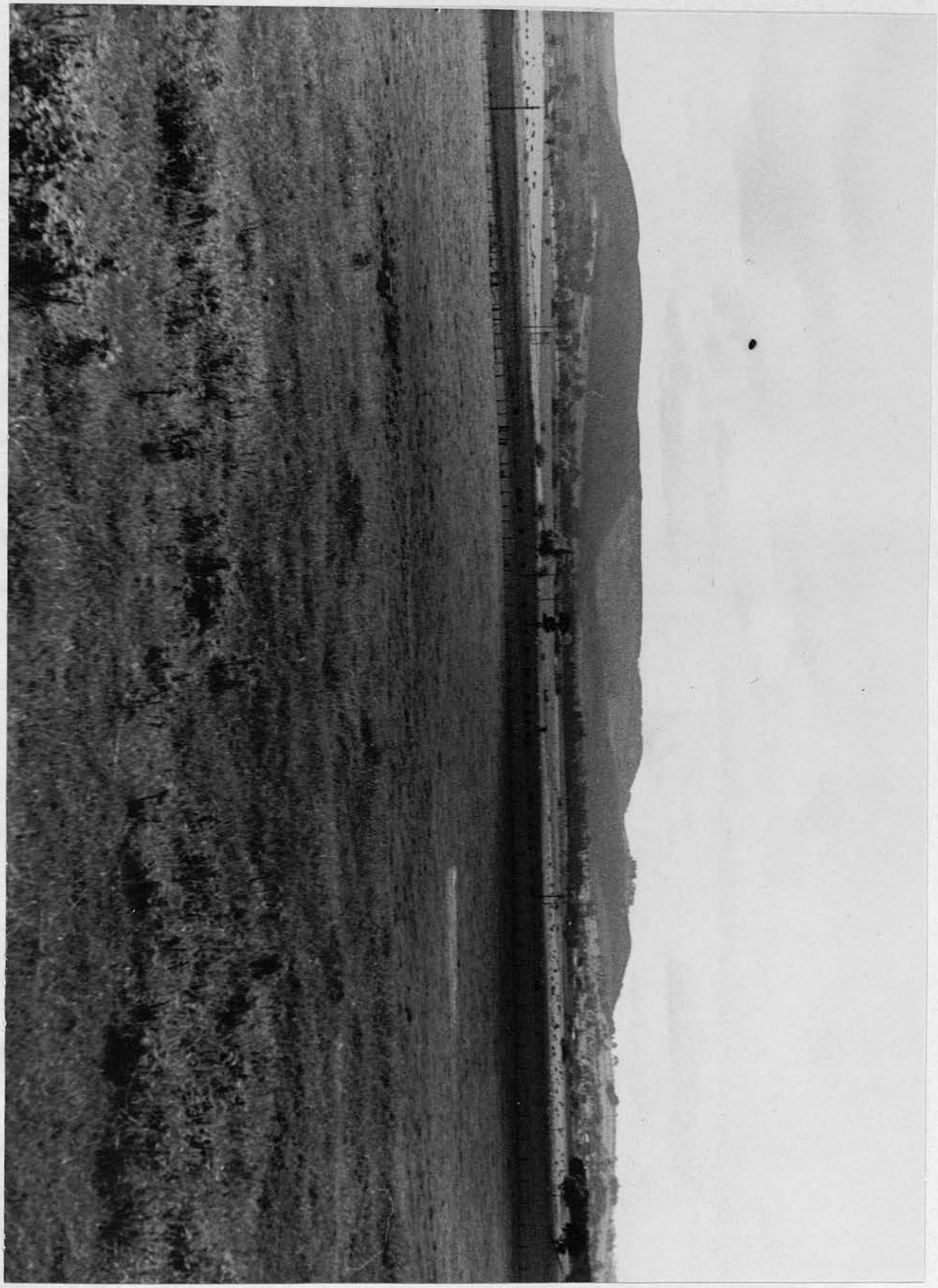
16 inches ..... laminated clay.

A significant point in relation to these exposures is that they all occur at or below 200 feet; no exposure of similar materials was observed above that level. This evidence strongly suggests that an extensive lake, in which laminated clays and silts were deposited, formerly occupied the mid Aln valley, downstream from Whittingham. Since the floor deposits of this lake do not appear to be present above 200 feet, and in view of the associated delta of sands and gravels, which slopes from 250 feet to approximately 200 feet, the latter figure may roughly represent the level at which the lake surface lay. The laminated nature of the lake floor deposits indicates that they accumulated in a glacial environment, a fact substantiated by the proglacial outwash nature of the deltaic sands and gravels at Whittingham. It remains uncertain whether or not the upper terrace was associated with a higher lake level, but it is almost certainly an earlier deposit, significantly composed of coarser materials. A discussion concerning the probable outlet for this lake and the barrier to drainage is deferred until later in this chapter.

The pattern of deglaciation in the Aln valley which can be reconstructed from the evidence so far presented, involves streams of meltwater that issued from subglacial meltwater channels terminating in the upper valley. This drainage system constructed two terraces of sands and gravels, the lower of which was certainly built out into a glacial lake in which laminated silts and clays were accumulating. That this deposition was chiefly in a proglacial environment, relatively free from stagnant blocks of ice, is suggested by the virtual absence of ice-contact phenomena. With the progress of glacier downwastage during deglaciation, it might be expected that the comparatively narrow upper Aln valley would have become ice free while considerable masses of stagnant ice remained in the broad basin of the mid Aln valley (bearing in mind that ice from the south-west moved over this area). On the contrary, these events did not ensue, possibly for the following reason. The approximate limits of penetration by the southern glacier lobe towards the north and north-east were established on the basis of meltwater channels in Chapter 4, and it was observed in Chapter 2 that the great Crawley/Shawdon channel marked the presence of the northern glacier lobe in that part of the Aln valley. Accordingly, the zone of contact between the two glacier lobes must have been roughly in the Whittingham-Low Barton area. When these glaciers began to wane, a certain amount of marginal recession would have accompanied vertical shrinkage, so that the area underlying the zone of contact became relatively ice-free at an early stage, although small remnant blocks of dead-ice may have lingered temporarily. An ice-dammed lake then formed in the mid Aln basin between the two receding ice masses, and vast quantities of proglacial meltwater, issuing from subglacial channels at the head of the Aln valley, built out two major outwash plains or deltas in the Whittingham area. Smaller amounts of debris were brought down the valleys of the Callaly and Swine Burns. There is no evidence, however, that confirms whether or not the upper terrace was deposited in a lake. An alternative possibility is that this formation may represent a proglacial valley train that was banked up against the western margin of the northern ice lobe as it lay across the Aln valley. Ultimately, to be as follows. The channel has already been interpreted as a subglacial formation cut under hydrostatic pressure, but it seems unlikely that the broad expanse of terrace extending into the Aln valley was also created subglacially before or during that period of channel formation. It is difficult to imagine a subglacial cavern large enough and sufficiently stable to accommodate the terrace deposits. Furthermore, no kettle holes, dead-ice hollows or fringing ice-contact slopes occur in association with the terrace. The period of terrace formation is more likely to have been much later, when the Glanton-Titlington ridge had become ice-free and meltwater was discharging proglacially through the Crawley/Shawdon channel. Indeed, this phase of fluvioglacial drainage probably coincided with the development of the englacial water-table in the Hedgeley Basin referred to in Chapter 3. Since the terrace deposits were subsequently eroded by a large volume of water draining through the Crawley/Shawdon channel, it seems likely that the mid Aln lake drained away or lowered in surface level before meltwater had ceased to flow southwards from the Hedgeley Basin.

Perhaps of greater significance in the context of this chapter is the coincidence in surface level of the Shawdon terrace with that of the Whittingham terrace. For this reason it is considered that both features were deposited approximately contemporaneously by volumes of proglacial drainage entering an ice-dammed lake in the mid Aln basin. The Shawdon terrace may therefore be referred to as an outwash delta. The two terrace fragments north-east of Bolton slope distinctly down-valley and demonstrate a drainage gradient in that direction. They also indicate that this relatively narrow section of the Aln valley might originally have been choked by these deposits from side to side. Indeed, this could have been partly or wholly responsible for impounding drainage from the upper and mid valley regions, resulting in a consider-

Photograph 5.d The mid Aln basin; area of laminated clays.



able lake. This potential barrier of sand and gravel should therefore be considered as a possible alternative to a dam of glacier ice, but the laminated nature of the lake floor deposits implies the close proximity of glacial conditions.

(3) The Mid Aln Basin. That part of the Aln valley, located between the two outwash formations at Whittingham and Shawdon, consists of a broad basin (Photograph 5.d) in which gentle slopes are quite devoid of any noticeable irregularities as they sweep gradually up from the floor to their limit at a clear break of slope at 200 feet. Although no satisfactory sections are exposed the heavy, clayey nature of the fields and of superficial exposures in small ditches, together with the presence of old clay pits in two places, strongly suggest that these gentle slopes are underlain by deposits that are predominantly clay. In areas where till mantles the bedrock surface, stones generally litter the ground quite freely, and boulders cleared from the fields are frequently piled up into large heaps at the field corners. Since there are no such characteristics present in the mid Aln area, it is considered that underlying clay deposits consist chiefly of the stoneless variety associated with lake floor deposits. This implies that the basin was inundated, approximately to a level of 200 feet, by the glacial lake and that till deposits below that height are generally mantled by a cover of stoneless lake clays.

The Eglingham Valley

That a glacial lake once occupied the mid Aln basin is strongly supported by the foregoing evidence, and it has been suggested that a mass of outwash deposits and/or glacier ice lying across the eastern extremity of the area might have been the effective barrier to drainage. In order to test the validity of this theory, it is necessary to discuss the various landforms of deglaciation present in the valley of the Eglingham Burn, a prominant tributary from the north-west entering the Aln at East Brizlee (Map 9).

The valley presently drained by the Eglingham Burn is a composite feature, consisting of several distinct sections that contrast considerably in their morphology. The stream, together with several minor tributaries, rises in a broad strike vale east of the Fell Sandstone ridge at Hepburn Moor, and flows southwards. In these upper reaches, the very small stream flows between shallow banks frequently composed mainly of peat and heather. At Bewick Moor, the stream enters a narrow, rock-cut section and plunges steeply from about 600 feet to 450 feet over a series of rapids and minor cascades, cutting through a thick infill of glacial drift. In the vicinity of Eglingham, the stream leaves the broad pre-existing valley leading westwards into the Hedgeley Basin and flows south-eastwards through a narrow rock-walled gorge cut through the former watershed. In the wide valley south-east of Eglingham, the Eglingham Burn flows in a broad, flat-floored gorge up to 50 feet deep. Exposures are generally poor, but occasional slumps, terracettes and the moderately steep angle of slope suggest that the gorge has been cut chiefly in glacial drift. The gorge deepens progressively until the stream is incised by over 100 feet at Shipley (Sy., Map 9). At this place, a gorge of similar form and dimension, presently occupied by the Shipley Burn, enters from the north, and beyond their confluence, a wide canyon leads the combined streams southwards into the river Aln.

The modern channel occupied by the Eglingham Burn, from its source on Hepburn Moor to its confluence with the Aln, is apparently the product of glacial events that altered the pre-existing drainage system in this area. That this stream alignment was established chiefly during deglaciation, when large volumes of meltwater flowed south-eastwards off Bewick Moor, can be demonstrated with reference to fluvioglacial landforms. The highest, and probably the earliest feature produced by meltwater activity on Bewick Moor (Map 6), is the 50-foot rock gorge at Dove Crags presently occupied by the modern stream. A remarkable pot-hole feature 20 feet deep, occurs on the right wall of this canyon, and 700 yards farther on, the canyon has deepened to over 70 feet. Evidence indicating the meltwater origin of this channel includes an abandoned loop segment and dry tributary channels, and adjacent deposits of fluvioglacial sand and gravel. Three hundred yards south of Blawearie, a meltwater channel leads down the steep face of the pre-existing valley head and is aligned parallel to the main canyon farther east. A narrow spur of drift, tapering from 70 feet to 5 feet in height is an erosional remnant fashioned by these two channels. One good section reveals 25 feet of stony till. The ridge crest has been breached in three places by small meltwater channels that clearly hang above the floor of the main canyon, and three other small meltwater channels furrow the adjacent hillside to the west. The steep gradients and oblique courses of all these channels indicate that they are probably subglacial formations.

It may therefore be concluded that considerable quantities of meltwater flowed from Bewick Moor into the broad and deep valley head between Harehope Hill and Eglinghammoor. A conspicuous meltwater channel cut into the deep col at 500 feet between Bewick Hill and Harehope Hill demonstrates that some fluvioglacial drainage from the Hedgeley Basin moved eastwards into this area. Concomitant deposits appear to be absent from the central section of this part of the valley, although the unknown depth of peat is perhaps largely responsible for the flat nature of the ground. The peat-flat is terminated abruptly, however, along its southern margin where a steep, ice-contact face of fluvioglacial deposits rises 35 feet above it. An impressive terrace, 480 yards long and 480 yards broad at its maximal extent, occupies much of the valley floor at this point. A poor exposure in the slumped sides of a disused gravel pit reveals that water-worn cobbles and gravel are interbedded with considerable thicknesses of sand. Steep ice-contact slopes bound the formation on three sides and a prominent kettle hole on its southern margin appears to have been used for the site of a mill dam. There are slight irregularities on the surface, but its overall form is relatively flat and it may be interpreted as a kame terrace. Apparently associated with this kame terrace are two massive eskers which diverge round a linear dead-ice hollow aligned parallel with them. Both eskers slope southwards and were presumably deposited by meltwater streams flowing in that direction. This assemblage of fluvioglacial materials was clearly deposited in and around stagnant glacier ice lying in the broad valley about Harehope, and appear to be related to a period when considerable volumes of water were discharged down the upper Eglingham valley.

The broad valley-head west of the Eglingham col contains numerous eskers aligned in a west-east direction (Map 8). Their crests undulate gently, but are normally less than 25 feet high. Dead-ice hollows and a few small kettle holes are associated with these formations which are probably continuations of the meltwater drainage system represented by kame terraces and related ridges north-west of Beanley. They all demonstrate the movement of meltwater in an easterly direction, towards the col at Eglingham, and the uphill gradient shown by most of the eskers suggests that deposition occurred in a subglacial environment. Alternatively, they were superimposed from englacial or supraglacial positions. The wide col at Eglingham is veneered by an unknown depth of till arranged in broad, drumlinoid swells. The esker system leads towards one of the low points through the col, and although it terminates shortly before the col crest (which is devoid of fluvioglacial features), the drainage line is continued by the meltwater channel beginning immediately to the east. The channel is merely a dry, shallow depression only 20 yards wide over the first 300 yards of its length, but thereafter becomes a steep-sided gorge over 15 feet deep. It winds directly downslope before turning sharply at right-angles to flow south-eastwards between massive ridges of sand and gravel lying in the broad valley south of Eglingham. An old borehole on the eastern flank of the deposits proved 15 feet of interbedded sand and gravel (records of the Geological Survey). South of Eglingham village (Map 9), these rather amorphous masses of sand and gravel terminate very suddenly with steep fronts and the meltwater channel fades away at precisely the same height, slightly below 300 feet. Beyond this point, the pre-existing valley is clothed from side to side with a remarkable expanse of drift, the plane surface of which slopes very gently towards the south-east, from slightly below 300 feet to approximately 240 feet in the vicinity of the road between East Bolton and Shipley Lane, where it fades away. Apart from several fields of improved pasture the surface of this flat is covered by a continuous layer of peat, proved by augering to vary in thickness from 9 inches to 1 foot 6 inches. Augering to a depth of 3 feet also revealed the presence of grey silt, sporadic clay lenses and sparse gravel layers immediately below the peat, but this superficial depth of penetration is not considered wholly satisfactory. Nevertheless, this evidence suggests that a fluvial deposit underlies the peat overburden, and the relatively consistent depth of the latter indicates that these fluvial deposits form a gently sloping plain. Despite considerable slumping along the banks of the Eglingham Burn, there are no exposures to provide information about the underlying materials; but on the right bank of a small tributary incised into the plain, over 2 feet of fine gravel is exposed, and a lower exposure adjacent reveals tenacious till. In the gorge-like lower section of this tributary, bedrock is exposed to within 10 feet of the surface of the plain. It may be concluded from this evidence that in some places, at least, only a shallow depth of fluvial deposits underlies the peat. Towards the Eglingham Burn, however, a much greater depth of drift is implied by the 60-foot gorge, and the present stream may well be re-excavating a buried channel.

The regularity of this wide expanse of near-flat ground is interrupted south of Bannamoor farm by vague mounds of gravel that gradually resolve themselves into a prominent esker extending in a south-easterly direction for almost 1,100 yards. An irregular crest-line and relatively gentle side slopes characterise this ridge, which rises to a maximum height of 20 feet. Superficial exposures and two auger bores prove that it is composed predominantly of sand with some water-worn gravel. The adjacent hollow, in which Kimmer Lough lies, is probably a kettle hole. Discontinuous terrace fragments at a similar elevation to that of the wide flat occur on the north-eastern side of the Eglingham Burn. Separated and indented by dead-ice hollows, they have been partially degraded by stream erosion. From Bannamoor, they extend south-eastwards as far as the point where they terminate against a complex of mounds and ridges. Augering proved that silt and sandy clay underlie these terraces and they are considered to be part of the same formation as the main surface. The morphology and nature of deposits south-east of these terraces indicate that fluvioglacial activity associated with the Shipley valley was responsible for their formation, so that they are quite distinct from those by Bannamoor and will be considered in a subsequent section of this chapter.

From the foregoing evidence, it may be concluded that the sequence of fluvioglacial landforms converging on the Eglingham col clearly demonstrates a former flow of glacial meltwater from Bewick Moor and the Hedgeley Basin, and it must therefore be linked with the northern lobe of glacier ice from the Tweed valley. The eskers and channels appear to be mainly subglacial in origin, but the kame terrace and massive ridges at Harehope could have been deposited in an open, ice-walled environment. South of Eglingham, these hummocky landforms terminate abruptly, and a gently sloping surface, underlain by gravels, silts and clays, extends continuously beyond them in a south-easterly direction for approximately 2 miles; this surface is up to 900 yards wide, south-west of the Eglingham Burn. The prominent esker and kettle hole south of Bannamoor, and the small group of deposits north-east of that farm, are interpreted as earlier formations, partially smothered by the thin layer of later sediment; they are probably part of the subglacial drainage system from the Bewick and Hedgeley areas. As deglaciation proceeded, it seems that a certain degree of backwasting of the northern ice lobe was important in the subsequent development of landforms in the Eglingham valley. The abrupt termination of ice-contact forms south of Eglingham village apparently marks the approximate position of an ice-margin that remained stationary long enough for a proglacial outwash plain to be spread beyond it. These deposits partly smothered earlier ice-contact forms in their path and probably destroyed or buried others. The presence of clay lenses in the stratigraphy suggests that proglacial deposition was occasionally in temporary lakes or ephemeral lakelets scattered over the area. Since the outwash plain slopes from slightly below 300 feet to approximately 230 feet, it may be related to the level of meltwater drainage that controlled deposition in the upper and mid Aln valley.

The Shipley Valley

The sequence of south-easterly directed fluvioglacial landforms in the Eglingham valley terminates at approximately one quarter of a mile south-

Photograph 5.e Hollingsheugh Hill and adjacent eskers (Map 9).

east of Shipleymoor Farm (Sm. F., Map 9), where a belt of similar forms aligned with a different orientation marks their limit. The latter deposits emerge from the Shipley valley and are related to a system of south-westerly directed meltwater drainage that spread across lower reaches of the Eglingham valley.

The wide, pre-existing valleys of Eglingham and Shipley are separated by a broad spur of land sloping south-eastwards down the Fell Sandstone dip slope from 600 feet to 300 feet. The upper Shipley valley is relatively devoid of meltwater deposits, but, in contrast, the lower section is rich in such formations. They appear to have been carried into the valley from the north-east, and are consequently linked with deglaciation in the coastal province of north Northumberland, outside the scope of the present study. That region has recently been investigated in considerable detail, however, by Parsons (1966), and his accurate descriptions allow important correlations to be made with the east Cheviot area.

In the vicinity of South Charlton, a broad col occurs on the eastern side of the Shipley valley, and, from the north-east, a discontinuous, but prominent esker can be traced for over half a mile, leading uphill towards the col. Close to the col, deposition is superseded by erosion, so that an impressive rock-cut channel, 50 feet deep, has been cut through the crest. The channel floor has an up/down profile and rises from slightly below 350 feet at the intake to a maximum of 360 feet, mid-way through. A short distance beyond the outlet, an interconnected system of five eskers spreads out in a fan-like manner (Photograph 5.e). Parsons (1966) noted that "The amplitude of the individual esker ridges varies considerably, but is generally greatest where the features change direction and at points where the distributaries branch away from the main ridge." Steep ice-contact slopes are also characteristic, and kettle-holes and dead-ice hollows form intervening depressions. No sections are present, but minor exposures and superficial debris consist predominantly of rounded gravel in a sandy matrix. This small esker complex is sharply bounded on the west by a massive accumulation of fluvio-glacial material that forms the broad-crested ridge called Hollinsheugh Hill (H.H., Map 9). A section in a small sand pit at the northern end of this feature has revealed the following:

24 inches	stony till.
24 inches	horizontally bedded sand.
12 inches	silt with rudimentary bedding, dipping towards the north.
18 inches	false-bedded sand indicating deposition from the east and north-east.
3 inches	horizontal band of sandy silt.
36 inches	unbedded sand.

No other sections are available, but the steep ice-contact slope descends from 450 feet to 325 feet, and minor exposures illustrate the sandy nature of materials underlying this slope. The banks of the stream, between 300 and 325 feet, are cut in till, so that up to 125 feet of fluvioglacial material may well underlie Hollinsheugh Hill. Parsons observed, "In the northern part of the complex the esker ridges lead directly into the eastern flank of Hollinsheugh Hill, indicating a common origin of deposition by meltwater from the east and north-east." He considered the huge ridge of Hollinsheugh Hill to be a different type of formation, however, and suggested that it "may well have accumulated along a local ice margin". The precise meaning of this statement is rather obscure, for it is not clear if Parsons is implying a kame terrace type of formation or a proglacial fan-like accumulation. However, he continued to discuss an alternative theory, according to which, "the whole

Longlee Moor complex may have accumulated within extensively crevassed ice in an advanced state of decay, the deposits of Hollinsheugh Hill collecting in a major crevasse extending approximately south-south-eastwards from the leeward side (with respect to the former direction of ice advance) of the high ground north of the col near South Charlton."

This sequence of events in which eskers lead uphill to a col where erosion supervened, followed by renewed esker deposits in fan formation has also been observed by the writer near the Burn o'Vat in the Dee valley, Aberdeenshire. At South Charlton, the uphill gradient of the proximal esker and the up/down profile of the col channel indicate subglacial drainage of meltwater under hydrostatic pressure. The possibility that Hollinsheugh Hill was also formed in such an environment is suggested as an alternative to Parsons' theory.

In the lower Shipley valley, beginning a short distance downstream from the esker complex, fluvioglacial sands and gravels are banked as discontinuous terraces on either side of the stream. Parsons noted that the terrace on the eastern side is higher in absolute altitude than that on the western side, and presumed it to be earlier in age. The terrace surfaces decline towards the south, although a subsidiary element of slope inclines towards the valley centre. For over a distance of one mile, the eastern terrace can be traced from 352 feet to its abrupt termination at 305 feet. There are no sections to disclose the exact nature of underlying materials, but minor exposures reveal sand and gravel. The western terrace begins as a very narrow strip of ground in the lee of a projecting spur of bedrock, but ultimately attains much grander dimensions, becoming up to 500 yards in width. The surface of this feature falls southwards from 295 feet to 238 feet at Shipley. Parsons observed 5 feet of gravel underlain by 10 feet of sand at the northern extremity of the terrace, and these deposits rested directly upon bedrock.

Photograph 5.f Fluvioglacial sand and gravel containing coal layers, right bank of Shipley Burn.

Photograph 5.g Fluvioglacial outwash deposits, right bank of Shipley Burn.

Photograph 5.h   Ablation till overlying the fluvioglacial deposits shown in Photograph 5.g.

Three good sections have become exposed in this western terrace due to recent stream erosion (1, 2 and 3, Map 9). For the most part, the deposit consists of relatively fine-calibre fluvioglacial material, with bands of sand, silt and clay figuring prominently in the sequence. Fine to medium gravel is frequently interbedded with sandy layers, and bands made up entirely of coal fragments are unusual features of this deposit (Photograph 5.f). Cross-bedding is common to several layers, the dip of which, together with the general inclination of other layers, indicates deposition from the north-east (Photograph 5.g). The interstratification of these sediments of different calibre suggests that variable conditions of flow characterised the period of deposition - the close juxtaposition of clay and gravel layers is indeed a contrast. The absence of stones larger than about 6 inches is most significant in relation to an overlying deposit, in which numerous boulders over 12 inches in diameter are scattered throughout. The two deposits are evidently quite separate in origin. The uppermost deposit is extremely intriguing, for it consists of clay, packed with stones and overlies the sands and gravels in all three sections. It varies in thickness from 3½ feet to 7 feet. The fine matrix is predominantly a red/brown, silty clay that binds together an heterogeneous admixture of stones, quite unsorted and unbedded (Photograph 5.h). A small percentage show some degree of water action, but the majority are sub-angular. At section 3, the most striking characteristic of the deposit is the large number of stones that are either chemically rotted throughout, or else possess well-developed weathering rinds. They mostly consist of various types of sandstone, but some pieces of igneous material are present. Several boulders up to 12 inches in size were dug out during the process of excavating a site for fabric analysis, and a good number of blocks over 2 feet across lay in the uppermost part of the section. Stones of lithology conducive to the preservation of

Photograph 5.i Kame terrace and esker, near Shipley.

striations are well-striated. These characteristics strongly suggest that the material is till, and since up to 7 feet of it caps underlying fluvio-glacial deposits, in sites remote from steep hillsides, it is considered to be later in age and to be in situ. Beneath the fluvioglacial deposits in section 1, however, 25 feet of a compact grey-coloured till extends to the bank-foot. Similar till is exposed beneath fluvioglacial sands and gravels approximately 1,200 yards further upstream, but the total depth of this deposit is unknown.

The riverward margins of these terraces have been extensively modified in places by subsequent stream erosion and it is virtually impossible to assess whether or not the deposits were originally bounded by ice-contact slopes in the centre of the valley. Parsons maintained that they are kame terraces, but the absence of kettle holes and ice-contact slopes must make this interpretation rather tentative. Indeed, the terraces could equally well have extended across the entire valley prior to dissection. Prominent ice-contact formations ultimately appear in the lower part of the valley, however, beginning in the col south of White House Hill. In this locality, a narrow terrace lies at a similar altitude to the broad western terrace on the opposite side of the Shipley valley and occurs in association with a kettle hole and a small channel that apparently led meltwaters over into Hinding Dean. Strong evidence that terrace deposition occurred in association with stagnant ice is present on the western margin of the terrace. The terrace itself is bounded on that side by a steep ice-contact slope, gently sinuous in outline. It descends to a broad elongated kettle hole of irregular outline, in which a prominent esker winds generally in a south-westerly direction (Photograph 5.i). The uneven crest of this ridge reaches a maximal height of 30 feet and has been breached at one point by meltwater escaping from an adjacent block of wasting ice. Towards the distal extremity, the crest of the esker climbs steadily uphill as it merges into the side of the massive terrace deposit on the west. The esker crest lies approximately 100 feet above the floor of the large kettle hole at this point. Two smaller kettle holes pit the esker near this extremity. West of the esker occurs a gently undulating terrace from which three short ridges project westwards between dead-ice hollows and terminate above the Eglingham outwash plain.

There is therefore considerable evidence that the Shipley terraces and related formations were deposited in an environment dominated by large masses of stagnant ice that ultimately melted out to form kettle holes and dead-ice hollows. The meltwater which furnished these vast quantities of fluvioglacial deposits flowed chiefly down the Shipley valley but subsidiary amounts possibly drained down the adjacent Smallburns depression. Conditions of drainage varied, so that periods of augmented flow, during which current-bedded sands and gravels were deposited, alternated with periods of very gentle or reduced flow, when fine silt and clay particles were able to settle out. The capping of till observed to overlie these deposits at numerous places, is an interesting formation. Although the material exposed in section 1 could possibly have been derived from adjacent hillsides by solifluction, it is elsewhere undeniably in situ. There are therefore two possible explanations of its position overlying fluvioglacial sediments.

(1) It represents the true ground moraine of a glacier readvance over earlier fluvioglacial deposits.

(2) It is an ablation moraine that sludged down from crevasses or became superimposed from downwasting ice onto subglacially deposited sands and gravels.

Although the contact between these two deposits is not clearly exposed in every section, it was observed in 1 and 3 that the layer of fluvioglacial sediment immediately below the till showed no evidence of disturbance by moving ice, and no wisps, lenses or masses of these deposits have been caught up in the overlying till. Furthermore, it is considered that the prominent esker and adjacent ice-contact forms south of Shipleylane are unlikely to have survived with such a high degree of preservation had they been overwhelmed by moving glacier ice. The kettle holes in particular do not appear to have been masked by any subsequent deposit of ground moraine. While it might be argued that these points are not conclusive, there is a distinct lack of evidence for an important readvance in adjacent areas in Northumberland. The following explanation is considered more acceptable. In the lower Shipley valley the deposition of fluvioglacial sediments occurred in an environment dominated by the presence of stagnant glacier ice, while only one mile farther upstream, the col channel and esker system at South Charlton were probably being formed subglacially under a substantial cover of ice. Such evidence reconstructs the former presence of the highly crevassed and decayed margin of a lowland glacier, downwasting and receding over rather irregular topography. A situation such as this frequently gives rise to subglacial formations in the sub-marginal zone closely linked to subaerial and subglacial deposits in the fragmented marginal zone. The surface and interior of such piedmont glaciers, like the present Malaspina glacier, often contain variable thicknesses of chaotic debris that may be composed largely of heterogeneous rubble and clay. This debris can sludge down into adjacent crevasses, where it may sporadically cover fluvioglacial sediments. The ablation debris can also be lowered by the gradual melting of underlying and contained ice, so that it eventually meets the surface below - bedrock, ground moraine or fluvioglacial deposits. The layer of till overlying fluvioglacial sediments in the Shipley area is therefore interpreted as ablation material and it is considered valuable evidence in supporting the theory that a highly crevassed piedmont glacier at one time receded northwards from this area.

It is thus concluded that the main zone of compact glacier ice at this time lay in the vicinity of South Charlton, where the interconnected esker system was deposited beneath the glacier farther down-valley, a considerable volume of fluvioglacial sediment was deposited on, within and beneath extensively crevassed ice. Up to 7 feet of till is known to overlie the latter deposits in places, and this may have been derived from the surface and upper layers of the stagnant glacier margin. These conclusions imply that further to the south-west, the ground was ice-free at this time and constituted the proglacial zone. A discussion of landforms in this area is therefore necessary to conclude the reconstruction of deglaciation.

The Eglingham-Aln Enclave

The general surface level of deposits south-west of the Eglingham Burn is clearly a continuation of that attained by deposits on the opposite bank (Map 9). Directly across from the Shipley esker, two broad, amorphous mounds apparently continue the same alignment of fluvioglacial drainage for a short distance. Superficial exposures and material revealed in a ploughed field indicate that underlying drift is chiefly sand and water-worn gravel. Although the ridge form is reasonably plain, it lacks the sharp slopes normally associated with ice-contact forms, so clearly exhibited by the Shipley esker. Indeed, considerable modification subsequent to deposition is suggested by these relatively obscured outlines. Downstream from the ridges, the uppermost part of the river bank is indented by numerous embayments and dry, gully-like depressions. The irregular plan of these and their close proximity to large kettle holes and dead-ice hollows, suggest that they too are dead-ice hollows. To the south-west of the ridges and dead-ice hollows, a remarkably continuous surface extends as far as Midstead, and for the most part, declines gently towards the south-west. Continuity of this surface is broken at four places, 4, 5, 6 and 7, Map 9. The prominent kettle hole, 4, is 15 feet deep and apparently contributed some water to the meltwater stream that cut the channel at 5. The meltwater channel at 6 is aligned at right-angles to the former, so that two fragments of the terrace have become isolated. The trend of channel 6 indicates that it was cut by water flowing roughly from south-west to north-east down the Aln valley, after the period of terrace construction, but before channel 5 was formed for it is truncated by the latter. The feature at 7 is a wide channel incised some 25 to 30 feet below the terrace surface near its distal extremity. The present trickle of water in the floor of this channel, even following prolonged rainfall, scarcely emerges above the tangle of vegetation that obscures it. The terrace surface on either side of channel 7 begins to slope more steeply as it descends into the Aln valley, until it ultimately peters away slightly above 200 feet. There are no sections in this terrace deposit, and only minor exposures and the nature of ploughed fields indicate the presence of sand and gravel, but this evidence in conjunction with the kettle hole, dead-ice hollows and meltwater channels, is considered sufficient to indicate that this sloping terrace is of fluvioglacial origin. It is difficult to assess the total thickness of these deposits, but nowhere does their surface descend below 200 feet. The bedrock exposed in channel 5 outcrops up to 200 feet, and numerous sections at and below 200 feet in channel 7 reveal till. Since much of the terrace surface lies above 225 feet, it is possible that at least 25 feet of sands and gravels were deposited on top of bedrock or till.

The following conclusions may be drawn from the foregoing evidence.

(1) Ridges and dead-ice hollows on the terrace edge south-west of the Eglingham Burn are related to similar formations on the opposite bank and testify to fluvioglacial deposition on contact with stagnant glacier ice. The large kettle hole at 4 may be considered as an outlier of this dead-ice environment, since no other ice-contact forms occur beyond it. The broad terrace surface slopes south-westwards from these ridges and dead-ice hollows, and it is believed that the amorphous outlines of the ridges may be in part due to modification and partial covering by the spread of these terrace deposits. Opposite Shipley farm, this terrace surface is plainly a continuation of similar features extending from the Shipley valley; a spot height of 238 feet beside Shipley and one at 237 feet on the cross-valley continuation, testify to this relationship. In view of the marked absence of ice-contact forms beyond point 4, and the broad extent of the terrace, these deposits are interpreted as proglacial outwash that issued from stagnant and crevassed glacier ice at Shipley.

(2) The conspicuous channel at 5 may represent a former outlet for proglacial meltwater from the Shipley area, but its alignment virtually at right-angles to the dominant slope of the terrace suggests that the eroding waters could have been derived from another source. In this respect, it must be realised that the extensive build-up of these outwash deposits occurred across the mouth of the Eglingham valley, down which considerable meltwater drainage from the north-west was possibly flowing simultaneously. Since the Eglingham deposits lie at a slightly lower level than those at Shipley, it may be considered that Eglingham meltwater probably became temporarily ponded from time to time. Clay layers in the Eglingham deposits possibly date from such occasions. While the ultimate escape of Eglingham drainage appears to have been along the course subsequently occupied by the present stream, the channel

at 5 may represent a former outlet cut by proglacial meltwater from the Eglingham valley.

(3) Although the Shipley outwash terrace indicates meltwater flow south-west towards the mid Aln basin, the only ultimate outlet for water from that area was down the Aln valley in an easterly direction. The feature at 6 is therefore interpreted as a channel cut by water establishing a route in this direction. It clearly post-dates terrace formation, but was itself truncated by a later phase of drainage possibly from the Eglingham valley.

Conclusion

The manner in which glacier ice receded from low ground peripheral to the south-east Cheviots tended not to produce abundant ice-contact landforms of fluvioglacial deposition. This is in great contrast to events that occurred simultaneously to the north, where vast systems of eskers, kame terraces and kettle holes dominate the lower hillslopes and adjacent basins. Since the direction of former ice movement in the south-east Cheviots approximately coincides with pre-existing valley alignment, most of the meltwater drainage ultimately flowed down these valleys. The relative absence of impediments to the efficient evacuation of water and debris may be an important factor in accounting for the anomalous absence of ice-contact forms in the Breamish and upper Aln valleys and in other valleys adjacent to the south. Perhaps of equal importance was the rather unique glaciological situation in which prominent lobes of glacier ice were apparently confluent in the axis of ground from Ingram to Whittingham. The marginal zones of these glaciers presumably responded promptly to climatic amelioration and a considerable amount of horizontal recession, in addition to vertical downwastage, is to be expected. Thus, as Breamish ice withdrew from the Ingram area, the proglacial drainage that became established down the lower Breamish valley was eventually responsible for the Ingram gravel plain. Similarly, as the great western lobe parted from its junction with the expanded Tweed glacier, in the vicinity of Whittingham, considerable quantities of proglacial outwash were deposited in the form of terraces. Meanwhile, northern ice was not only stagnating over the Hedgeley Basin and the coastal province east of the Fell Sandstone ridge, but also marginal recession enabled the deposition of conspicuous outwash trains from Shawdon Dean, the Eglingham col and the lower Shipley valley. Since the mid Aln basin became ice-free at a relatively early stage, it became the site of a glacial lake in which laminated sediments accumulated to a level of at least 200 feet. The delta-like deposition of outwash sands and gravels in the Whittingham and Shawdon area was probably in association with this lake, and since the Eglingham and Shipley terraces slope to terminal heights of slightly above 200 feet, they too, may have been related to a level of drainage linked with the presence of the mid Aln glacial lake. The ultimate escape of water from this lake must have been down the lower Aln valley, for there is no other outlet below 325 feet. Stagnant remnants of the northern ice lobe possibly lay across this route and were perhaps partly responsible for the impediment of drainage that gave rise to the mid Aln glacial lake, but the extensive masses of outwash debris were probably the most effective barriers. Finally, normal drainage became established along this route as the lake eventually drained, and it is presently occupied by the river Aln.

The widespread occurrence of terrace surfaces at approximately 200 feet suggests that the outlet level for drainage from the mid Aln region lay at that height for a considerable length of time. Since the lower Aln valley lies outside the scope of this thesis, detailed mapping was not undertaken in that area. However, Parsons (1966) mapped and described a large meltwater channel near Warkworth station (4 miles south-east of Alnarck). The channel is aligned in a southerly direction and extends for a distance of $1\frac{3}{4}$ miles. This conspicuous feature, whose intake lies just below 200 feet probably functioned as a major routeway for drainage issuing down the Aln valley as the water became diverted southwards beneath decaying ice occupying the coastal province of Northumberland. As such, it possibly determined the surface level of the mid Aln lake and consequently the upper level of deposition in that lake.

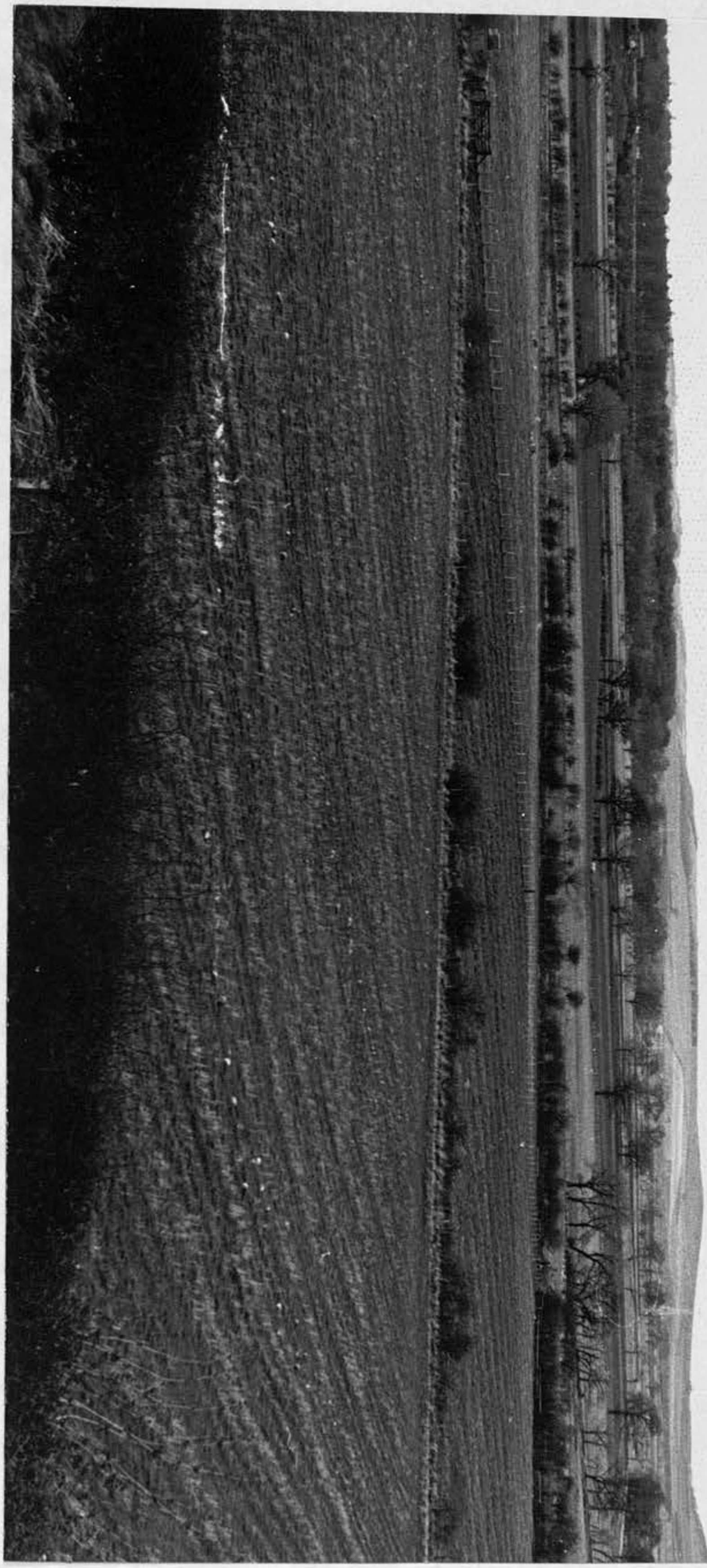
CHAPTER 6.

FLUVIOGLACIAL PHENOMENA IN THE MILFIELD BASIN AND ADJACENT AREAS

Introduction

To the north of Wooler extends a prominent tract of low-lying country called the Milfield Basin (Map 2). Lying mostly below 200 feet, it covers an area of approximately 12 square miles. The Milfield Basin is the largest member of the series of Cementstone basins that constitutes the sub-Cheviot depression and is contained between the steep edge of the volcanic massif on the west and the precipitous, west-facing escarpments of Fell Sandstone on the east. Its much wider extent in comparison with adjacent basins, is due mainly to the sharp westward curve assumed by the volcanic massif at Wooler. The steep hill-front overlooking the basin apparently led the Geological Survey to infer the presence of a major fault in this vicinity, but such topography could equally well have developed through weathering and slope processes under bio-climatic environments different from that of the present; a modest amount of glacial erosion may also have been involved. Whether or not faulting was responsible, it is certainly the case that a marked re-entrant interrupts the sweep of the massif four miles north-west of Wooler and accommodates the lower section of the Glen valley. The bedrock floor of the Milfield Basin is entirely obscured beneath thick deposits of glacial drift, including till, laminated silts and clays and stratified sands and gravels. The maximum thickness of these deposits remains unknown, but at least 145 feet is present in one place. The bore-hole that recorded this depth of material reached almost down to sea-level and terminated before rock-head was encountered, and so a most unusual depression must underlie the drift cover. Common (1953) favoured "ice-gouging and overdeepening" in explanation of the Milfield depression. Since the adjacent

Photograph 6.a The Milfield lake basin. Trees on left skyline are on the outwash delta emerging from the Glen valley. Viewed from the south-west.



escarpments of Carboniferous rocks appear to have been considerably scoured and streamlined by glacial erosion (Map 3), ice movement over this particular area may well have been sufficiently active to have severely eroded the soft Cementstone rocks of the Milfield Basin. The river Till winds slowly across the basin, in which it is joined by two major tributaries, the Wooler Water and the river Glen. All three streams meander considerably over this relatively flat country and have been confined between artificial levees in an attempt to avoid serious flooding. The central part of the basin, where the rivers Till and Glen meet, lies slightly above 100 feet O.D. From this locality the ground rises very gently, almost imperceptibly, towards the south and east until the margin of the basin is reached at between 150 and 200 feet. To the west and north the gentle slopes that characterise the floor of the basin are bounded somewhat abruptly by relatively steep slopes rising 20 to 25 feet up to a broad terrace surface (Photograph 6.a). This extensive feature emerges distinctly from the Glen valley, in which its apex begins at 225 feet, but the greatest expanse lies out in the Milfield Basin where it extends chiefly towards the north on both sides of the river Till as far as its termination at Etal. A narrow, rather fragmented, strip of the same deposit fringes the edge of the volcanic massif on the south side of the Glen west of Akeld. Numerous exposures revealing laminated silt and clay occur on the floor of the basin and, together with those exposing sand and gravel underlying the terrace, have given rise to considerable speculation in the literature regarding their mode of origin. A brief summary of salient points in the literature is therefore relevant at this stage.

The first description of deposits in the Milfield Basin was presented by Milne Home in 1876. The occurrence of sands and gravels, mainly overlying, but occasionally interbedded with stoneless clay, led him to believe that the

Milfield Basin had formerly been the site of an extensive lake. A much fuller account of the area was written by Gunn (1895) who observed many sections and exposures and acquired information from various local inhabitants. He fully agreed that the deposits represented those of a former lake and believed that it "was probably formed about the close of the Glacial period". He also considered that "The thick deposits of clay in the old lake may have been to a great extent derived from the melting of the ice at the close of the Glacial period." It is therefore clearly apparent that Gunn connected these deposits with the close proximity, or, at least, the recent presence of glacier ice. The level reached by this old lake was not stated, but he inferred that it must have risen above 180 feet; e.g. "At Nesbit there seems a good-sized patch of Boulder Clay which must, however, have been covered by the waters of the lake when they stood above 180 feet", and later, "It would seem, therefore, that the waters must have covered a good deal of the low sandy ground below 200 feet ..." The evidence on which he based these assumptions is by no means clear, but is likely to have been the 185-foot level at which the associated terrace lies at Lanton in the Glen valley, Apparently in agreement with Milne Home, Gunn believed that the lake waters were dammed back by the moundy deposits of sand and gravel about Crookham and Etal. The next reference to "Lake Ewart" followed several years later when Butler (1907) somewhat imaginatively described a hypothetical journey across the lake "in a Neolithic boat". He reconstructed a situation, in which he envisaged the present Till valley blocked between Pallinsburn and Duddo, so that "Till, disappointed of an outway by Black Bank gorge ..... would swell with anger and fill his valley from side to side till his fulness reached the contour line of 200 feet"; he continued, "Were the barriers at the Black Bank and Pallinsburn gorges to be 220 feet high, the outline of Lake Ewart would not be very appreciably altered, and the Till would then drain away the overflow by passing between Mattilees Hill and Gatherick, where there is a line of minimum elevation not rising higher than 217 feet above sea-level, ... once past this, there would be a downhill slope to the head of Haiden Dean." Haydon Dean is a prominent meltwater channel cutting through the low col east of Duddo. The intake begins near the village at slightly above 225 feet, and from there, the channel may be traced continuously in an easterly direction to Ancroft, where it terminates at 75 feet. Butler clearly interpreted this feature as a lake overflow channel, the then current explanation for such dry gorges following Kendall's work in the Cheviot and Cleveland Hills, and it appears to be his only evidence in support of a former lake level at 220 feet. Considering the origin of Lake Ewart, Butler surmised that "If ..... the Tweed Valley was occupied by a wide glacier, spreading over the river bank on both sides, this would effectually block the direct avenues to the Tweed from Pallinsburn or Black Burn, three miles distant, either with its moraine, or with its own solid ice. So long as this glacial condition lasted, the Till would flow down Haiden Dean ....." It is unfortunate that satisfactory evidence was not presented to corroborate this theory. While the position of Haydon Dean suggested a maximum level for Butler's Lake Ewart, this channel was not his only evidence of the former existence of the lake, because he continued, "There are some independent evidences of Lake Ewart which may be given. The existence of beds of clay, when not of glacial origin, that is, not intermixed with gravel and boulders, but pure, deep, and continuous, points to deposition in still lake water of the finer sediment brought down by rivers." He endorsed this statement with reference to sites in the Milfield Basin where such deposits had been proved, and it must also be recalled that Milne Home and Gunn had earlier described the numerous sections on which they based their Lake Ewart theories. In addition to Lake Ewart, Butler suggested that a smaller lake formed at a later date near Kirknewton, in the Glen valley; he accounted for this in the following manner: "As the Tweed glacier may have dammed the Till, so may an Akeld glacier, coming right across the valley to the high ground of Akeld Steads opposite, have dammed the Glen till a lake was formed, surface 170 to 180 feet above sea-level ....." His sole reason for suggesting a lake at this level was to account for an anomalous dry valley that winds across the terrace surface from Coupland to Thirlings, north of the Glen. A somewhat indirect reference was made to Lake Ewart by Burnett in 1927, arising from his investigation of the Chatton Basin, out of which the Till flows into the Milfield Basin. In the vicinity of Chillingham village, he observed "true laminated gutta-percha clays", which "appear to have been deposited in the headwaters of the great lake which extended from Chillingham, by Chatton and Wooler, to Etal, where the water was impounded by a barrier of kettled gravels and dead ice". He described traces of "the lake terrace, running just below the 200-foot contour" and plainly assumed this to be the highest level reached by the ponded waters. However, the exact nature and origin of the lake terraces observed by Burnett are not clear; he even admitted that "they are not easily seen at close quarters", and did not indicate if they were gravel terraces or terraces cut into the ground against which the shores of the lake abutted. The most recent appreciation of the Milfield Basin was that by Common (1953) who added few observations to those by previous workers. He did not discuss the lake concept in detail, but apparently accepted that two higher water levels formerly existed, for he concluded, "the 140' lake in Milfield and also at a higher level c.200' may be post glacial in origin, dating from a period when the lower Tweed was flooded and temporarily estuarine".

The foregoing discussion illustrates that a considerable volume of literature already exists on deposits occurring in the Milfield Basin. It is generally agreed that these are lake deposits, but there is a distinct conflict of opinion on the origin and extent of the lake. Briefly, these are as follows:

(1) Milne Home and Gunn inferred that the lake was post-glacial in age.

(2) Butler and Burnett suggested that Lake Ewart formed during a late stage in the wastage of the Tweed glacier.

(3) The same memoir in which Burnett's views are expressed contains a brief summary on the sequence of glacial events by Carruthers (1927). He firmly believed that the lake surface lay at 140 feet and that the water was ponded by the "Cornhill kettle-moraine". A precise age for the lake was not suggested, but since he evidently supposed that this moraine had originally extended much further and was subsequently obliterated in the Milfield Basin by the action of the Till, Glen and Wooler Waters, a post-glacial age may be inferred.

(4) Common's views on the origin of the lake are somewhat obscure, and without any supporting evidence, he referred the lake to a post-glacial period when the lower Tweed valley was rather mysteriously "flooded and temporarily estuarine".

Since the sequence of events responsible for deposits and landforms in the Milfield Basin and adjacent areas is by no means clear from the foregoing hypotheses, it is necessary to (1) reconsider the evidence from exposures referred to in previous literature and to include information from sections observed by the writer, and (2) to re-interpret the associated landforms, following the detailed mapping of them.

The Laminated Clays and Silts

Since former railway cuttings, quarries and pits are now completely overgrown, a careful investigation was made of every available exposure along the banks of all streams leading into the river Till. Information yielded by the numerous sections observed has been tabulated in the appendix. For convenience of description and reference, the relevant parts of the Till valley have been considered in two sections, the Chatton Basin and the Milfield Basin. With reference to the former (Map 6), present exposures and those reported in the literature demonstrate that laminated clays and silts occur over relatively wide areas below 200 feet, as Burnett (1927) earlier observed. Indeed, exposures above that level are not in such materials. Laminated sediments in the Chatton area consist primarily of red and grey stoneless clay, with occasional lenses of silt and fine sand. In a few sections, the upper layer sometimes contains small stones and is seldom laminated. This is considered to be a modified deposit produced chiefly by solifluction, in which laminated clay has become mixed with a gravel layer, or else soliflucted till from higher levels.

The Milfield Basin and valleys tributary to it (Maps 2 and 5) contain considerable thicknesses of silts and clays that were presumably laid down in standing or near-standing water. These sediments are extremely well laminated, breaking up into leaves along the silt layers, and for the most part, vary in colour between red, brown and grey/green. The total depth of these deposits has not been proven, but a borehole at Humbleton appears to have penetrated them for at least 70 feet, and at another place in the same vicinity, for at least 45 feet. It remains uncertain from the borehole evidence quoted in the Geological Survey Memoir that the clay referred to was entirely of the laminated deposit; it is conceivable that part of the bore was through relatively stoneless till. Sections revealing the nature of till in this area are seen only along the Wooler Water, and while the till is not densely packed with stones, it contains a considerable scattering of them and any borehole through this material is unlikely to give the impression that the clay is stoneless. The Milfield Basin may thus contain great thicknesses of laminated clays and silts over a wide area below 150 feet (such deposits have not been observed above that level). Those recorded by Gunn (1895) in the Glen valley occur at a similar elevation and are part of the same deposit, but laminated clays exposed along the Wooler Water could be considered as a separate formation for the following reasons: (a) they are less distinctly laminated; (b) they occur as sporadic outcrops and vary in colour and composition from one place to another within a distance of 600 yards; (c) they occur at between 200 and 225 feet, much higher than those elsewhere in the Milfield Basin. These reasons are by no means conclusive, however, and the clays could have accumulated in a shallow extension of the lake in the Wooler Water valley, where conditions of sedimentation would have been slightly different. Two of the borehole records for the Milfield Basin refer to thin seams of gravel contained in the stoneless clay.

It is therefore concluded that the Milfield and Chatton Basins were formerly occupied by an extensive lake in which floor deposits of laminated silts and clays accumulated to approximately 150 feet and 200 feet respectively; the surface possibly stood at a slightly higher level. The latter suggestion is perhaps endorsed by the occurrence of laminated clays in the Wooler Water valley at between 200 and 225 feet, but it is by no means certain that these deposits were formed in the same lake. Owing to the paucity and relatively poor quality of sections in the Chatton Basin, the total thickness of deposits there is not known, and it may be considerably in excess of the observed maximum of 15 feet. Such deposits do not always have a distinct topographic expression, but the somewhat featureless, gently sloping tracts of the Till valley north of

Newtown and south of Greendikes (Map 6), are considered to be partly the result of inundation by lake waters, with the concomitant deposits smoothing out any minor irregularities on the pre-existing slopes. It is significant that the pronounced break of slope above these gentle slopes occurs almost precisely at 200 feet. The central part of the Milfield Basin and the slopes below Doddington and Fenton Town (Map 2) are of a similar nature.

The Sands and Gravels

The Milfield Basin is relatively free of ice-contact fluvioglacial landforms such as eskers, kame terraces, kettle holes and dead-ice hollows. In view of the abundance of such forms in similar, adjacent basins, the situation in the Milfield Basin may be considered anomalous. Laminated sediments, consisting chiefly of clays and silts, underlie the surface of the low-lying eastern and southern parts of the basin, but a large expanse of the remaining area, stretching from the mouth of the Glen valley northwards to Etal, contains a considerable depth of sand and gravel. These deposits extend as a conspicuous terrace formation, partly degraded and fragmented, that slopes out radially from its apex in the Glen valley. From a maximum height of approximately 225 feet on the apex, this terrace can be traced with remarkable continuity to its northern extremity at Etal in the Till valley, where the feature terminates at approximately 125 feet. The terrace surface also slopes north-eastwards until it fades away at just below 125 feet near Fenton Town, but the continuity of this section is interrupted by the trench presently occupied by the river Till. Only a narrow strip of the terrace remains on the south side of the Glen valley. Although subsequent gullying by short streams flowing off the steep edge of the volcanic massif has partly fragmented this narrow strip of terrace, the surface clearly slopes south-eastwards to Akeld, where the last fragment lies at approximately 150 feet. Concerning the sand and gravel, only a limited amount of information can be obtained from sections presently exposed, but Gunn (1895) was able to record much more detail when he worked in this area. He was in no doubt that the sand and gravel directly overlie the laminated clays and silts previously described. For example, he observed 3 feet of moderately fine gravel on top of the clay at Flodden Tile Works, 7 feet of gravel over 20 feet of fine clay just to the north, and 2 to 4 feet of gravel over clay in the adjacent bank of the river Till. South of that area, Gunn noted the following; "the sand and gravel overlying the clay are thicker on the Ewart Estate than they are further north, this estate being mostly on sand, running sand, with stiffer sand and fine gravel in thin layers. There are occasionally thin seams of clay - grey, blue or red." Additional information obtained from the Ewart Estate by Gunn seems to indicate that not only does the sand and gravel deposit thin northwards, it also thins in a south-easterly direction. For example, a well near Ewart Park was sunk through 24 feet of sand, but at Ewart Brick and Tile Works the clay was capped by only 6 to 7 feet of sand. Sections presently exposed on the left bank of the Till downstream from Flodden Tile Works reveal 2 to 4 feet of fine gravel overlying laminated clay and silt, but exposures no longer occur in the vicinity of Ewart Park. It further emerges from Gunn's account that the clay disappears beneath an increasing thickness of sand and gravel when traced towards the apex of the terrace, for the last recorded exposure of clay was at the base of the river bank half a mile north-east of Coupland Castle. A section in the river bank adjacent to Coupland Castle consisted entirely of sands and interbedded gravel, the calibre of which became coarser towards the base. It is evident from these records that the deposits of sand and gravel are thicker and coarser near the proximal end of the immense terrace formation. The rather fragmentary evidence from present exposures tends to support this conclusion. For example, near the apex of the deposit, the river Glen has cut down through at least 25 feet of coarse gravels and cobbles; the stones vary in shape from sub-angular to sub-rounded, and fragments up to 12 inches in diameter are common. On the right bank of the Till approximately $3\frac{1}{4}$ miles from the previous locality a 15-foot section in the terrace consists predominantly of horizontally bedded sand of an extremely fine calibre.

The final observation that may be made on the nature of the terrace materials, is that an upper layer of small to medium sized gravel consistently appears over much of the terrace. For example, near Yeavering in the Glen valley, a shallow exposure shows sand overlain by a foot of medium-sized gravel. Gunn observed 4 to 5 feet of moderately coarse gravel at the top of a section, near Coupland Castle; he mentioned also "an old pit S.W. of Gale Wood, showing moderately fine gravel with some sand", and, "in a small stream 300 yards N.N.E. of Gale Wood:-

> 4 to 5 feet of well-washed and rounded gravel, resting on a denuded surface of fine light brown stratified running sand with some gravel, about 5 feet seen."

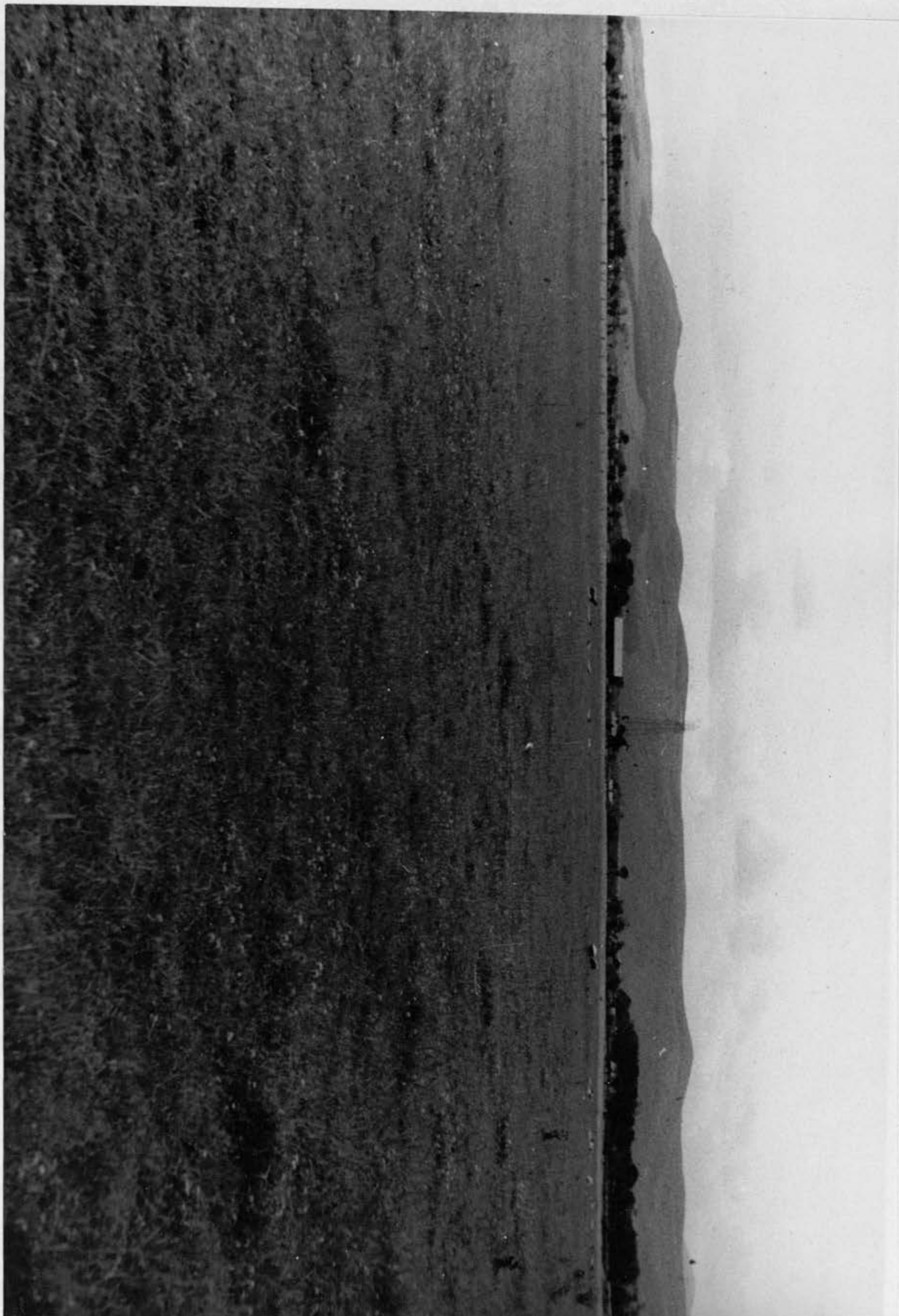
On the right bank of the Till at Milfieldford Plantation, a few feet of fine gravel on top of sand is presently exposed in a poor, slumped section. Farther downstream Gunn observed a pit close to the east bank of the Till, near the south end of Pheasant's Wood. This showed the following section:-

> Sand and sandy earth, 2 to 3 feet.
>
> Gravel, rather small, with thin irregular seams of coarse gravelly sand, 6 to 7 feet.

On the left bank of the river near Ford bridge he saw 10 to 12 feet of rather fine gravel, with a few inches of clayey sand about 5 feet from the top. Finally, it may be added that fine to medium gravel litters the majority of fields on the terrace. Shallow pits dug for a recent (1966) line of electric pylons near Bog Plantation all show at least 2 feet of fine gravel in a sandy matrix.

The earliest interpretation of these deposits was that put forward somewhat tentatively by Gunn when he wrote that "the underlying gravels and sands of the plain west of Coupland ..... must certainly be regarded as old delta deposits formed by the Glen as it entered the lake". The subsequent revision of this area by Carruthers (1932) provided no new information and his account in the Cheviot Memoir consists very largely of quotations from Gunn's earlier writings. Common (1953) casually accepted the terrace formation as a delta, but apart from noting the slope of its surface "to be symmetrically disposed on either side of a Coupland-Galewood line", he added no new information or theory with which to establish the genesis of the terrace. It is therefore evident that previous literature on the East Cheviot area does not provide a comprehensive interpretation of the Milfield sands and gravels. For this reason the following paragraphs present a detailed morphological description of the terrace, which, in combination with an analysis of its internal composition, points to an adequate interpretation of its origin.

In relation to the width of the valley it occupies, the river Glen flows through a relatively narrow, winding channel that is seldom more than 2 to 4 feet deep. Downstream to Akeld, the stream bed consists chiefly of cobbles and gravel, although small sand lenses are occasionally present, and these materials appear to have been derived from the terrace edges. In a downstream direction from Akeld, the modern alluvium is predominantly composed of sand and/or loam. Small patches of fine to medium sized gravel are sometimes present, but it is clear that the modern stream is incapable of transporting the coarser debris over which it flows farther upstream. In this respect, Photograph 6.b Looking over the Milfield outwash delta towards the steep, north-east flank of the Cheviot massif.



the calibre of materials present on the modern floodplain closely reflects that of materials in the adjacent terrace edge. Indeed, the volume and power of the present river Glen seems inadequate to account for the abundance of gravel and cobbles on the floodplain upstream from Akeld, and it seems probable that were it not for the presence of such materials in the undercut bluffs of the terrace, the modern floodplain would consist of quite different sediment. In view of this, and since the terrace surface lies from 15 to 30 feet above the river, it is difficult to account for the extent and composition of the Milfield terrace in terms of recent or current fluvial processes, as Gunn (1895) was well aware. The volume of water that flowed down the Glen valley during the stage of terrace deposition must certainly have been considerably larger than that of the present stream. Furthermore, the immense quantity of sand, gravel and cobbles appears to have been derived from a source no longer present, and the deposition of these materials was to a level up to 30 feet above that of the modern stream. In order to establish the sequence of events leading to the formation of the Milfield terrace, its extent and morphology, together with those of related landforms immediately adjacent, were mapped on the six-inch scale.

1. The Milfield Delta : Although the dominant characteristic of much of this terrace is its plane surface (Photograph 6.b), the uniform smoothness is broken in several places by distinct, dry channels (Map 2).

Channel A begins at the bank top above the river Glen east of Coupland. From modest dimensions only a few feet deep and wide, the channel soon becomes a very prominent dry valley, 15 feet deep and up to 250 yards wide, and describes a sharply defined sinuous course marked by undercut bends. Despite the winding nature of its course, channel A is aligned predominantly in a north-easterly direction and terminates abruptly on the side of the trench occupied by the river Till. At its terminus, channel A is approximately 20 feet deep and lies over 15 feet above the modern floodplain of the Till. An additional feature of this channel is the presence of narrow terrace fragments within its walls. They appear to be remnants of a former floor level of channel A, approximately 3 feet higher than the ultimate one.

Channel B begins abruptly on the terrace surface as a shallow depression, no more than 3 to 4 feet deep. Although it is a much smaller feature than the previous channel, the early part of its course winds north-eastwards similar to that of channel A. The lower reaches of channel B become up to 100 yards broad, and terminate 3 feet above the floor of channel A.

Channel C is less distinct than the two previous forms, but may be traced continuously as a broad depression running northwards from Lanton to Milfield. The form of this feature is extremely amorphous in the vicinity of Lanton, where it lies only 2 to 3 feet below the terrace level. Centuries of cultivation have probably destroyed much of the original form. North of Milfield, however, it is up to 9 feet below the terrace surface and ultimately fades on the surface of a terrace distinctly lower in level than the main one. The Sandy House Burn is clearly responsible for the narrow gully cut into the floor of channel C east of Milfield.

Channel D is located much further north than the previous channels and occurs near the distal extremity of the Milfield terrace. Although the channel begins as a shallow, flat-floored depression 170 yards wide, it eventually narrows and becomes a sharp, gorge-like feature almost 40 feet deep. Two dry tributary gullies enter at the point where channel D turns through 90 degrees to run generally northwards. The channel begins at 130 feet and apparently grades into the present floodplain of the Till, slightly below 100 feet.

The alignments of channels A, B and C indicate that they were cut by streams of water flowing out of the Glen valley, and were undoubtedly associated with the drainage system that deposited the sands and gravels of the Milfield terrace. A slight fall in the base level of erosion later caused channel A to erode 3 feet below its former floor, fragments of which remain as narrow terraces. Since channels A and C terminate at approximately 130 feet, they were probably associated with the same level of drainage immediately before they ceased to function. This drainage flowed northwards down the Till valley. Because channel D grades almost to the level of the present Till floodplain, it may be considered to have operated last of all, and presumably reflects the change in drainage conditions that terminated terrace deposition in the Till valley south of Etal.

In addition to the abandoned river channels, other irregularities are present on the Milfield terrace, interrupting the general smoothness of its plane surface. For example, in the Ewart area there is a small number of shallow depressions, but since these mostly occur near the former Brick Works, they are probably old clay pits. Nevertheless, the broader undulations are probably the result of shifting stream courses, possibly related to the forerunner of the Glen as it moved towards the position presently occupied by that river. Between Cat Corner and Warren Plantation (Ct.C. and W.P., Map 2) the terrace surface is very irregular. There is a relief amplitude of 10 feet at one place, but this may reflect the change in level from the main terrace to the lower one associated with channel C, mentioned above. Indeed, the zone of undulating topography appears to be confined to this lower level. For example, near Warren Plantation the terrace surface slopes down towards the river in a series of small ridges and depressions. Although the height difference between ridge crest and succeeding hollow is normally less than 5 feet, these undulations are well-marked features. It is suggested that they were formed by a stream (or perhaps a series of streams) which continually shifted its course as it adjusted to a lowering base level. It finally eroded the permanent channel now utilised by the river Till. That the various depressions just referred to are not associated with blocks of dead-ice is suggested by (a) the absence of definite ice-contact slopes normally present in kettle holes and dead-ice hollows, and (b) the pronounced linearity of the features in a direction parallel with the present drainage.

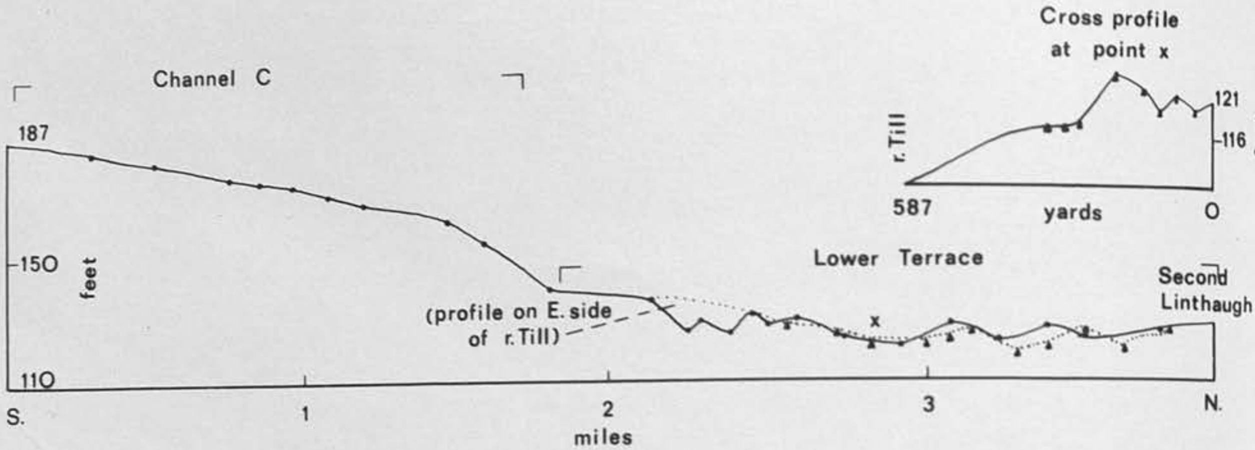
From the foregoing evidence, it may be concluded that the sands and gravels composing the Milfield terrace were deposited by a stream or series of streams that issued from the Glen valley and flowed eastwards, north-eastwards and northwards. Channels A, B and C appear to have functioned simultaneously for some time and the lower terrace level into which channel C merges is probably associated with this phase of drainage. Other irregularities on the terrace surface are relatively minor features, possibly produced by a braided stream network and/or the lateral migrations of the Glen and Till as they established themselves in their present channels following a pronounced fall in the base level of erosion. Since channel D grades almost to the level of the present floodplain, it appears to have been cut at a later stage in the drainage development, becoming ultimately abandoned when the trench now occupied by the Till became permanently fixed.

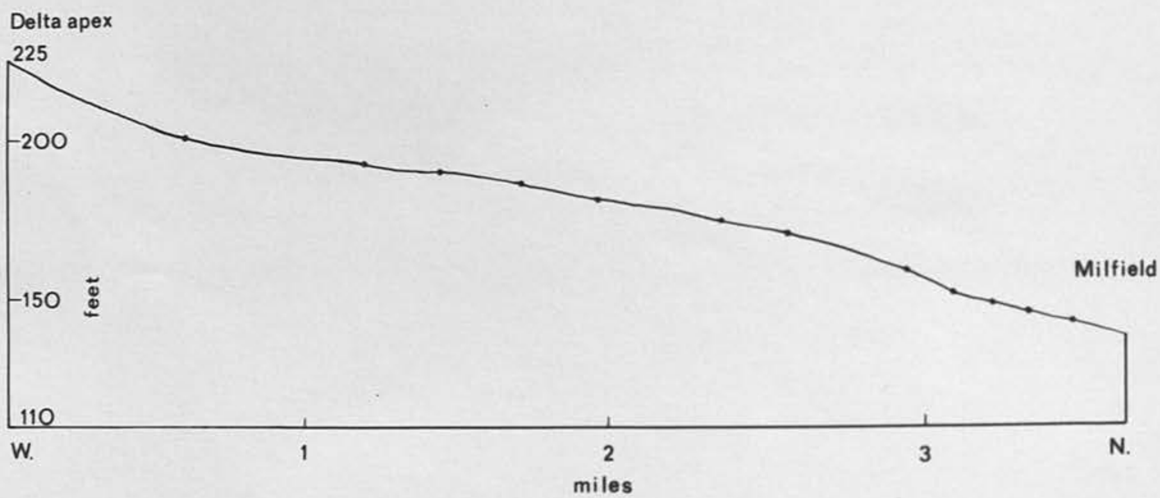
Having established that a system of drainage emerging from the Glen valley deposited and partly dissected the terraced sands and gravels, it is necessary to determine the areal extent of the deposition and the slope of the terrace surface. Although several of the Ordnance Survey's spot heights are conveniently located on the Milfield terrace, a programme of levelling was completed along selected routes to determine the nature of its surface slope more accurately. The following observations are based on an analysis of height information from these two sources.

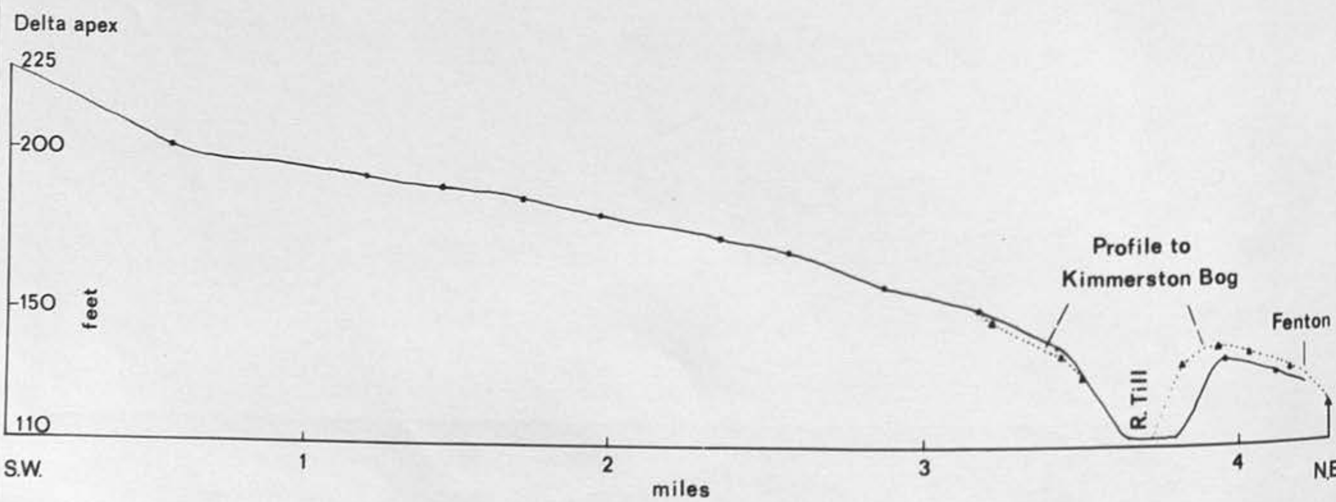
East and south-east of a curving line connecting Fenton Town, Newton, Akeld Steads and Akeld, there are no terrace fragments that can be linked with the Milfield terrace, and it is reasonably clear even from the $2\frac{1}{2}$-inch map that longitudinal slopes on the terrace surface are radially disposed from an apex in the lower reaches of the Glen valley. South of the river, a narrow strip of the terrace terminates at Akeld where it has been severely undercut by the vigorous Akeld Burn. The terrace surface lies at approximately 150 feet and no obvious continuation of it can be traced east of the stream. It seems probable that terrace deposits originally extended a short distance farther to the south-east, for the abrupt termination of the last terrace fragment suggests that they have been either removed or strongly modified by subsequent events. North of the river Glen the terrace surface slopes continuously to its eastern extremity at Ewart. The distal edge between Ewart and Thirlings is not marked by a sharp break of slope, but from a level of between 150 and 144 feet, the surface declines more steeply and descends to approximately 120 feet on the fringe of the Till floodplain. It is unlikely that these relatively gentle slopes at the distal margin of the terrace are simply undercut banks that have become extremely degraded by mass movements and other slope processes, for the undercut walls of channels A and B have retained steep profiles. For this reason it may be suggested that the deposition of sand and gravel extended no farther east than Ewart. The information recorded by Gunn (1895) from a section exposed in the clay pit beside Ewart Brick Works supports this conclusion. The pit is located on the lower part of the slope leading down from the terrace surface, and at that point only 6 to 7 feet of sand were seen to overlie the laminated clays, whereas a well sunk into the surface some distance west of the margin proved at least 24 feet of sand. This evidence indicates that the sand thins rapidly in an easterly direction and suggests that the

Photograph 6.c Fine to medium gravel overlying fine sand in the Milfield delta. Redscar section on right bank of the Till.

Photograph 6.d Gravel wedges in the Milfield delta. Same section as above.


Channel C
187
150
feet
110
S.
1
2
3
N.
miles
(profile on E. side
of r. Till)
Lower Terrace
x
Second
Linthaugh
Cross profile
at point x
r. Till
121
116
feet
587
yards
O


Delta apex
225
200
150
feet
110
Milfield
W.
1
2
3
N.
miles


Delta apex
225
200
150
feet
110
Profile to
Kimmerston Bog
R. Till
Fenton
S.W.
1
2
3
4
N.E.
miles

distal edge of the terrace between Ewart and Thirlings represents the original frontal slope of the feature. A similar conclusion is reached by tracing the slope of the terrace surface in a north-easterly direction from its apex to the vicinity of Fenton (Figure 6.1). Apart from minor riverward slopes immediately adjacent to the Till, caused by subsequent modification, a regular profile is maintained to a height of 140 feet at Fenton Wood. From this place, the terrace surface declines eastwards with a steeper gradient until it dies away at 121 feet, where deposits of fine sand lie only a few feet above the Till floodplain. From Fenton Wood the terrace surface also declines in height towards the north, terminating beside Kimmerston Bog at 120 feet.

The disposition and morphology of the terrace surface thus indicate that the deposition of its constituent sands and gravels was effected by a system of drainage that flowed in shifting stream courses aligned radially out of the Glen valley. That these deposits were built out into a large body of water standing in the Milfield Basin is strongly suggested by the following evidence.

(a) The sands and gravels directly overlie laminated clay and silt.

(b) The existence of a generally even surface on which points roughly equa-distant from the apex are at a similar elevation.

(c) The presence of a distinct slip-off slope at the distal extremity of the terrace.

(d) The disposition of the bedding, for example, in the section at Redscar, where the following is exposed (Photograph 6.c): Grid Ref. 3955/6336.

18 to 36 inches ..... turf, soil and disturbed sand and gravel.

10 to 24 inches ..... unsorted sand and gravel; stones are 1 to 3 inches in size.

28 to 33 inches ..... well-sorted sand, grit and gravel; the beds are steeply dipping like foreset beds; cross-bedding common.

41 inches to an
unknown depth ..... horizontally bedded fine sand; upper layer sharply defined from overlying materials; ripple bedding present in the top few layers.

The total height of this section is from 15 to 20 feet, but the lower part is badly slumped. Three prominent gravel wedges extending from the unsorted gravel layer into the lower layers of fine sand (Photograph 6.d) are additional features of this section; their significance will be discussed in more detail at a later stage.

Thus the presence of delta-bedded gravels on top of finer sediments in the Redscar section, the consistent presence of an upper gravel layer in other sections, and the abundance of gravel litter all over the terrace surface indicate the progressive encroachment of the terrace into the Milfield Basin. Consequently, the lake floor deposits, consisting of laminated clay and silt and bottom-set beds of fine sand, became overlain by the coarser sands and gravels. On the basis of this reasoning it is apparent that the lake was in existence before the narrower part of the Till valley north of Redscar bridge became filled from side to side with sand and gravel. Since the laminated clays apparently extend to a depth of at least 45 to 70 feet, a considerable length of time necessary for the accumulation of these sediments is implied. Accordingly, the lake that once occupied the Milfield Basin must have formed long before the coarser sediments were built out to their ultimate limits.

The lowest height at which lake waters could have drained southwards out of the Till valley is between 300 and 325 feet at Crawley Dean, and since there is no indication that the lake surface ever rose to such a level, it is deduced that its outlet was always towards the north. Lake floor deposits certainly occur beneath the terrace gravel as far north as First Linthaugh

(1L. Map 2) and they are believed to lie near the surface in Kimmerston Bog (Gunn 1895). Indeed, the relatively flat ground extends northwards from Kimmerston Bog and although there are no sections to reveal underlying materials, no stones were observed in ploughed fields and shallow drainage ditches exposed silt. Presumably the lake floor deposits lie close to the surface in this part of the valley also. The section at First Linthaugh is the most northerly in which laminated clays and silts are exposed, yet the terraced sands and gravels extend as far down-valley as Etal. At First Linthaugh, the clays are exposed up to 5 feet above river level, but 200 yards downstream, a somewhat slumped exposure on the same river bank shows that gravel is probably present from river level upwards, to a height of 16 to 18 feet. The exposed material is well rounded gravel, consisting primarily of pebbles from $\frac{1}{4}$ to 3 inches in size, but stones up to $6\frac{1}{2}$ inches in diameter are also present. Coarse sand is partly interbedded and partly forms a matrix for the gravel. At the previous section, the presence of clay causes prominent slump features at the bank-foot and a distinct seepage of water issues from the spring line at the base of the gravels. No such phenomena occur at the latter section, where a steep, gravel-strewn slope descends to river level. It is therefore possible that the clay deposits terminate at some point between the two sections. One problem arising from this gravel section concerns the relatively coarse calibre of the deposit in comparison with the upper gravel layer exposed elsewhere on the terrace north of Milfield. A probable solution to the anomaly will be suggested at a later stage in the chapter.

It may therefore be concluded so far that the sands and gravels of the Milfield terrace were built out into a lake in which laminated clay and silt had been accumulating for a considerable length of time. Since these deposits occur at least as far north as First Linthaugh, the lake must have filled the valley to that point, but may not have extended any farther. The Milfield terrace is therefore considered to be a delta deposit. Because the main surface level terminates chiefly at about 140 feet, the lake surface presumably remained at this level during the deposition of the sand and gravel that compose the foreset and top-set beds. Arising from this situation is the problem of a barrier across the Till valley necessary to impede drainage and cause the formation of a lake. Before discussing the probable nature of this barrier, however, there are other phenomena associated with the Milfield delta that must be considered.

A distinct fall in lake level following the stage when the main delta surface was constructed is indicated by channels A, B and C, which are incised into that surface. The fall appears to have been approximately from 140 feet to 125 feet, the level to which the channel outlets grade. North of Milfield, the surface of the sand and gravel deposits does not decline regularly down valley. Levelling profiles on the surface on either side of the Till indicate that the surface lies mostly between 120 and 127 feet and undulates considerably between these heights (Figure 6.1). Clear riverward slopes are also present. It is difficult to determine whether these irregularities are entirely the erosive work of drainage adjusting to a lower level (ultimately to that followed by the present river Till) or if they are chiefly the result of deposition at the lower level of drainage. Whichever explanation is valid, it is clear from levelling results that there is no continuous down-valley (or up-valley) slope on the surface of the deposits north of Milfield. It has already been suggested that the lake may not have extended beyond First Linthaugh, and since the terraced sands and gravels extend farther north, continuously to Ford Forge and discontinuously to Etal, considerable deposition rather than erosion at the lower level of drainage may well account for the extent of these deposits. This

Photograph 6.e — Massive ice-contact forms in the Cornhill-Etal belt of fluvioglacial deposits.

Photograph 6.f — Eskers in the Cornhill-Etal area.

Photograph 6.g Kame terrace at Branxton and the Pallinsburn kettle hole.

Photograph 6.h Kame terrace south of Pallinsburn.

lower stage in the drainage development was therefore of sufficient duration to enable - (a) prominent channels to become incised into the delta surface; (b) the extension of sand and gravel deposits as far north as Etal. The drop in lake level undoubtedly involves the drainage barrier that initiated the lake and the full implication of this will be discussed later in the chapter.

Channel D indicates that the lake finally drained away rapidly, for the channel floor descends uniformly from its intake at approximately 125 feet to the outlet, grading almost to the level of the present floodplain, just above 100 feet. The main drainage route at this stage must have been that now occupied by the present river, so that channel D eventually became deserted when all the water concentrated into the course now followed by the river Till. The trench in which the river Glen flows was presumably excavated at this time.

2. Landforms Adjacent to the Milfield Delta:

(a) In the Till valley: North of First Linthaugh, the terraced sands and gravels lie adjacent to the eastern extremity of a large system of fluvio-glacial deposits. These are arranged as massive ridges and terraces that extend with unbroken continuity towards Crookham from East Learmouth (Maps 1 and 2). Ice-contact slopes bounding the ridges and terraces are everywhere steep and freshly preserved. The majority of ridges are massive, with broad or narrow crests standing frequently 60 to 120 feet above adjacent depressions (Photograph 6.e). Seldom do they exhibit sharp sinuousities, for they normally run with relatively straight or gently winding courses (Photograph 6.f). Even more impressive than the ridges, however, are the extensive terraces pitted with kettle holes and fringed by crenellate ice-contact slopes (Photographs 6.g and 6.h). On both the northern and southern margins of the sand and gravel belt terrace surfaces lie at between 200 feet and just over 225 feet. The surface of the large terrace occupying a central position in the deposits (west of Pallinsburn) apparently represents a lower and possibly later level of deposition, for it occurs at between 150 feet and slightly over 175 feet. The surface of this terrace is very flat in some places, but gently undulates in others. Throughout the area of sands and gravels it is noticeable that wherever the crest of a ridge rises to approximately 200 feet, the ridge form gives way to an expanse of terrace and in some places, terraces are connected by narrow ridges. Perhaps the most remarkable landforms occuring in association with the sand and gravel deposits are the enormous kettle holes that lie between the ridges and terraces in some places, for example, the one south of Pallinsburn. This feature extends as a narrow, linear depression for 2 miles, bounded by steep, crenellate ice-contact slopes rising over 100 feet above the flat, marshy floor (Photograph 6.g). The original floors of these hollows are obscured by an unknown depth of infill, including peat, which may be quite considerable. The entire belt of deposits is contained within a broad, pre-existing valley lying between a low ridge of Fell Sandstone on the north at Cramondhill and the fringe of the outlying volcanic mass at Branton.

Gunn (1895) provided the first detailed information concerning the nature of this belt of deposits. From sections then available, Gunn fully illustrated the irregular bedding and sorting that he observed to be characteristic. At one point he stated, "From the number of cases mentioned in which the sand and gravel beds are highly inclined it will be inferred that they must have been deposited on very steep slopes, and probably in rather a tumultuous manner, as their irregularity would indicate." The shrewdness of Gunn's observations are endorsed by the ample documentation he gave to the sand and gravel deposits, but rather surprisingly, he offered no explanation to account for their origin and arrangement. Gunn's observations were confirmed by Carruthers (1932), but whereas the former implied that the disposition of the beds was due to original deposition on steep slopes, Carruthers disputed this and concluded that the deposits were a kame and kettle moraine, similar in nature and origin to "dead ice" moraine on Alaskan glaciers described by Gilbert (1904). It seems that Carruthers envisaged a situation in which the margin of the Tweed glacier in this locality became covered with water-laid sand and gravel - the material being composed partly of re-worked superficial moraine and partly of river deposits issuing from the Glen and Bowmont valleys. Furthermore, he apparently considered that these deposits were built up into a tumultuous assemblage of hillocks lacking any obvious arrangement. These conclusions may be criticised chiefly on two accounts.

(1) If the Bowmont/Glen valley was sufficiently free of glacier ice to allow the transport of large quantities of sand and gravel down from the northern flanks of the Cheviots, it is more likely that this material would have been confined to the pre-existing valley of that stream rather than be transported over the watershed and onto the Tweed glacier lying some distance to the north.

(2) Detailed mapping of the sand and gravel deposits illustrates that they are arranged in long, prominent ridges and terraces, clearly aligned either from west to east or from south-west to north-east (Maps 1 and 2).

In a gravel pit at Cleghorne Knowe, south of East Learmouth, Gunn observed beds of "brown sand alternating with fine gravel which have a dip to the N.E. of about 20^{o}"; the ridge in which this bedding was observed trends from south-west to north-east, and so it may be inferred that the stream which deposited these materials flowed in the same direction. The alignments of the ridges, terraces and large dead-ice hollows are generally parallel with the direction of former ice movement as deduced from drumlin orientation and striations. Since the surface slope of a glacier and meltwater drainage normally follow the same direction as ice movement, and in view of the bedding at Cleghorne Knowe, it is concluded that the streams of meltwater which constructed the ridges and terraces of sand and gravel between Learmouth and Crookham flowed with the same alignment as the landforms. Consequently, the ridges and terraces are interpreted as eskers and kame terraces, respectively, and not as lateral moraines.

With regard to the relative age of these fluvioglacial deposits, Gunn (1895) noted how one of the prominent drumlins in this vicinity "rises like an island of clay above the general level of the billowing sea of gravel", and ultimately concluded "though the clay generally occupies the higher ground, compared to the gravel, I have no doubt that the latter is the newer deposit". Following his work on the geological revision of this area, Carruthers (1932) also discussed the drumlin field and came to a similar conclusion. The validity of such a conclusion is further endorsed by the fact that the eskers, kame terraces and kettle holes are all phenomena associated with stagnant, dead-ice, whereas the adjacent and buried drumlins are landforms that require actively moving glacier ice to account for their streamlined form. The high density of kettle holes, the presence of enormous dead-ice hollows and the extent of the kame terraces strongly suggest that during the main period of fluvioglacial deposition, this zone of the Tweed glacier had become not only dynamically dead, but highly fragmented into broken masses of ice. Torrents of meltwater transported vast volumes of sand and gravel into this environment and deposition probably occurred marginal to ice, in tunnels beneath and within ice, in open, ice-walled channels and on top of the ice. The strong lineation of these dead-ice phenomena perhaps results partly from a former crevasse pattern.

Photograph 6.i Upper reaches of the Haydon Dean meltwater channel (Map 2).

The large belt of fluvioglacial deposits extends eastwards as far as the left bank of the Till at Crookham. Only one minor ridge of sand and gravel occurs east of the river, preserved in the core of a prominent meander loop south of Etal. The absence of ice-contact landforms east of the Till is most probably explained in the following manner. A west-facing escarpment of Fell Sandstone rises with moderately steep slopes from the east bank of the Till in this vicinity to about 500 feet (Map 2). This high ground was evidently sufficient to divert easterly-directed meltwater drainage either to the south or to the north. Although three short eskers south of Crookham are aligned towards the south-east, there is little evidence from the majority of adjacent forms indicating a major flow of water southwards. North of Crookham, several long eskers from the west swing towards the north-east. Rivers of meltwater following this alignment would eventually have been guided into the broad, valley-like embayment leading north-eastwards to Duddo, where the Fell Sandstone escarpment swings sharply round towards the north-east. The col at the head of this embayment is breached by a prominent meltwater channel, fed by two shallow but conspicuous intake sections. The most northerly branch begins slightly above 225 feet in Duddo village and curves generally eastwards round the foot of a steep crag of Fell Sandstone. The southern branch begins as a broad, amorphous channel at 200 feet, and runs east and then north to join the northern channel in a broad peat-flat. Emerging from this boggy depression with a more precise form, the channel apparently climbs several feet uphill into a narrow col then winds down the Fell Sandstone dip-slope in an easterly direction. The crest in the floor profile west of the col may be due to the considerable growth of basin peat that occurs in many poorly drained parts of this countryside. From somewhat modest dimensions near its intake (Photograph 6.i), this Haydon Dean meltwater channel ultimately becomes over

75 feet deep and is cut chiefly in bedrock. Extending for a distance of over $4\frac{1}{2}$ miles from the northern intake at Duddo to the outlet at 75 feet O.D. at Ancroft, Haydon Dean was almost certainly excavated by meltwaters flowing from the fluvioglacial drainage system that deposited sands and gravels west of Etal. The inception of this channel was possibly in a subglacial environment, particularly if the bedrock profile beneath the peat is up/down at the col, but proglacial drainage may also have utilised and enlarged Haydon Dean during a later stage in the recession of the Tweed glacier. Since the intake levels of this channel, at approximately 225 and 200 feet, roughly coincide with the crest heights of several eskers and terraces west of Etal, the channel probably controlled the upper level of deposition for some time.

The crests of some eskers and the terrace surface west of Pallinsburn lie at approximately 150 to 175 feet, but follow the same alignment as those adjacent and higher. Furthermore, it seems reasonable to suppose that the easterly flow of meltwater down Haydon Dean had ceased by the time ice wastage had caused the drainage level west of Etal to lie at 150 to 175 feet, and it becomes necessary to determine a possible outlet for this drainage as it emerged from the eastern extremity of the esker/kame terrace belt of deposits. Since there is no route at this level across the Fell Sandstone ridge east of the present Till, the solution to the problem must occur in the vicinity of Etal. Indeed, following its course through a rock-girt valley in the volcanic massif the river Till flows between soft banks cut in glacial drift in the sub-Cheviot depression as far as Etal; but at this place, precisely where the last fragment of the Milfield terraced gravels terminates, the river enters an impressive rock-cut gorge. Only a short distance downstream from the entrance, where rock walls rise 20 feet above the river, the undercut right bank exposes an almost vertical face of Fell Sandstone, rising over 80 feet in height. Downstream from this point, the gorge continues as a deeply incised rock-cut feature, bounded by precipitous walls towering over 100 feet above the river in places, until its outlet into the Tweed gorge at Tweedmill. The Till gorge is not located in the floor of a prominent pre-existing valley, for although the first few hundred yards are aligned along the foot of the Fell Sandstone ridge, it is difficult to detect this gorge in the landscape - especially when viewed from a short distance away. Furthermore, because it is a narrow canyon cut chiefly in bedrock, it does not represent the re-excavation of a drift-filled depression; it simply winds sinuously, following the lowest available route through undulating drumlin topography. Since the present river flows quietly and slowly through the gorge, accomplishing little or no modification to its banks, the magnitude of this impressive canyon seems explicable only by reference to a former period when much larger volumes of water passed down towards the Tweed. The rims of the gorge at its entrance near Etal lie at approximately 150 feet. It is therefore possible that streams of water responsible for sand and gravel deposits at that elevation west of Etal became concentrated at the foot of the Fell Sandstone escarpment and flowed northwards to the Tweed valley, excavating the deep canyon now occupied by the river Till. It is also possible that the meltwater stream responsible for the Till gorge formed part of the drainage system that cut the very large channel south-east of Tweedmouth. Aligned from south-west to north-east, the latter channel is over 2 miles long and significantly begins just below 150 feet.

In the vicinity of Crookham, there is strong evidence suggesting that large blocks of dead-ice lingered in the area until the floor of the gorge had been cut to at least 100 feet O.D. For example, west of Crookham, the bottom of an enormous dead-ice hollow lies between 100 and 125 feet, and its eastern extremity is bounded by steep ice-contact slopes rising to the crests of eskers at Crookham village. This girdle of sand and gravel is breached by a conspicuous channel, however, the sides of which rise precipitously for at least 40 feet above its broad, flat floor. The size of this feature and the smooth, straight form of its walls, suggest that the streamlet presently flowing through an artificial ditch in the channel floor has accomplished relatively little work, even during past periods of higher run-off. A small stream of this nature might be expected to cut a narrow sinuous channel, with interlocking spurs and undercut meander bends, whereas the dimensions and form of the Crookham channel indicate that a much larger stream once flowed from the Pallinsburn dead-ice hollow and into the Till. Such a stream was undoubtedly formed by the melting out of the large mass of dead-ice that formerly occupied the hollow. Since the outlet of the ensuing meltwater channel lies at approximately 100 feet, the Till gorge must have been cut to that level while dead glacier ice still lay in the Crookham area.

The following conclusions may be drawn on the basis of the foregoing discussion.

1. The large system of eskers and kame terraces between East Learmouth and Crookham was deposited by a meltwater drainage system that flowed chiefly towards the east and north-east in a marginal zone of the wasting Tweed glacier.

2. In conjunction with enormous dead-ice hollows and kettle holes, the broad extent of the kame terraces and of some of the eskers suggests that much deposition occurred in open cavities between walls of dead and fragmented ice.

3. An upper level of deposition at 200 to 225 feet (approximately) is characteristic of eskers and kame terraces occupying positions marginal to the belt of deposits. This level was probably controlled by the intake levels of Haydon Dean.

4. A second, common level of deposition at 150 to 175 feet, shown by

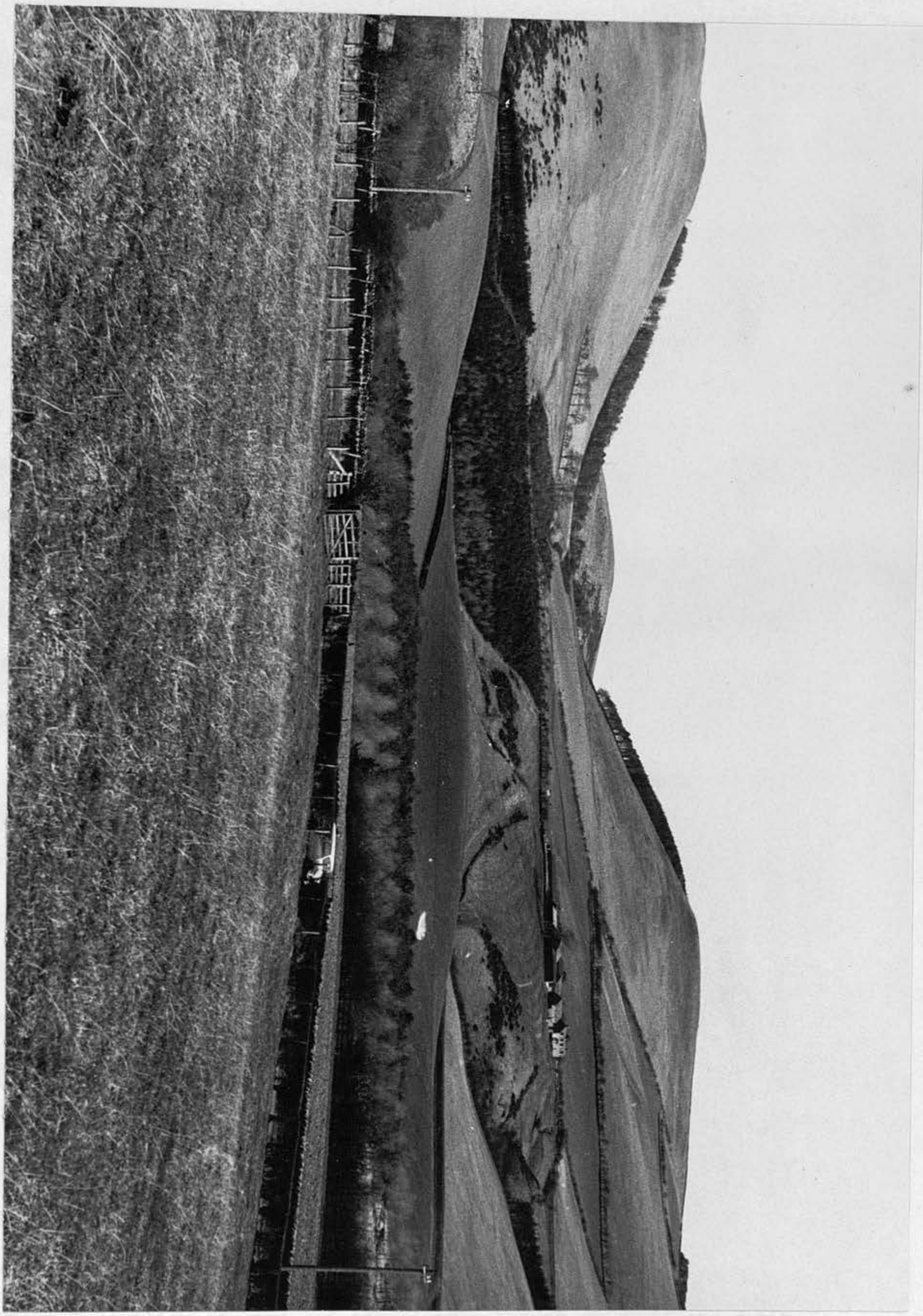
eskers and the kame terrace west of Pallinsburn, seems to have been determined by a drainage route which became established in a northerly direction towards the Tweed.

5. The sinuous rock-cut gorge presently occupied by the river Till was produced chiefly by that drainage, when blocks of dead glacier ice were still melting out amongst the sands and gravels west of Crookham.

The Relationship of the Milfield Delta to the Learmouth-Crookham Fluvioglacial Deposits: Immediately west of Crookham, the esker/kame terrace system comes into close juxtaposition with the lower gravel terrace extending northwards from Milfield. The relative ages of these two depositional stages are reasonably clear. At Old Heatherslaw (O.Hw., Map 2) a broad mound of fluvioglacial sand and gravel rises over 25 feet above the surface of the surrounding terrace lying approximately at 125 feet. A spot height on the crest of the mound is at 156 feet. Although the mound appears to be an outlier of the esker system, its sides are more gently sloping and less sharply defined, except where they have been truncated by channel D. West of channel D the fluvioglacial deposits appear to have been partly degraded to form a group of low amorphous mounds at about 125 feet. These contrast conspicuously with the steep ice-contact slopes of adjacent eskers. The esker-like ridge in the meander core south of Etal rises to 150 feet, but it too is characterised by slopes that are much less sharp and steep than those of eskers to the west. Like the mound at Old Heatherslaw, it is partly surrounded by the terraced gravels. The foregoing evidence suggests that the eastern extremity of the esker/kame terrace belt has not only been partly truncated and washed over by water flowing down the Till valley, but also partly buried and surrounded by terraced sands and gravels. In this instance the latter must be younger in age. However, the main delta surface south of Milfield terminates at between 140 and 150 feet, the height at which the crests of many eskers and the terrace surface west of Pallinsburn lie. It may therefore be possible to correlate these two sets of deposits and suggest that the upper level of deposition was controlled by a drainage outlet northwards to the Tweed valley, through decaying glacier ice. An abrupt fall in drainage level is indicated by channels on the delta surface, and presumably this led to the draining away of the Milfield lake, causing deep incision north of Etal to form the rock-cut gorge. That this fall in drainage level was arrested for a time at a height of approximately 125 feet is suggested by the spread of terraced sands and gravels between Milfield and Etal, but ultimately renewed incision caused channel D to form and led to the establishment of the river Till in its present channel well below the terrace surface.

(b) In the Glen Valley: The Milfield delta deposits can be traced continuously up the Glen valley (Map 2) on the north side of the river to a point 550 yards west of Lanton, where they terminate at 192 feet. A degraded strip of the same delta feature occurs south of the river Glen; extending from Akeld to Old Yeavering, it terminates at a similar elevation immediately opposite. Farther upstream (500 to 1,250 yards) on the north side of the valley, the narrow terrace fragments lying just above 200 feet are probably remnants of the same delta deposit. Approximately 300 yards upstream from West Newton, the small patch of terrace lying at 225 feet represents the farthest up-valley extent of the Milfield delta gravels. Although only a relatively minor feature, it provides an extremely important link in the sequence of landforms in the Glen valley. The proximal edge of the terrace is an ice-contact slope rising from a large dead-ice hollow occupying a considerable area of the valley floor. The crenellate edge of a kame terrace rises 40 feet above the hollow and 25 feet above the small fragment of delta. Rising in an up-valley direction to a height of 275 feet, the surface of the kame terrace merges into the edge of

Photograph 6.j Eskers and kame terrace in the Glen valley at the apex of the Milfield delta.



a massive ridge of fluvioglacial sand and gravel almost filling the Glen valley at Canno Mill. The crest of this ridge, at 325 feet, lies over 100 feet above the adjacent dead-ice hollow and the modern floodplain of the Glen, while only a short distance away, on the other side of the valley, the surface of a prominent kame terrace at Crookhouse also lies at 325 feet and no doubt represents the same phase of fluvioglacial deposition (Photograph 6.j). Beginning in the deep trench between the Crookhouse kame terrace and the massive ridge, a sharp-crested esker extends along the valley floor for 700 yards. Lying much lower than the neighbouring features, the highest point on its undulating crest occurs approximately 25 feet above the river Glen. This esker forms an eastern boundary to the large dead-ice hollow previously mentioned, and its crest lies mostly at a height similar to that reached by the proximal remnant of the Milfield delta. The esker terminates precisely at the point where the delta begins.

It is therefore clear that the upper level of fluvioglacial drainage represented by the Canno ridge and associated kame terraces was followed by a phase of deposition at a much lower level, during which the esker was formed. Since the apex of the Milfield delta has an ice-contact proximal slope, beginning precisely where the esker terminates, it is considered that the delta is a proglacial outwash deposit, formed by meltwater drainage issuing from an ice-contact environment immediately up-valley. Since the kame terrace near Westnewton is partly truncated by the delta apex it is considered that glacier ice had receded a short distance from a former down-valley position before the main phase of outwash deposition began. Subsequently, however, glacier ice must have remained stationary in the Crookhouse area sufficiently long to allow the extensive deposition of sand and gravel that formed the Milfield delta.

3. High-Level Terrace Fragments in the Milfield Basin: Narrow terrace fragments at a higher level than the Milfield delta occur at several places in the Milfield Basin, and appear to be remnants of deposits that were formerly more extensive. One such terrace occurs at Old Yeavering in the lower Glen valley. Rising like an island from the adjacent delta level, 30 to 35 feet lower, the terrace is roughly 450 yards long and 250 yards broad. An old sand pit at the western end is 20 to 25 feet deep. Although the sides are badly slumped, the upper 4 to 6 feet were clearly observed in 1962 and showed the following: below the soil mantle (6 to 14 inches) is approximately 3 feet of horizontally bedded gravels, fine to medium grained and sub-rounded in shape. These gravels are underlain by steeply dipping beds of sand (with occasional seams of fine gravel), arranged similar to the foreset beds characteristic of deltas. The beds dip towards the centre of the valley. Other terrace fragments that possibly belong to the same level of deposition occur farther down-valley, 800 to 200 yards west of Akeld. They are much smaller and more doubtful features, however, lacking exposures, but since their surfaces lie at approximately 200 feet, they may be considered as possible remnants of the upper terrace level. The most extensive strip of terrace at 200 feet occurs in the vicinity of Lanton. Curving northwards for over a mile round the lower slopes of Lanton Hill, the terrace edge rises 20 to 25 feet up from the Milfield delta to a surface that has a maximum width of nearly 200 yards. Poor exposures on this feature reveal sand and gravel; these materials presumably underlie the entire terrace. A farm on the terrace is significantly called Sandy House, while the Geological Survey Drift Map shows fluvioglacial sand and gravel coincident with the terrace.

These high-level terraces are clearly earlier in age than the Milfield delta, for they appear to have been fragmented and degraded before the delta deposits were banked up against them. Since their relative position is similar to that of the kame terraces at Crookhouse and near Westnewton, they are possibly related to the same phase and level of fluvioglacial deposition. In this respect, two alternative explanations are possible for the terrace fragments:

(a) They are remnants of kame terraces, the original extent and slopes of which have been modified by subsequent erosion; but they may never have been considerably more extensive.

(b) They are remnants of a proglacial outwash deposit that formerly filled the lower Glen valley and possibly occupied much of the Milfield Basin.

The valley-ward dip of the bedding at Old Yeavering suggests a kame terrace origin, the deposition occurring in a narrow marginal ice lake, but distinct ice-contact slopes are lacking; the latter problem is possibly explained by subsequent modification, however. In support of the second alternative is the occurrence of laminated lake clays up to a height of 200 feet in the Chatton area and in the Wooler Water valley, and since the terrace surface at Sandyhouse lies at 200 feet, the possibility that its level was controlled by the former presence of a lake at that height is perhaps quite valid. The 200-foot level of deposition is approximately coincident with that shown by eskers and kame terraces west of Crookham. It has already been indicated that 40 to 70 feet of stoneless clay, which is probably laminated lake clay, occur in the Milfield Basin, and since a considerable length of time is necessary to allow such enormous thicknesses of fine sediment to accumulate, the existence of a lake at a stage earlier than that in which the Milfield delta was subsequently formed is perhaps a reasonable explanation. However, the absence of conclusive evidence prevents any confident interpretation of the high level terrace fragments in the lower Glen valley and on the western side of the Milfield Basin.

In the vicinity of Nesbit, where a broad embayment occurs in the Fell Sandstone escarpment on the eastern side of the Milfield Basin, the surface of a broad terrace lies at approximately 160 feet. A small number of poor exposures reveal sand and sub-rounded gravel, and since these deposits are far removed from any present stream, they may be considered fluvioglacial in origin. The slopes descending from the terrace surface cannot be compared with those characteristic of ice-contact landforms composed of similar deposits. In this respect, the Nesbit terrace slopes seem to have been considerably modified, but the large embayment in the terrace is possibly a dead-ice hollow, thereby implying that the feature is an ice-contact landform.

The somewhat amorphous ridge rising 20 to 25 feet above the Milfield delta level near Kimmerston is also composed of fluvioglacial sand and gravel; several cobbles on the surface are 8 to 9 inches in diameter. The feature extends as a broad ridge, aligned north-west to south-east, and is bounded by gentle slopes, much less distinct than those of eskers in the Crookham area nearby. The ridge crest lies at approximately 160 feet.

The absence of sections hinders a satisfactory explanation of the two landforms described above, but they clearly consist of fluvioglacial sand and gravel. The marked vagueness of their slopes suggests that they have been considerably modified subsequent to deposition. In view of the level at which their surfaces lie, it is suggested that both landforms were washed over and modified during stages of the Milfield lake and that they possibly represent fragments of fluvioglacial deposition older in age than the period of lake development in the Milfield Basin.

4. Evidence from Areas Adjacent to the Milfield Basin: Since areas south of the Milfield Basin were most probably ice-free during the stage (or stages) of lake development, and since substantial parts of the Chatton basin and the

Wooler Water valley lie below 200 feet, a careful investigation was made in these areas to see if the streams had formed terraces or deltas that could be linked with the former presence of narrow extensions of the Milfield lake.

(a) The Till Valley: A reconstruction of the former lake to a level of 200 feet shows that it would have extended up the Till valley as a narrow neck of water to terminate in the vicinity of New Bewick. From New Bewick to the Weetwood gorge (beyond which the Till flows across the Milfield Basin) the river and its modern floodplain are confined within the relatively narrow trench cut in glacial drift. Immediately south of New Bewick, however, the river flows across broad haughlands, up to 900 yards wide, and these extend up-valley for approximately 1¼ miles. It is difficult to determine whether or not the considerable expanse of sand and gravel composing these haughlands represents a former period of active aggradation controlled by a lake level at 200 feet. If the haughlands were formed during such a phase of aggradation rather than solely by the lateral erosion of adjacent glacial deposits, then distinct terraces ought to have been developed as the river readjusted itself to a lower base level when the lake ultimately drained away. Terraces do not occur in the area south of New Bewick. Indeed, terraces are entirely absent from the Till valley between New Bewick and Weetwood, so that there is little evidence upon which to base the former presence of a lake at 200 feet, except for the occurrence of laminated clays.

(b) The Hetton Valley: A substantial part of this broad valley (Map 3) lies below 200 feet, and would undoubtedly have been occupied by a wide arm of the Milfield lake had its surface ever lain at that height. The Hetton Burn is a small stream that presently flows in a narrow channel cut through superficial deposits. In the lower section of the valley the stream is incised by 10 to 15 feet below the surface of a distinct terrace that slopes from approximately

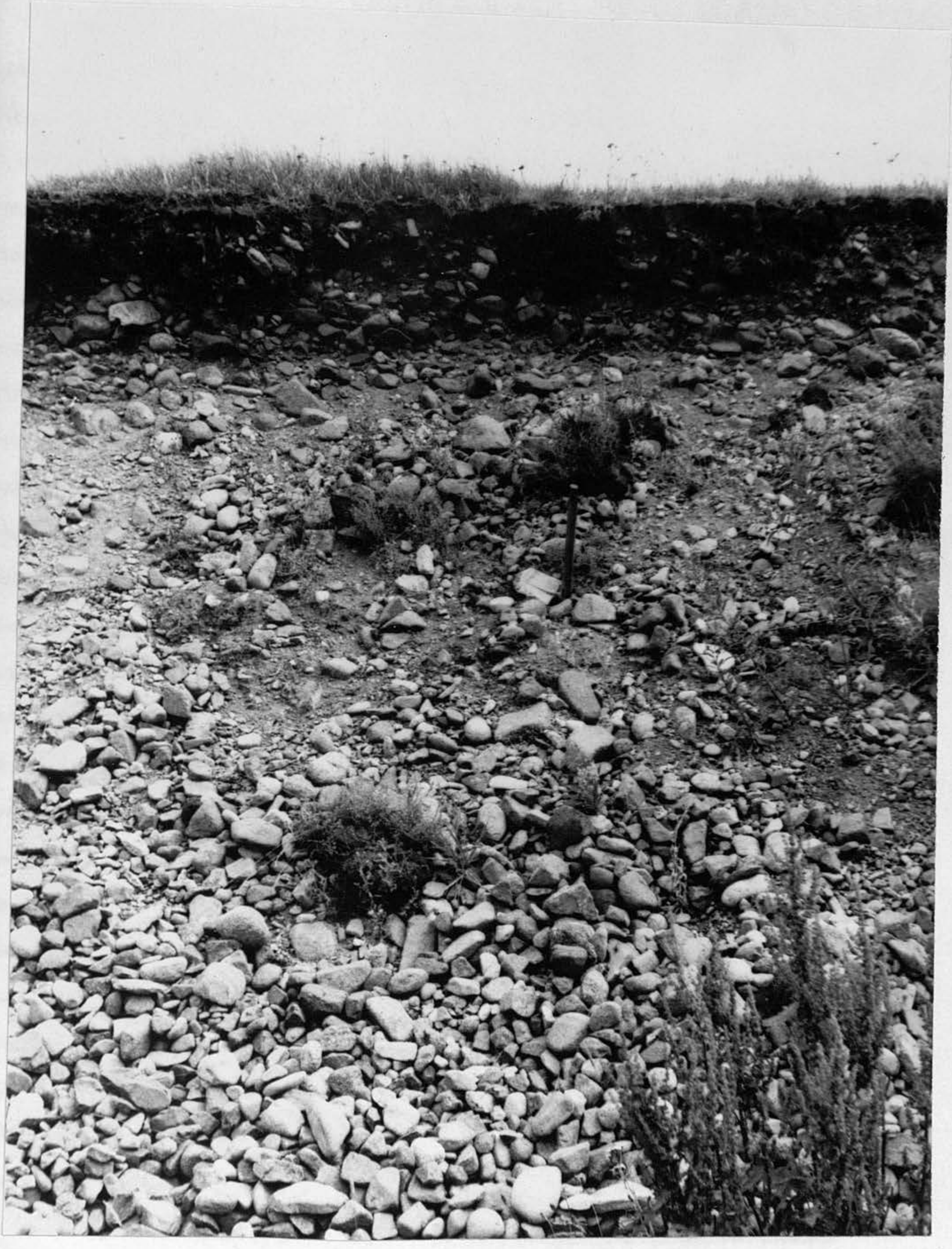
175 feet to 150 feet. Exposures are poor and occur in only three places, two of which reveal sand and gravel near the surface. The other shows that a tenacious, sandy clay, packed with stones, lies 3 to 4 feet below the terrace surface; this material is probably till. On the basis of this somewhat inadequate evidence it may be tentatively suggested that the sand and gravel is a relatively thin deposit overlying the till. Since the terrace surface slopes down-valley from approximately 175 feet to 150 feet, it is perhaps more valid to correlate it with the Milfield delta than with a possible lake level at 200 feet. The absence of delta deposits grading to 200 feet in the Hetton valley does not necessarily disprove that the valley was ever occupied by lake waters to that level. Indeed, the presence of a lake at 200 feet would have reduced the stream's catchment area to such an extent that only small quantities of debris could have entered the lake. This reasoning seems especially valid when it is considered how little water is supplied to the present stream by the entire valley. The absence of a delta at 200 feet is therefore to be expected, while the 3 to 4-foot layer of sand and gravel composing the terrace between 175 and 150 feet is in keeping with the size of the valley.

(c) The Wooler Water Valley: Before entering the Milfield Basin, the Wooler Water flows through a broad vale developed on Cementstone bedrock lying between the Fell Sandstone escarpment of Weetwood Moor and foothills of the volcanic massif. Within this pre-existing vale the river has truncated several eskers that form part of the large system south of Wooler, and it is clear that fluvio-glacial deposits were formerly much more extensive in this area; whether or not they ever filled the vale from side to side is difficult to determine. The lower section of the valley lies below 200 feet and since laminated clays are exposed along the river bank, it has been suggested that they possibly accumulated in an arm of the Milfield lake. Apparently supporting this sug-

Photograph 6.k The Wooler Water gravel plain.

Photograph 6.l Section in the Wooler Water gravel plain.

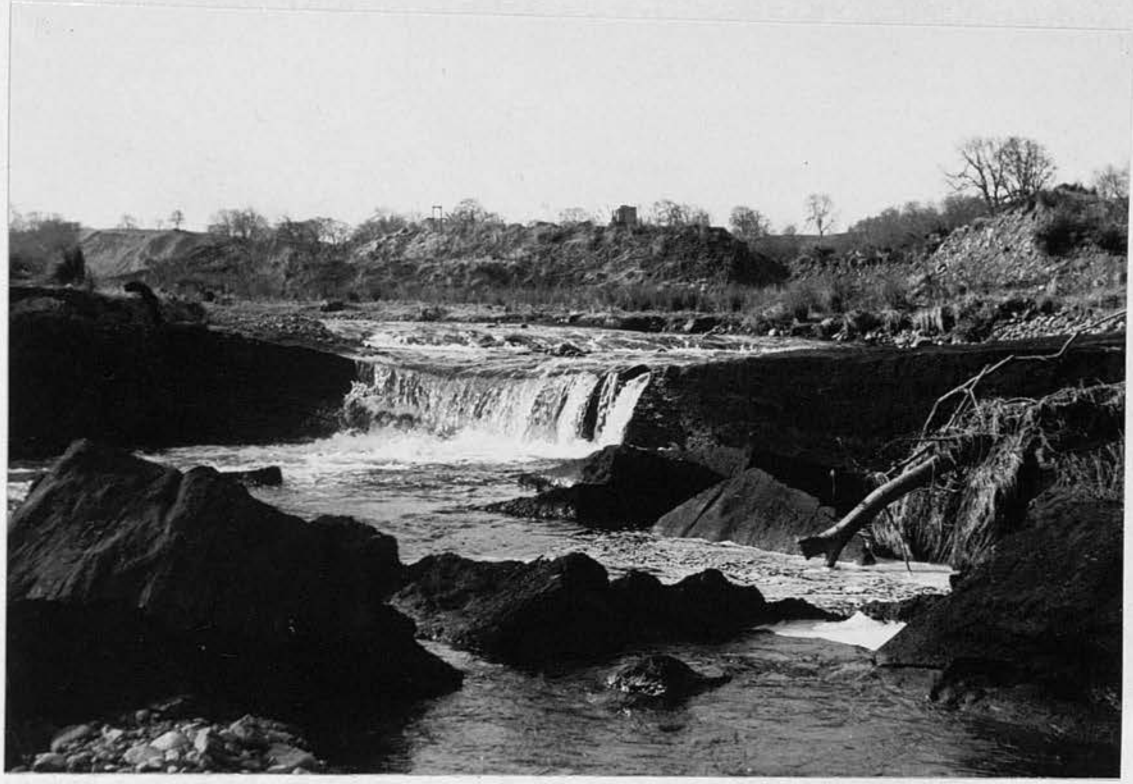
Photograph 6.m Coarse gravel and cobbles composing the Wooler Water gravel plain.



gestion is the prominent gravel plain filling the valley between the truncated esker system and the foot of the escarpment (Photograph 6.k). The river is presently located at the western side of the gravel plain and actively undercuts the esker system in several places. At Coldgate Haugh the river flows over a bed of coarse gravel and cobbles no more than 3 to 4 feet below the level of the gravel plain, but from this point onwards the river becomes progressively incised below that level until it is over 15 feet lower (Photograph 6.1). Clear sections are constantly present along the river banks, and show that the surface of the gravel plain is consistently underlain by a layer of poorly bedded gravels and cobbles. The stones vary in size from one-inch pebbles to boulders over 3 feet in diameter and are contained within a gritty matrix. The majority are water-worn to some degree, but many vary in shape from well-rounded to sub-angular (Photograph 6.m). These deposits are identical in size, shape and lithology to those exposed in the truncated ends of adjacent eskers and were clearly derived from the erosion of these landforms. The consistent presence of this gravel layer over a distance of at least $1\frac{1}{4}$ miles is quite remarkable. It is mostly 6 to 8 feet thick, although 11 feet 6 inches occur at Earle Mill. The long axes of the stones generally dip upstream, indicating that deposition was by water flowing down-valley (in relation to the present stream). Beginning approximately at 300 feet, the gravel plain extends continuously towards the river Till near Weetwood, and terminates there at 145 feet. This range in altitude is not entirely consistent with deposition in a lake in which floor deposits accumulated to above 200 feet, but since the surface of the Milfield delta terminates at between 140 and 150 feet, it could be considered that the Wooler Water gravel plain was partly related to the lower lake level indicated by that delta. However, considerable doubt concerning the glacial age of the gravel plain arises from the following phenomenon recently exposed

Photograph 6.n The peat bed exposed by the Wooler Water (August 1963).

Photograph 6.o The peat bed overlain by the gravels of the Wooler Water gravel plain.



Photograph 6.p	Laminated clay overlain by coarse gravel, right bank of the Wooler Water.
Photograph 6.q	Fine sand overlain by coarse gravel, right bank of Wooler Water.

POLLEN DIAGRAM WOOLER

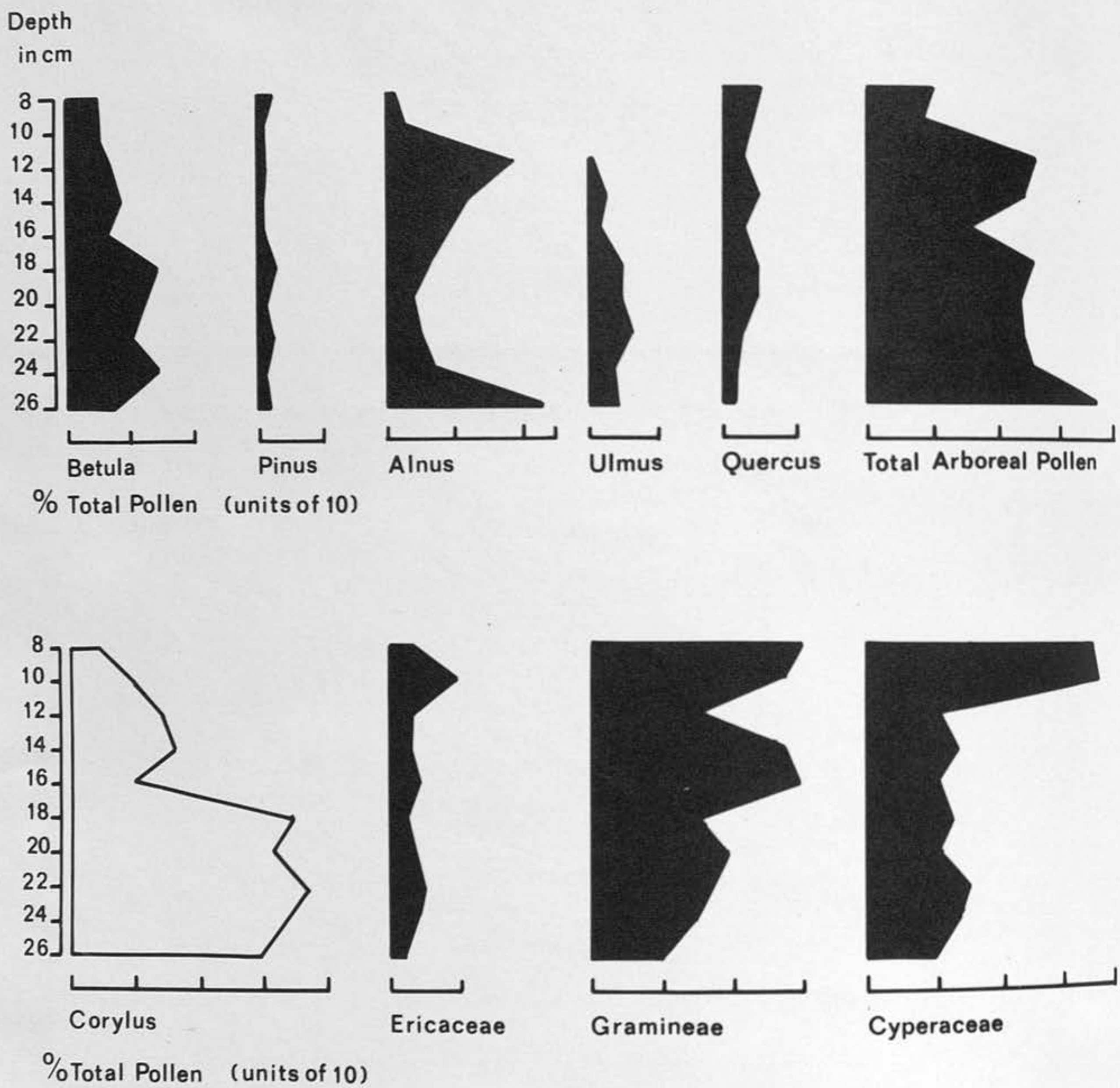


Figure 6.2

(1963) by the Wooler Water. Between points i and ii, near Earle Mill (Map 5) a thick bed of peat has been continually exposed and eroded by the river since 1963. As far as can be determined from the river-cut sections the peat bed thins from a maximum depth of over 7 feet near Earle Mill both upstream and downstream; the bed also becomes thinner on either side of the river. The peat bed thus appears to be a basin formation. The gravel layer referred to above rests directly on top of the peat and must be younger in age (Photographs 6.n and 6.o). An analysis of the pollen content in the peat, undertaken by S. E. Durno, demonstrates that it grew during the Atlantic period (Figure 6.2). Consequently, the gravel plain, which is younger than the peat, is in no way related to the former presence of an ice-dammed lake in the Milfield Basin and must represent aggradation accomplished by the Wooler Water since Atlantic times.

Upstream from the peat bed, the laminated clays which were interpreted as lake floor deposits lie directly beneath the gravel layer at the base of the river bank (Photograph 6.p). A short distance farther upstream, however, the clays are no longer present and the gravel layer rests on beds of sand and silt (Photograph 6.q). The latter are of considerable interest. They are up to 7 feet thick and are remarkably similar in colour to the laminated clays, which are pink, red-brown and grey-green. The sands and silts are distinctly cross-bedded, dipping in a down-valley direction. Since they are located between 200 and 225 feet, they possibly represent a deltaic deposit built out by the Wooler Water into the lake in which laminated clays accumulated to a level of 200 feet. Subsequently, these lake deposits became buried by the gravel plain.

The Atlantic peat bed appears to rest partly on coarse sand and gravel deposits and partly on till, so that its precise stratigraphic relationship with the lake sediments is difficult to establish. The distinct basin form of the peat bed, however, indicates that growth took place in a pre-existing hollow over 7 feet deep, and since the peat has been heavily compressed by the overlying gravels, its original thickness was probably considerably in excess of 7 feet. The presence of a hollow in which peat growth could have become established and the absence of lake sediments beneath the peat can be accounted for in the following manner. The adjacent esker system contains many large kettle holes, some of which contain peat. Furthermore, south of Haugh Head the river has undercut the fluvioglacial deposits to such an extent that only a relatively narrow ridge remains between the river and the large kettle hole adjacent. The river removes considerable quantities of sand and gravel each year when in flood and it seems inevitable that it will ultimately encroach upon the kettle hole. When this happens, peat and other deposits in the kettle hole will become covered with a layer of coarse gravel deposits derived from adjacent fluvioglacial landforms and redistributed by the river. It seems unrealistic to assume that a sequence of events such as this could not have taken place in the past. For these reasons it is suggested that the peat bed near Earle Mill possibly accumulated in a kettle hole that was inundated by the Wooler Water after Atlantic times and covered by up to 11 feet of coarse gravel deposits.

The following conclusions may be made on the basis of evidence from the Till valley and the valleys of the Hetton Burn and Wooler Water.

(a) Apart from deposits of laminated clays, there are no other deposits or landforms that can be confidently associated with a former lake level at 200 feet. Sand and gravel haughlands in the Till valley south of New Bewick were possibly formed when a lake surface lay at 200 feet and the pink, red-brown and grey-green sands and silts exposed in the Wooler Water valley may represent a deltaic deposit at approximately 200 feet.

(b) A sand and gravel terrace sloping from 175 feet to 150 feet in the Hetton valley was probably deposited contemporaneously with the Milfield delta in a shallow arm of the lake. Terrace fragments on the south bank of the Till west of the Weetwood gorge lie at 145 feet and could be linked with the same lake, but powerful lateral erosion and redistribution of fluvioglacial sands and gravels accomplished by the Wooler Water since Atlantic times (and probably before that period) seem to have obscured any evidence of aggradation related to a 150-foot lake surface in the Wooler Water valley.

Conclusion

The occurrence of laminated clay and silt deposits in the Milfield Basin and the lower reaches of adjacent valleys to the west, south and east indicate that these areas were at one time submerged beneath an extensive lake. Within the Milfield Basin the lake floor sediments have not been recorded above a height of 150 feet, but they occur up to 200 feet in the Chatton area of the Till valley and south of Earle Mill in the Wooler Water valley. Boreholes have shown the laminated clay to be at least 40 feet thick and as much as 70 feet possibly occurs. To enable this thickness of such fine sediment to accumulate the lake must have been present for a very long period of time. The laminated clays terminate precisely where a belt of large eskers and kame terraces is aligned across the valley leading northwards from the Milfield Basin, the lowest route available for escaping drainage. Since these ice-contact landforms accumulated in a marginal zone of the Tweed glacier, it is suggested that glacier ice formed the effective barrier that impounded drainage from the south in the Milfield Basin, forming an extensive lake. The floor of a large dead-ice hollow at Crookham lies below 125 feet and has not been modified by lake waters or lake sediments; it was therefore occupied by glacier ice when the lake was in existence. A broad terrace of sand and gravel slopes radially out from the Glen valley. Declining in height from an apex at 225 feet to a frontal slip-off slope at between 144 and 150 feet its surface lies up to 35 feet above the modern floodplains of the Till and the Glen. The calibre of this deposit becomes progressively finer towards the distal extremity and the fine sands located at a level of 120 feet near Fenton Town are interpreted as bottom-set deposits. The entire formation represents a delta built out into the Milfield lake when its surface lay approximately at 150 feet. A minor deltaic formation appears to have been constructed contemporaneously in the Hetton Burn valley, and the small terrace remnants bordering the Till west of Weetwood possibly formed at the same time. Since a prominent ice-contact slope bounds the proximal edge of the delta apex, upstream from which occur eskers, kame terraces and a dead-ice hollow, it is considered that the delta sands and gravels are proglacial outwash deposits derived from an ice-contact environment west of Kirknewton in the Glen valley. Conspicuous dry channels furrow the Milfield delta in some places and indicate a pronounced fall in lake level to a height of 125 feet. North of Milfield an irregular terrace surface suggests that the gravel deposits were extended farther north at that period, resulting in a modification and partial smothering of eskers lying at the eastern extremity of the great systems at Crookham. This event indicates a marked recession of the glacier margin. The terraced gravels terminate at Etal (125 feet) precisely where the entrance to a spectacular rock-walled gorge occurs. Since the rims of the gorge were cut at a level of 150 feet, it seems that this feature owed its inception to drainage flowing north from the Milfield lake when its surface stood at that level; this drainage presumably flowed towards the Tweed in a subglacial course that took a line of least resistance, winding sinuously through rolling drumlin topography. An impressive dry channel incised into the terraced gravels south-east of Crookham grades almost onto the present floodplain of the Till just above 100 feet. It was probably deserted as drainage escaping from the Milfield Basin finally became entrenched and established in the course now occupied by the river Till. The absence of terraces within this river trench suggests that the ultimate adjustment of drainage was accomplished relatively quickly (presumably when the ice barrier wasted away). The high-level terraces that slope out of the Glen valley to a height of approximately 200 feet, are possibly linked with a lake level at that height; laminated clays in the Chatton Basin and in the Wooler Water valley may belong to the same period. There is insufficient evidence, however, upon which to base any confident conclusion concerning events linked with a possible lake level at 200 feet.

CHAPTER 7.

THE CHEVIOT ICE CAP

Introduction

Previous literature dealing with glacial events in Northumberland refers mostly to ground below 1,000 feet where phenomena produced by glaciation are generally abundant. A small number of meltwater channels and some till deposits occur above 1,000 feet, but the evidence of glaciation is much less apparent in the higher parts of the hills and so they have not received a great deal of attention. Consequently, it has never been fully debated whether or not the Cheviot massif and the higher summits of the Cheviot Hill range to the west could ever have been independent centres of glacier dispersion.

The detailed field work completed as a basis for this thesis was confined to the eastern part of the Cheviot massif, but since the concept of a Cheviot ice cap is intimately linked with high ground farther to the west, it is proposed to include relevant evidence from brief field investigations and from previous literature on the latter area.

Earlier workers in the Cheviots such as J. Geikie (1876), Clough (1887, 1888), Smythe (1912) and Raistrick (1931) did not doubt that the higher parts of the Cheviots had been formerly covered by glacier ice. For example, when discussing the mass of ice that occupied the Tweed valley, Geikie (1876) stated, "The Cheviots appear to have been quite buried underneath this wide sea of ice". The map produced by Geikie (1894) to show the "British Isles During The Epoch of Maximum Glaciation" indicates a flow of ice south-eastwards from the summits of the Cheviot Hills and he evidently believed that valley glaciers had existed in the Cheviots at some stage, for he stated (1894), "I have noticed true moraines also at the head of certain valleys in the Cheviot Hills". Clough (1897) was clearly aware that the lower ground south of the

Cheviot Hills had been covered by ice from a western source, for he spoke of "a general ice-sheet, wrapping up the whole district, and covering all the lower elevations of the Cheviot Hills", but he also considered that "the top-most Cheviots and some of the head valleys were certainly occupied by one or more local ice caps. Thus, the hills on the south side of the Rede ..... were probably on the borders of a cap covering Carter Fell." In the Cheviot Hills memoir, published the following year, Clough further emphasised his theory of a local ice cap when he stated "The higher summits, Cheviot, Hedgehope, Comb Fell, Cushat Law, &c., seem never to have been overridden by foreign ice, but to have acted as independent centres of glaciation." Subsequent publications by Smythe (1912) and Raistrick (1931) provided general accounts of ice movements in Northumberland and evidently accepted the former presence of local ice in the Cheviots. For example, Raistrick (1931) observed that "In Northumberland, local glaciers had flowed almost radially from the Cheviots to the east; northward, then east from the northern flanks, mingling with the Tweed glacier; east and south down the Coquet and the Rede, and into the North Tyne valley; and also south-east into the North Tyne from the Carter Fell-Peel Fell hills." Smythe and Raistrick appear to have based their conclusions on work previously done, by Clough, however, and they present no original observations in support. A re-examination of the "Central Cheviots" by Carruthers et al. (1932) led to the following conclusion; "Indeed, the case for an early 'Cheviot glaciation' is as yet by no means proved. Much can be said against such a conception. Not only is the gathering ground for an effective ice-cap very small, but, as has often been noticed, there is a remarkable scarcity of Cheviot granite boulders in the surrounding lowlands. ..... one would have expected them to be plentiful enough, had the Cheviots ever been an active centre of dispersion." In this way Carruthers et al. expressed their doubts that the Cheviots at one time contained local glacier ice, but they ultimately stated, "And finally the whole character of the local drift around the Cheviot reminds one of snow-scree material, or at any rate of the debris of small valley glaciers, perhaps hardly more than half-consolidated neve." The question of a Cheviot ice cap was again raised by Common (1953). He concluded, "Field observations tend to support the view that Cheviot did NOT, indeed could NOT, supply the amount of neve to form a local ice centre." Since Common did not mention what his field observations were, however, his statement must be considered inconclusive. Nevertheless, in a recent appreciation of the glacial period in Britain, Sissons (1964) accepted the latter view and considered it likely that "the Cheviots were never completely ice covered". A similar suggestion was made by the same author in 1965.

In view of the conflicting opinions concerning the validity of a Cheviot ice cap it seems appropriate within the context of this thesis to examine the available evidence in more detail. Since glacial erratics have been recorded up to heights of 1,900 feet (Clough 1888) in the Cheviot massif, it seems certain that glacier ice was present to at least that elevation. What now remains to be established is whether (a) the ice was derived locally from the higher summits, or (b) the ice came from external sources and built up around the massif to a level of 1,900 feet, leaving the higher ground ice-free, or (c) the foreign ice covered the Cheviot massif entirely.

Evidence from Meltwater Channels

Extensive systems of glacial meltwater channels occurring on the north-east and south-east flanks of the Cheviot massif (Chapters 2 and 4) suggest that the meltwater drainage was associated with two separate masses of glacier ice, one of which moved south-eastwards round the massif from the Tweed valley and the other north-eastwards round the massif from the west. The channel systems converge in the Breamish valley. On the hillsides in both areas there is a distinct upper limit to the concentration of meltwater channels, below which such phenomena are very numerous. Furthermore, the approximate upper limit of concentration declines in height towards the Breamish valley from the western extremities of both areas. Another characteristic of the channel systems is their location peripheral to the massif as a whole. The central parts of the east Cheviot massif are remarkably devoid of meltwater channels that can be linked with either the north-east or south-east system. Meltwater channels are not entirely absent from the inner and higher parts of the massif, however. For example, two large channels breach the eastern watershed of the Lambden Burn (Chapter 2). Since they were both cut by streams of meltwater flowing north-eastwards and eastwards out of the upper Lambden valley it was concluded that glacier ice was thicker in that valley than in areas to the north-east and east beyond the cols in which the channels are cut. Because the channels probably formed subglacially following the superimposition of englacial streams it is possible that glacier ice had previously flowed towards the north-east and east through cols. The glacier ice that formerly occupied the Lambden valley can therefore have come from one of two possible sources: (a) it could have moved into the valley from the west as part of the Southern Uplands ice sheet which flowed generally eastwards between the Lammermuir Hills to the north and the Cheviot Hills to the south; (b) it could have been part of a local ice cap that built up on The Cheviot. An easterly component to its movement in the Lambden valley-head may have been induced by the pressure of the much larger ice mass flowing eastwards down the Tweed valley.

Another part of the east Cheviot massif in which the alignments of meltwater channels do not conform to those of the north-east and south-east channel systems is the mid-Breamish valley. On both sides of the valley the meltwater channels (Maps 7 and 8) indicate that meltwater flowed either down the valley or radially out of it, suggesting that the occupant glacier ice had come from high ground at the valley-head. The frontal margin of ice in the Breamish valley did not progress beyond Ingram, since all features of fluvio-glacial erosion and deposition east of that locality are related to either northern or southern ice. With regard to the east Cheviot massif in general, the approximate minimal upper limits of encroachment by northern and southern ice, as indicated by meltwater channels, have already been established in Chapters 2 and 4; those relating to ice in the Breamish valley may similarly be discussed. The highest evidence of meltwater erosion on hillsides south of the river Breamish is at 1,100 feet near North Pike, on the High Knowes-Het Hill spur. Eastwards, the highest channel on each successive spur is at a lower elevation, until the Middledean channel at 900 feet terminates the sequence (Maps 7 and 8). The transfluent lobe that pushed south-eastwards between Hogden Law and High Knowes built up in height to over 1,100 feet, which is the level of the col floor. North of the Breamish meltwater channels are less numerous and approximate limits are even more tentative. With reference to Maps 7 and 8, channel 36 begins at 1,400 feet and is the highest member of this series; channel 40, the most easterly, lies at 900 feet, similar to the Middledean channel, the most easterly of the series south of the river. This limited evidence suggests that the surface of glacier ice in the Breamish valley built up in height to at least 1,400 feet in the vicinity of channel 36 and descended generally eastwards to the Ingram area, where channels intaking at 900 feet probably mark its farthest extension in that direction.

The foregoing accounts of meltwater channels occurring in central parts of the east Cheviot massif indicate the following possibilities.

(a) The Lambden valley and possibly the upper Harthope valley were occupied by glacier ice characterised by a surface slope towards the north-east and east. The ice surface was higher than about 1,450 feet.

(b) The mid-Breamish valley contained glacier ice with a surface which sloped radially out of the valley in an easterly direction. The minimal upper limit of the ice surface was 1,400 feet.

These conclusions by themselves cannot perhaps be taken as incontrovertible evidence that the glacier ice that formerly occupied the Lambden and mid-Breamish valleys originated quite separately from the glacier masses which flowed round the north-east and south-east peripheries of the massif. Yet the alignments of meltwater channels associated with ice in the central part of the massif distinctly contrast with the orientation of channels in the north-east and south-east systems, and strongly suggest that the surface slope of the central ice was radial to the core of the massif on its eastern side. It should again be emphasised that the north-east and south-east channel systems are clearly defined by upper limits, each of which declines in height towards the Breamish valley, probably reflecting the approximate surface slope of the two ice masses. Since it is believed that the fluvioglacial activity responsible for the channel systems occurred chiefly in the marginal zones of the two ice masses, the occurrence of other meltwater channels at much higher levels towards the central parts of the massif is difficult to reconcile with the same two ice masses. Indeed, the upper limits of meltwater channels on the north-east and south-east flanks of the massif are so striking that it seems unrealistic not to consider them meaningful in establishing approximate limits of incursion of extraneous glacier ice into the fringes of the massif. It is difficult to understand why the relatively broad expanses of territory extending farther in towards the heart of the massif should be so devoid of meltwater channels similar in form and alignment to those of the north-east and south-east systems if the invading ice masses had penetrated beyond the limits suggested by the channel systems. A possible explanation is that the Lambden, Harthope and mid-Breamish channels were formed during a previous and more extensive glacial stage, but there are factors which render this suggestion improbable. For example:

(a) Channels in the Lambden, Harthope and mid-Breamish valley are equally fresh in appearance and similar in form to those of the north-east and south-east. Had the former originated during a previous period of deglaciation they might reasonably be expected to have suffered severe modification by periglacial processes operating during the following glacial period, and in this case would contain much greater amounts of debris.

(b) If the northern and southern ice masses penetrated into the east Cheviot area only to the approximate limits suggested by the two large channel systems, considerable areas of the central part of the massif would have remained ice-free (assuming that there was no local ice cap). Since considerable valley systems occur above 1,250 feet, for example, those associated with the Lambden, Harthope and Breamish, the former occurrence of glacier dammed lakes in several valleys might reasonably be expected had these areas remained ice-free. Such lakes would almost certainly have endured for substantial periods of time because they would owe their existence to glacier advance rather than recession. Furthermore, the presence of deeply rotted bedrock and the prevalence of periglacial conditions in the ice-free areas would have ensured that abundant debris was available for seasonal meltwater drainage from snow, ice and frozen ground. Under such conditions large deltas ought to have been constructed, particularly in the major valleys, had ice-dammed lakes ever formed. Not only is there a complete absence of deltas, but also there is no other evidence, such as floor deposits, shorelines and overflow channels to even suggest the former presence of ice-dammed lakes in the Cheviot massif. While the absence of such phenomena does not necessarily prove that ice-dammed lakes never existed in the upper valleys, it certainly throws considerable doubt on the suggestion that these areas remained ice-free while glacier masses encroached upon the north-east and south-east flanks of the Cheviot massif.

It is therefore possible that meltwater channels in the central part of the Cheviot massif were formed during the same period of deglaciation as those on the north-east and south-east flanks. It is also possible that glacier ice was present over central parts of the massif when the peripheral areas were inundated by the northern and southern ice masses. If the upper limit of the meltwater channel systems produced by the latter provide a valid limit of glacier incursion, then the central parts of the massif must have been simultaneously occupied by glacier ice that was nourished locally. The local ice cap was presumably confluent with the extraneous ice and may have been instrumental in preventing further encroachment of the invading glacier masses. The alignments of meltwater channels associated with the local ice suggest that it flowed out radially from the high core of the massif, chiefly in easterly directions. The approximate coincidence in the slope of the ice surface with the pre-existing valleys probably explains, in part, the relative infrequency of meltwater channels over wide areas of the inner massif, for fluvioglacial drainage would probably have utilised the pre-existing valleys.

Detailed field work was not undertaken in the western part of the Cheviot massif, but the entire area was studied stereoscopically on vertical aerial photographs and some field checking was done. One of the most striking facts about that area is the small number of meltwater channels present, even although the topography is similar to that in the north-east and south-east

Cheviots and the valleys are aligned approximately at right-angles to the movement of ice down the Tweed valley and eastwards from the Solway basin. Impressive channels do occur in a few places, however, and indicate two separate alignments of fluvioglacial drainage. For example:

(a) On the north-west fringe of the massif large rock-cut meltwater channels occur at Paston, Hare Law and on Yetholm Law (Maps 1 and 4). The orientation of these channels is roughly parallel with the edge of the massif and the former direction of the Tweed glacier (or northern ice mass) which over-ran that area.

(b) Farther in towards the heart of the massif, in the middle reaches of the Kale and Bowmont valleys, the few channels observed are aligned conspicuously down-valley in a north-westerly direction. In the Bowmont valley there are large channels near Atton Burn and Mow Law (Map 4), and in the Kale valley a deep channel is located in the col west of Hownam Rings (Map 12).

From the sparse evidence of meltwater channels in the west Cheviot area it seems that fluvioglacial drainage followed two separate trends, approximately at right angles to each other. Channels on the north-west fringe of the massif are located at 500, 550 and 750 feet and illustrate a flow of meltwater towards the north-east, roughly parallel with the Tweed valley and its drumlin field. These channels are cut through cols, thereby crossing watersheds in the same way as channels on the north-east flanks of the massif, and were probably formed by the superimposition of englacial streams. Channels in the Bowmont valley are at 700, 750 and 850 feet and their alignment strongly indicates meltwater drainage flowing north-north-west, leading down-valley away from the Cheviot watershed. This suggests that the glacier ice had also formerly flowed in that direction. Elsewhere in the western Cheviots hillsides and cols are notably devoid of meltwater channels, suggesting that the movement of fluvioglacial drainage was predominantly down the pre-existing valleys. If Tweed ice had encroached far into the western part of the massif, the alignment of meltwater streams during the period of deglaciation would have been mostly at right-angles to the main valleys and systems of superimposed channels similar to those in the north-east and south-east Cheviots would probably have become located in cols and valley-heads. It has been suggested elsewhere (Clapperton 1960) that the area south of Hawick at the western extremity of the Cheviot Hills is devoid of meltwater channels because the direction of former ice movement and of subsequent fluvioglacial drainage approximately coincided with that of the major valleys. As a result, most of the meltwater drainage was channelled into the pre-existing routeways. In this fashion it may be argued that the relative absence of meltwater channels in the west Cheviot massif indicates that the former direction of ice movement was predominantly down-valley, related to source areas on the watershed.

The occurrence and alignments of meltwater channels in the Cheviot Hills is perhaps an insufficient basis upon which to conclusively establish the former presence of a local ice cap with its surface sloping down from the high ground in the centre of the massif and from summits adjacent to the west. Nevertheless, it is difficult to explain satisfactorily the absence of meltwater channels over broad areas of the inner massif and to account for the orientation of a small number of channels radiating from the core of the massif if it is assumed that the glacier ice that covered the Cheviots came entirely from external sources.

Evidence from Striations

The following discussion is based entirely on the striations that have been recorded in previous literature on the Cheviot Hills and on the relevant one-inch sheets of the Geological Survey. All of these striations are indicated on Map 12.

Within the territory composed of andesite and granite, the outcrop of rock surfaces is extremely limited. The steep hillsides are generally clothed with a continuous mantle of grass, bracken and heather, and the flatter interfluves at high levels are mostly covered with blanket peat. In places where bedrock does appear at the surface, it is usually in the form of tors or steep crags. Abundant scree litters the slopes below these outcrops, indicating that the rock is susceptible to and has experienced considerably disintegration. There are no smooth polished rock surfaces. In some of the interior head streams, deeply rotted bedrock (overlain by till in places) similarly precludes the preservation of glacial striations. It is therefore hardly surprising that only a small number of striations has been recorded within the igneous massif. One group of these occurs on andesite at the north-west fringe of the massif near Hownam. Since the striations indicate a direction of ice movement towards the north-east, they were undoubtedly formed by ice flowing down Teviotdale towards the Tweed valley. The only other striations on igneous rocks were recorded by Clough (1888), who observed "one well marked set on a fine-grained ash a little over a mile E. of the top of Thirl Moor, the direction is E.S.E. The top of the thick quartz vein, Baker Crag, half a mile S.E. of Carshope, is polished and striated in a direction slightly E. of S., probably by ice action." These striations are clearly related to ice movement outwards from the Cheviot watershed, flowing in a south-easterly direction, an alignment that is parallel with the main river valleys.

Farther west in the Cheviot Hills range, outside the igneous massif, striated rock surfaces appear to be more abundant. For example, Clough observed and recorded striations near the valleys of the North Tyne, the Tarset Burn and the Rede. From the average direction of these striations, Clough indicated that there had been a movement of ice parallel with the principal valleys, generally towards the south-east. In the lower reaches of these valleys and on the more open Fell country adjacent, sets of glacial striations are fairly frequent, and Clough recorded 20 separate observations. With only two exceptions the striations pointed in directions between 3 and 26 degrees north of east; the two exceptions were aligned from west to east. Clough clearly believed that this evidence indicated a movement of glacier ice quite separate from that recorded in the higher parts of the valleys, for he concluded, "This mass of moving ice can have been nothing less than a general ice-sheet, wrapping up the whole district, and covering all the lower elevations of the Cheviot Hills."

The Scottish side of the Cheviot Hills is not nearly so well documented as the Northumbrian area, as no sheet memoirs have been published for it by the Geological Survey. The first geological and geomorphological account of the Scottish side was provided by J. Geikie in 1876 and since then only Common (1953) has further investigated the area (and then only a small section of it). Although he provided few specific examples, J. Geikie discussed the grooved and striated nature of the upper valleys in the Cheviot Hills and firmly believed that glacier ice had flowed outwards from the main watersheds in a direction parallel with the principal valleys. For example, he stated, "In the upper valleys of the Cheviots, the scratches coincide in direction with the valleys, which is, speaking generally, from south to north in Scotland", and in relation to the agent that had produced the grooves and striations, he concluded, "Thus it becomes evident that the denuding agent, whatever it was, that gave rise to these ridges and scratched rock surfaces must have pressed outwards from all the dominant watersheds, and, sweeping down through the great undulating strath that lies between the Cheviots and the Lammermuirs, must have gradually turned away to the east and south as it rounded the northern spurs of the former range so as to pass south-east over the contiguous maritime districts of Northumberland." Striations recorded by J. Geikie and Peach and published on sheets 11 and 17 by the Geological Survey for Scotland (1883) indicate ice movement towards the south-east from high ground north of the Liddel Water on the south side of the Cheviot Hills, while between 5 and 6 miles farther to the north-west, on the other side of the watershed, there is abundant evidence, such as striations, grooves, crag and tail formations, drumlins, etc., showing former ice movement from south-west to north-east (Clapperton 1960).

On the basis of recorded glacial striations it may therefore be concluded that the following ice movements affected the Cheviot Hills:

(a) The central range of hills, forming the present watershed, supported glacier ice that flowed south-eastwards and northwards-north-eastwards, presumably guided by the upper reaches of the principal valleys.

(b) The northern flanks of the hill range were smothered by a powerful stream of ice that flowed north-eastwards down Teviotdale and curved eastwards in the lower Tweed valley.

(c) The southern flanks of the hill range were covered by an extensive sheet of glacier ice that moved in a direction slightly north of east and which appears to have come via the Tyne gap and the low Fell country north of it.

Although striations have not been recorded in the east Cheviot massif and only a small number has been reported from the western part of the massif, it may be inferred that if the range of hills to the west supported substantial masses of local ice, then the much higher ground in the centre of the Cheviot massif would very probably have nourished an independent ice cap also.

Evidence from the Glacial Drift

In addition to fluvioglacial phenomena and striations another useful indicator of the former direction of ice movement is the occurrence of erratic stones, either freely on the ground surface or within the glacial drift. The direction of former ice movement, in turn, suggests the probable source of the glacier mass. The early Geological Survey officers such as Gunn and Clough were intensely aware of the importance of recording erratic stones in the drift and their accounts of erratics, published in the various sheet memoirs, provide considerable evidence on the relative movements of the ice masses once present in the Cheviot Hills.

In the northern part of the Cheviot massif Gunn and Clough (1895) recognised several erratic stones in the Elsdon Burn valley. The till there "consists of a stiff chocolate clay with boulders of porphyrite, well-rounded hard green, probably Silurian sandstone, basalt, Carboniferous sandstone, vein quartz, and purple and yellow quartzites". Gunn and Clough stated that this material had "clearly come over from the west country by the head of Tuppie's Sike, a height of over 900 feet". This suggestion is in keeping with the evidence from meltwater channels and striations, indicating that Tweed ice encroached upon the massif at least to the vicinity of the Elsdon Burn. Erratics were not observed farther in towards the centre of the massif, however. In a similar fashion, many erratics of yellow Carboniferous sandstone and Silurian sandstone were observed on the north-east flanks of the massif, up to a height of 800 feet (on the slopes of Akeld Hill and White Law, and in the Humbleton valley) but were not reported from more central parts of the massif (Gunn and Clough, 1895). In an earlier memoir Clough (1888) described erratics on both the north-east and south-east flanks of the Cheviot massif. For example, he observed that "near the margins of the Lower Old Red area, a clay containing very many Carboniferous, &c. rocks, and essentially of a foreign origin, has advanced on to it; e.g. on the N.E. margin by Skirlnaked, on the E. by Brand's Hill and Middleton Crags, and on the S.E. by Hazeltonrig and Ewe Hill, and on the S. a quarter of a mile S. of Lord's Seat. ..... The Skirlnaked clay has probably come from the N., for in the adjoining Map (110 S.W.), a track of foreign boulders can be seen crossing the Porphyrite hills from the N. towards this locality. The Lord's Seat clay, on the other hand, appears to have come from the S.W., for many boulders of the Acklington basalt dyke have been carried on to the N. side of the Netherton Burn, ...". He concluded from this evidence that "In all probability both the E. and the S. margins have thus once been overridden by foreign ice up to a height of about 1,000 feet." The directions of former ice movement indicated by these erratics fully agree with the evidence provided by meltwater channels in the same areas. The channels also endorse Clough's conclusion that "foreign" ice encroached to at least 1,000 feet on the northern, eastern and southern margins of the Cheviot massif.

Discussing the glacial drift present on the southern flanks of the Cheviot Hills range Miller and Clough (1887) observed that "The boulders of far-derived origin are all from the west"; for example, stones from Galloway, Criffel, East Cumberland and the Lake District were recorded in the valley of the North Tyne. The alignment of this valley and the fairly low ground at its head appears to have allowed glacier ice encroaching from the west to flow relatively unimpeded in an easterly to south-easterly direction. Similar erratics occur on the western extremity of the Simonside Hills at a height of 1,000 feet. All of these western erratics endorse the direction of ice movement suggested by striations. Erratic stones from the Cheviot igneous area appear to extend no farther west than Redesdale. In this locality, Miller and Clough noted that the igneous erratics "are spread over the greater part of the ground that lies east of the Rede .....", but, "on the west side of the valley, away from its central hollow, not a fragment has been detected". Furthermore, these authors also observed that while these erratics are plentiful in upper Redesdale, in lower Redesdale they have become few and very small. This evidence is clearly crucial in assessing the directions of former ice movement in this part of the Cheviot Hills. Since the nearest outcrop of "porphyrite" occurs north of the Rede valley, it is suggested that the glacier ice responsible for transporting fragments of that material into upper Redesdale flowed southwards from the Cheviot watershed. This movement of ice was either too weak to penetrate as far as the western side of Redesdale, where ice from the west was probably present, or else the latter ice subsequently removed any igneous erratics from that side of the valley. No erratics stones from western sources, such as south-west Scotland and the Lake District occur in the drift of Redesdale, even although this valley lies only 10 miles east of the North Tyne valley, where they are abundant. In conjunction with the striations (which point down-valley in Redesdale) this evidence suggests that local ice from the Cheviot Hills occupied upper Redesdale and was sufficiently powerful to prevent this valley from being overridden by the western ice stream from the Solway basin and by ice from the Tweed valley.

Similar evidence suggesting that the upper parts of some of the valleys east of Redesdale were also occupied by local ice rather than by western ice occurs in the following observation by Clough (1888); "..... the massive blocks of black glossy Porphyrite which made such a show in the drift in the glens a mile N. of Whiteburnshank, and on the E. side of the Usway below Fairhaugh, and in many other places on the W. side of Bloodybush Edge, Cushat Law, and Wether Cairn ridge, do not appear to have been carried over to its E. side". A further indication that local glacier ice was present in the Cheviot

Hills occurs in the vicinity of Bloodybush Edge, where many granite boulders are at a height of about 1,900 feet, near the "Shivering Stone". Since bedrock at that point consists entirely of andesite, the granite is most likely to have come from the nearest outcrop, which is approximately $1\frac{1}{4}$ miles to the north-north-east. A movement of glacier ice south-south-westwards from The Cheviot is thereby suggested.

Since investigations in the east Cheviot area subsequent to those by Miller and Clough did not add to the list of recorded erratics, a close inspection was made of drift sections occurring in the upper valleys. During the investigation of these sections, the first problem to be considered was that concerning the validity of interpreting the material as glacial drift. Since some authorities have considered the Cheviot massif as an ice-free enclave, the drift lying in the upper valleys could have been interpreted as solifluction debris. The steep slopes, the presence of friable and rotted bedrock, in conjunction with periglacial climatic conditions would certainly have produced extensive deposits, filling the valley bottoms. Close inspection of the drift in all the valleys suggests that the deposit is glacial till, however, for the following reasons:

(a) The majority of stones have been derived from fresh, unrotted bedrock, whereas soliflucted debris might reasonably be expected to contain a much higher percentage of the rotted pre-existing bedrock.

(b) Most solifluction deposits are composed of flattish, often elongated stones that are sharply angular, but the deposit under consideration consists chiefly of cuboid stones and boulders that are conspicuously sub-angular in shape. A few boulders are even sub-rounded. Sub-angular stones normally occur in glacial till where they tend to predominate over stones of other shapes.

Photograph 7.a Till overlying deeply rotted bedrock in the upper Breamish valley.

(c) The majority of solifluction deposits are composed of stones derived from underlying bedrock. Where this is of uniform lithology, the stones in the deposit are all of the same lithology. Bedrock underlying the drift in the upper valleys of the Cheviot massif is mostly uniform, consisting of either granite or andesite. The drift is certainly composed entirely of igneous material, but a great variety of igneous types is represented. Indeed, the lithological variety in the drift is too great for the material to have been derived solely from underlying bedrock.

(d) In most solifluction deposits the stones can clearly be seen to lie with their long axes orientated down-slope, but such an arrangement appeared to be lacking in the drift sections under consideration.

(e) In most solifluction deposits derived from underlying bedrock, the stones are usually relatively uniform in size, although the calibre might vary with depth and/or position on the hillside. The drift deposits in the upper valleys of the Cheviot Hills contain a heterogeneous admixture of stones varying in size from less than one inch to over two feet.

(f) A certain degree of layering or fissility is often present in solifluction deposits, but such structures are entirely absent from the drift under consideration.

For the foregoing reasons, the drift that lies in the upper valleys of the east Cheviot area is interpreted as glacial till and not as solifluction debris. The drift sections (Photograph 7.a) to which particular attention was paid are located in the upper reaches of the following valleys and/or their tributaries; the Lambden Burn, the Harthope Burn, the river Breamish (tributaries Ainsey Burn, Hareshaw Cleugh) and the Shank Burn. Up to 50 feet of bright red till occurs in places; boulders 6 to 8 feet in size are contained in the till in the Lambden valley. In some of the gullies tributary to the upper Breamish, several feet of till overlies deeply rotted bedrock. The purpose of the investigation was to search for stones that were not of igneous origin, for it was considered that had the area been overridden by extraneous glacier ice, at least a small number of erratic stones from the sedimentary rocks west of the Cheviot massif would have been present in the drift. As far as could be ascertained, each section was completely devoid of sedimentary erratics. It is therefore suggested that since the glacial drift consists entirely of igneous material, it was probably deposited by glacier ice that was derived locally.

Additional evidence indicating that the central part of the Cheviot massif was not overridden by the general "mer de glace" from the west comes from a consideration of the Cheviot granite in the drift. The granite bedrock occurs over an area of approximately 22 square miles, and since it is a conspicuous type of pink granite, it should occur fairly liberally and obviously in the drift of Northumberland had the granite area been covered by the Tweed/Solway ice sheet. However, when Smythe (1912) considered the drift deposits of northern Northumberland, he was obliged to conclude, "curiously enough, the characteristic augite granites of that region are of comparative rarity in the drift deposits of the county; what specimens have been transported by ice occur to the east of Cheviot .....". Since the granite area descends on the north to a level below that known to have been covered by the Tweed glacier, it is to be expected that some erratics of this material would occur farther to the east, as Smythe observed, but it is suggested that the relative scarcity of Cheviot granite erratics in the drift deposits of north-east England can be explained by the fact that the central Cheviot massif was either ice-free or else covered by a local ice cap that did not spread far beyond the granite area. Since the occurrence of glacial till indicates that glacier ice was

Photograph 7.b The Bizzle corrie. The moraine is visible near the corrie mouth. Prominent tor on the upper flanks of The Cheviot.

Photograph 7.c	Large block on morainic ridge at the mouth of the Hen Hole.
Photograph 7.d	View towards the exit of the Hen Hole.

Photograph 7.e	Roches moutonnées in the Lambden valley (Dunsdale Crag).
Photograph 7.f	View down the Bizzle outwash apron towards the ice-eroded side of the Lambden valley.

formerly present in the upper reaches of the valleys, the latter alternative is considered the more probable.

Evidence from the Bizzle and Lambden Valleys

Evidence supporting the concept of a local ice cap on the Cheviot massif is twofold in the Bizzle and Lambden valleys (Maps 4 and 5), and concerns landforms of both glacial erosion and glacial deposition.

(a) Erosional Forms: If the central Cheviot area had ever been overridden by extraneous ice from the west, the pre-existing valleys would have been modified no more (probably less) than the valleys known to have been affected by that ice mass elsewhere. In the latter instances, the greatest modification to the cross-profiles of the valleys has apparently resulted from the deposition of till - particularly on the lee side of the valleys in terms of ice movement (J. Geikie 1894; Clapperton 1960). The valleys leading northwards from The Cheviot to the Lambden valley do not show such characteristics of asymmetry, however. Indeed, the valley sides are markedly oversteepened and appear to have been widened by erosion rather than constricted by deposition. This is particularly clear in the valley of the Bizzle Burn. From 2,400 feet to 2,000 feet, the V-shaped valley is predominantly grass covered and characterised by moderately steep slopes; bedrock outcrops are few. From 2,000 feet to 1,700 feet, however, the valley walls diverge in gentle curves, so that an amphitheatre-like form is assumed (Photograph 7.b). The valley floor is broad and relatively flat between 1,500 and 1,300 feet, and precipitous bare cliffs rise up to 300 feet in height, particularly on the east-facing side. The present stream cascades over a prominent rock-step into the flat-floored section of the valley below. The foregoing characteristics of valley form are unlikely to have been produced by stream erosion or by glacier ice moving over the area from the west. The Bizzle amphitheatre is therefore interpreted as a corrie, developed during a period when a local valley glacier flowed northwards. A similar, but less well-defined form occurs in the upper reaches of the College Burn. The stream headwaters begin in a broad embayment where steep sides rise over 300 feet above the flat floor lying between 1,900 and 2,000 feet. The valley becomes more constricted in a westerly direction, but the walls steepen and bare crags become characteristic. The valley floor descends irregularly over two rock-steps into the section named the Hen Hole, where rock buttresses tower nearly 300 feet above the more gentle lower slopes (Photographs 7.c and 7.d). As in the Bizzle valley, such characteristics are more logically explained by the former presence of a local glacier flowing outwards from the broad dome of The Cheviot, rather than by the incursion of ice from the west. Although the valleys of the Bellyside and Goldscleugh (Map 5) Burns are less spectacular, they too probably owe their oversteepened forms to local ice.

Much of the Lambden Burn valley is distinctly asymmetrical in cross-profile. The profile of the upper section of the valley, however, from its head on the slopes of The Cheviot to the point at which it turns westwards from its northerly course, is quite symmetrical. The most prominent asymmetry occurs along the middle section of the valley which is aligned almost due east-west. The steeper side faces south. Although asymmetry in valley cross-profiles may be due to several factors, including structure and aspect, there is certain evidence in the Lambden valley suggesting that ice erosion was most important in this instance. For example, in the vicinity of Dunsdale Crag, the south-facing side of the valley is an almost vertical wall of rock and exhibits distinct roche moutonnée forms (Photographs 7.e and 7.f). Relatively smooth slabs of rock face up-valley, but have abrupt craggy re-entrants facing

Photograph 7.g	Section exposing till on left bank of Lambden Burn.
Photograph 7.h	View down the Hen Hole (College) outwash apron; left wall of the valley oversteepened by ice erosion.

down-valley. While it seems likely that structure influenced the development of these forms, it is suggested that ice erosion was chiefly responsible for their genesis. They are currently losing their form as active screeing continually dislodges large blocks from the craggy lee faces. It is no doubt significant that this evidence of relatively powerful ice erosion occurs immediately down-valley from the point at which the Bizzle/Bellyside valleys jointly enter the Lambden valley. In direct contrast with these erosional features, the less steep north-facing side of the Lambden valley is mantled with up to 40 feet of glacial drift (Photograph 7.g). On the basis of the foregoing evidence it is suggested that glacier ice flowed northwards down the Bizzle, Bellyside, Goldscleugh and upper Lambden valleys. On being obstructed by the south-facing side of the Lambden valley, the ice from the Bizzle/ Bellyside area in particular effected considerable erosion as it was diverted westwards.

The evidence of ice erosion in the College valley is less distinct and the fact that this valley follows the line of a fault (or crush) must also be considered in any assessment of ice erosion. Nevertheless, it is clear that asymmetry in valley form occurs at the point where the valley bends northwards from its earlier westerly course. Here, the western side of the valley has a steep concave/uniform slope, whereas the eastern is distinctly convex (Photograph 7.h), and it is possible that a certain degree of "undercutting" by ice erosion on the west wall has been responsible for the asymmetry in profile. A similar situation occurs in the vicinity of Southern Knowe where the oversteepened western side of the valley contrasts with the more gentle eastern side, precisely where the Lambden ice stream would have entered the College valley and effected such oversteepening.

Photograph 7.i The Bizzle moraine, rock-step in the background.

Photograph 7.j Section in the Bizzle moraine.

Photograph 7.k Large blocks near the exit of Hen Hole. The dark vegetation in the middle distance grows on the morainic ridge extending across the valley. View is looking down-valley.

If the inner parts of the Cheviot massif had been overridden by ice from the west, the features of ice erosion just described would not have been formed. It is therefore suggested on the basis of erosional forms that the glacier ice that formerly occupied the Lambden, Goldscleugh, Bellyside, Bizzle and College valleys flowed down these valleys away from catchment areas heading on The Cheviot.

(b) Depositional Forms: The Bizzle corrie contains mounds and ridges that are unique in terms of their composition and arrangement in the eastern Borders. Varying in height from 5 to 40 feet, these landforms are assembled in a crescentic formation near and parallel with the east side of the corrie, but show no dominant pattern in the centre of the broad corrie floor (Photograph 7.i). They lie between 1,300 and 1,500 feet. The internal composition of the mounds is revealed where the present stream has cut sections through them (Photograph 7.j). They are composed entirely of loose debris consisting chiefly of subangular rubble, the stones varying in size from grit to blocks up to 6 feet in diameter. Large blocks also litter the surface on and between the mounds. Whereas the scree cones at the base of the corrie walls show some degree of gradational sorting, the debris exposed in the mounds appears to be much more heterogeneous. The hummocks are therefore interpreted as a moraine. Since the mounds and ridges on the eastern side of the corrie have evidently accumulated as scree or avalanche cones, banked up against a mass of firn or ice lying in the corrie, it seems probable that much of the morainic debris was derived in a similar fashion; the material perhaps travelled only a short distance down-valley from its source. Owing to the relatively small scale of the corrie and its catchment area, it is suggested that the moraine was built on and around a mass of snow and firn that possibly lay on a base of true glacier ice which had a small degree of forward movement. There is no evidence upon which to base an absolute age for the moraine, but relative to other glacial phenomena in the eastern Borders, it may date from late-glacial times and is unlikely to be younger than Zone 1. Even if this tentative age interpretation is mistaken, the presence of the moraine implies that the northern slopes of The Cheviot formerly received a supply of snow sufficient to nourish a small glacier towards the end of the last glaciation. It is therefore reasonable to assume that in the period leading up to and during the last maximum glaciation a much greater volume of glacier ice was able to accumulate in this vicinity. The broad ridge littered with large blocks aligned across the College valley a short distance beyond the mouth of Hen Hole, at approximately 1,300 to 1,350 feet (Photographs 7.k and 7.c), probably represents the same glacial phase; extensive aprons of outwash slope steeply down-valley from the moraines in both valleys (Photographs 7.f and 7.h).

The occurrence of landforms of glacial erosion and deposition in the College, Bizzle and Lambden valleys suggests that during the last maximum period of glaciation and probably a subsequent cold oscillation, the central Cheviot massif supported local glacier ice and was neither an ice-free enclave nor overridden by external ice from the west.

The Significance of the Tors and Deeply Rotted Bedrock

While it is generally agreed that in most areas formerly overridden by glacier ice the preglacial and interglacial rock waste was completely removed, allowing the ice to attack the fresh bedrock beneath, residuals of weathered and rotted rock remain in some places because of the selectivity of glacial erosion (Flint 1957). Deeply weathered bedrock has been reported from North-East Scotland (Fitzpatrick 1963), North-West Scotland (Godard 1965), and the Cairngorm Mountains (Sugden 1965). Each area also contains abundant evidence of the former presence of glacier ice and it is evident that the presence of deeply rotted bedrock is not necessarily indicative of unglaciated enclaves, particularly since till frequently overlies the rotted material. Similar reasoning can be applied to the deeply rotted bedrock which occurs in the Cheviot Hills, for it is present at levels below the higher subglacial meltwater channels and must therefore have been overridden by moving glacier ice. For example, the rotted bedrock outcropping in the valleys of the Lambden, New, Hawsen, Broadstruthers and Common Burns is below the level of the subglacial meltwater channels located in the cols west and south of Broadhope Hill. For this reason the central part of the Cheviot massif cannot be considered as an unglaciated area simply because of the presence of soft, deeply rotted bedrock.

Although Linton (1949, 1955, 1959) has argued that tors are unlikely to have survived in areas covered by glacier ice, Sugden (1965) has recently suggested that tors in the Cairngorm Mountains remained relatively intact even after the passage of a powerful ice sheet and a subsequent plateau ice cap. In the Cheviot Hills well-defined tors occur at various levels between 1,300 and 2,500 feet, but two of the most impressive lie at 1,725 and 1,750 feet (Great Standrop and Little Standrop). Although there are no meltwater channels or till deposits above this height, glacial erratics were transported to a level of 1,900 feet on Bloodybush Edge, only 3½ miles to the south-west (Clough 1888). This evidence indicates that the Standrop tors must have been overridden by moving glacier ice, which did not destroy them. Since the outer parts of both tors are composed predominantly of large blocks of granite that have clearly weathered out in situ along joints, it is difficult to assess whether they represent the most recent stage in their development, having weathered out from solid bedrock subsequent to deglaciation, or if they are much earlier forms which survived the passage of glacier ice. The first alternative seems unlikely in view of the absence of loose gravel that could reasonably be expected to have resulted from post-glacial weathering on a scale sufficient to have loosened the outer "rind" of the tors. The second alternative is perhaps valid when it is considered how relatively inefficient glacier ice would have been as a powerful agent of erosion at a height on the massif of 1,725 to 1,750 feet. Furthermore, at the onset of glacial conditions the tors and the crevices between the loose blocks would probably have become smothered with permanent snow and ice before being overwhelmed by moving glacier ice, and in this way survived relatively intact as rigid blocks beneath the weak glacier ice. Between 2 and $2\frac{1}{2}$ miles to the north-east, prominent tors occur at a much lower level on the shoulder of ground above the Harthope valley. The most conspicuous of these, Housey Crag, lies at 1,275 to 1,300 feet; Long Crags are at 1,300 to 1,325 feet and Langlee Crags occur at 1,300 feet. In contrast to the Standrop and Coldlaw tors there is an almost complete absence of loose blocks and debris on and around the tors above the Harthope valley. Furthermore, the latter are distinctly elongated and Common's suggestion (1953) that they were streamlined by glacial erosion is a possible explanation. Since ice movement would probably have been more powerful at this lower level, farther towards the edge of the massif, a slight degree of glacial erosion may well have removed any loose clitter of rock originally surrounding the tors.

There is therefore good reason to believe that deeply rotted bedrock and tors in the Cheviot massif were formerly smothered beneath moving glacier ice and the tors remained as prominent features in the landscape. For this reason the latter should not be considered as indicators of unglaciated areas in the eastern Borders. Indeed, it is perhaps because the tors and rotted bedrock were covered by a relatively thin and weak cap of local ice that they have survived, because exposure to severe periglacial conditions in ice-free enclaves surrounded by glacier ice might have entirely removed them.

Evidence from Present Precipitation

Measurements of annual precipitation falling on the Cheviot massif are generally lacking, but records were kept in the vicinity of Broadstruthers (Map 5) during the years 1875 to 1895 inclusive (British Rainfall). Broadstruthers (1,000 feet) is located roughly 3 miles north-east of The Cheviot summit, but since the height of the rainfall recording station is given as 1,692 feet, (corrected to 1,694 feet in 1891) and the top of Broadhope Hill (lying approximately one mile south-west of Broadstruthers) lies at 1,694 feet, it is clear that the rain gauge was located on the summit of that hill. The table below represents the annual precipitation falling at the Broadstruthers station for the 20 years during which records were kept. (The figure of 118.13 inches for the year 1888 seems so anomalous compared with those recorded for the other years that it is questioned in the British Rainfall records and has been omitted from the present considerations.)

ANNUAL PRECIPITATION AT BROADSTRUTHER

Year	Amount (inches of rainfall)	Year	Amount (inches of rainfall
1875	36.50	1885	33.13
1876	79.00	1886	59.75
1877	64.13	1887	46.38
1878	50.19	1888	
1879	77.63	1889	56.50
1880	45.75	1890	44.63
1881	72.75	1891	61.38
1882	55.13	1892	76.50
1883	44.13	1893	54.50
1884	45.00	1894	85.50
		1895	84.00

For the 20-year period, the average annual precipitation was 58.57 inches. Lying directly on the lee side of the broad bulk of The Cheviot, only two miles distant, the top of Broadhope Hill is almost 1,000 feet lower in altitude. It seems reasonable to assume, therefore, that a substantially greater amount of precipitation falls on The Cheviot. Indeed, on many occasions the summit of Cheviot and Hedgehope Hill (2,348 feet) can be observed to be enshrouded in cloud while adajcent hills at slightly lower levels remain quite cloud-free. It is estimated that an annual average precipitation of at least 65 inches probably falls on The Cheviot. The significance of this estimation is perhaps most relevant in the context of the following discussion.

The Formation of an Ice Dome

With the onset of colder climatic conditions leading to the last glacial maximum, the high ground of south-west Scotland and the Tweedsmuir massif would probably have become the first catchment areas for glacier ice in the Southern Uplands of Scotland; both areas rise over 2,500 feet in height and presently receive an annual average precipitation of over 80 inches. It seems unlikely that the Cheviot Hills, with their lower precipitation, developed glaciers or permanent snowfields during the earlier stages of the glaciation because the snowline in Britain appears always to have risen rather sharply towards the east. However, as the North Atlantic became progressively cooler (e.g. 8 to 9 degrees centigrade cooler off south-west Ireland; Manley 1959), and ice caps and permanent snowfields developed in Britain, the snowline must have continually lowered as a result of decreasing annual temperatures. Manley (1955) has suggested that the presence of large masses of glacier ice would imply that at sea level in eastern Scotland the summer averages of temperature were only a few degrees above freezing point, while in winter "they may have been between

10^{o} and 15^{o}F." Under such conditions it seems very probable that even although annual precipitation was somewhat less during the glacial periods (this is by no means certain, however) sufficient would have fallen on the Cheviot Hills range and the Cheviot massif to sustain permanent snow when the snowline had become sufficiently depressed. There are no records to indicate which air streams presently bring most snow to the Cheviot massif, but the writer is aware that some of the heaviest snowfalls are related to the influx of easterly air masses. Nevertheless, Arctic air from the north-west and north also cause substantial falls of snow in the eastern Borders. During the onset of the last glacial maximum, a time must have come when the North Sea became partly or totally frozen over, thereby reducing the potential source of moisture for easterly air streams. However, since most of the precipitation presently falling on the Cheviot massif seems to be associated with westerly air masses, it is likely that a similar situation occurred during glacial times, except that most of the precipitation fell as snow. Comparing the development of glaciers in west and east Scotland, Linton (1959) emphasised that because substantially lower temperatures prevailed in the east, it was possible to generate a corrie glacier in the eastern Highlands "with some 30 inches less in the way of annual precipitation than in the mountains of the western seaboard".

According to Manley (1955), it may be reasonable to say that in a disturbed temperate climate a summit 1,000 m. broad is likely to retain a snow cap and form a "dome" if it rises 200 m. above the local firn line. Since the summit of The Cheviot is approximately 2,900 m. broad, it is likely to have developed an ice dome when the firn line was considerably less than 200 m. from the summit, according to Manley's reasoning. Once a permanent dome of firn and ice had become established, it would have continued to grow higher and higher above the snowline with each annual increment of precipitation. Rime deposit would also have formed a significant part of the accumulation, in view of the damp and unsettled climate prevalent during the glacial period (Manley 1951, 1959). Some of the broad summits farther west in the Cheviot Hills, such as Peel Fell (1,975 feet) and Carter Fell (1,815 feet), and the summits in upper Teviotdale, such as Greatmoor Hill (1,964 feet), Cauldcleuch Head (1,996 feet), Tudhope Hill (1,961 feet) and Wisp Hill (1,950 feet) probably supported considerable domes of snow and ice.

It is difficult to assess how far eastwards glacier ice had advanced from the major catchment areas of the west by the time that ice was emanating from summits in the eastern Borders, but it seems reasonable from a climatological standpoint to suggest that the central part of the Cheviot massif and major valleys such as the Rede, Coquet, Kale and Bowmont were always occupied by local glacier ice. The local ice most probably originated in broad domes of ice that built up on the high, broad summits along the Border watershed.

Conclusion

In the preceding paragraphs considerable evidence has been presented to support the concept that the Cheviot massif and much of the Cheviot Hills range nourished independent centres of glacier dispersion during the last maximum period of glaciation in the eastern Borders. The orientation of landforms such as meltwater channels and glacial striations, together with the occurrence of glacial erratics and features of glacial erosion and moraines, strongly suggests that glacier ice flowed somewhat radially from the core of the Cheviot massif. Similar phenomena illustrate a distinct movement of ice away from the Cheviot Hill summits, following the alignments of the principal valleys on either side of the watershed. Meanwhile, the peripheral hills at lower elevations, particularly below 1,000 feet, were overwhelmed by the Tweed glacier north of the watershed and by the Solway ice south of the watershed. Although precipitation figures are generally lacking for the Cheviot area, the fact that The Cheviot probably receives as much as 65 inches annually at present suggests that, as lowering temperatures in glacial times depressed the snowline, The Cheviot would have been able to support a considerable dome of glacier ice. Some of the summits farther west appear also to have developed small ice caps. It is therefore concluded that sufficient evidence exists in the Cheviot area to strongly support the concept that an ice cap formerly existed on the Cheviot massif and on areas adajcent to the west.

CHAPTER 8.

THE DEGLACIATION OF THE EAST CHEVIOT AREA

In the preceding chapters the landforms produced during the dissolution of the last period of maximum glaciation in the east Cheviot area have been discussed systematically. Certain conclusions concerning the nature of the downwastage were put forward at the end of each chapter, but as yet, the general pattern of events over the entire area has not been outlined. It is therefore the purpose of this chapter to summarise the conclusions arising from the previous chapters and to consider broadly the deglaciation of the east Cheviot area.

The Extent and Movement of Glacier Ice

With the onset of cold climatic conditions leading to the last maximal development of glaciers in Southern Scotland and Northern England extensive ice caps probably accumulated first of all in the west and in the Tweedsmuir massif. Outlet glaciers from the ice caps spread initially down the major valleys radiating from the centres of ice accumulation. Ultimately, as the regional snowline lowered with the increasing severity of the climate, ice domes built up on the broad summits of the higher hills in the Cheviot area. The Cheviot massif in particular, rising to an extensive plateau over 2,500 feet high, appears to have supported a considerable ice cap. From this local area of accumulation and similar centres to the west, such as Peel Fell and Carter Fell and especially the broad hill-tops in upper Teviotdale, tongues of glacier ice spread down the main valleys. By this time, the wide expanse of lowland comprising Teviotdale and Tweeddale, had become filled with an enormous mass of glacier ice moving north-eastwards and eastwards to the coast. It was guided between the belts of high ground formed by the Moorfoot/Lammermuir ranges on the north and the Cheviot Hills in the south. Simultaneously, the congestion of ice in the Solway Basin, caused by the confluence of glaciers from the Lake District with those from southern Scotland, produced a stream of ice moving generally eastwards through the Tyne gap and north-eastwards over the southern flanks of the Cheviot Hills. Glaciers nourished on the Cheviot summits became confluent on both sides of the range with the major streams of ice and were diverted towards the north-east in the mid and lower reaches of the valleys they were occupying. East of the Cheviot massif the three masses of glacier ice appear to have become confluent at an altitude of approximately 1,000 to 1,400 feet. At a relatively early stage in the glaciation, the combined streams of ice seem to have flowed directly into the North Sea Basin, but subsequently, ice from the Tweed valley was forced to move powerfully towards the south, apparently by the lateral spread of ice from the Midland valley of Scotland. This caused ice moving across the south-east flanks of the Cheviot massif to swing sharply south-eastwards and join the southerly movement of ice over the coastal province of Northumberland. These directions of former ice movement have been deduced principally on the basis of glacial striations, the distribution of erratics, the preferred orientation and dip of stones in the till and the alignments of fluvioglacial landforms.

<u>The Period of Downwastage</u>

With the gradual return to less severe climatic conditions the snow-line would ultimately have risen above the Cheviot ice cap, leading to a rapid thinning and downwastage of the ice. As a result, extensive systems of melt-water drainage became abundant in the vicinity of the Cheviot massif. The first channels cut on the ground by the meltwater appear to have been associated with the Cheviot ice cap, for they lead north-eastwards and eastwards through cols at the head of the Lambden valley at heights of 1,500 and 1,375 feet; these are the highest meltwater channels in the east Cheviot area and were evidently formed subglacially following the superimposition of englacial streams. Apart from these two prominent features meltwater channels are generally absent or few in number west of The Cheviot and in the central part of the massif. It is suggested that meltwater drainage followed the direction of former ice movement in these areas and flowed down the pre-existing valleys. For this reason few dry channels occur on the hillsides, but much of the drift infill washed out of the valleys was possibly removed by meltwater streams. Ice-contact fluvioglacial deposits are also generally absent from the west and central parts of the Cheviot massif.

In direct contrast with the western and central areas of the massif, the north-east and south-east margins contain a zone of territory, roughly 1 to 2 miles wide, where the hillsides are extensively furrowed by remarkable systems of meltwater channels. In the north-east, vast channels and channel systems were cut into cols through spurs radiating from the massif and in broad valley heads. Frequently up to 100 feet deep in bedrock the channels are aligned with few exceptions parallel with the curve of the massif. This alignment is that followed by the Tweed glacier (or northern ice mass) as it spread out of the Merse and into northern Northumberland. Spurs and valley heads in the south-east contain equally impressive channels and channel systems. These are aligned parallel with the curving margin of the massif and were clearly cut by fluvioglacial drainage directed into this alignment by the western ice mass (which over-rode the south and south-east flanks of the Cheviot massif). A large number of meltwater channels in the east Cheviot area have up/down longitudinal floor profiles and others possess remarkable complexities in form, such as abandoned loops, isolated knolls and reticulate tributary networks. Such characteristics strongly suggest that the channels were cut subglacially. Since many of the meltwater streams were evidently in subglacial positions when flowing across the spurs and in the valley heads, it has been concluded that they reached these positions by being superimposed from englacial courses; the upper and lower portions of the stream courses presumably remained englacial. This conclusion applies chiefly to channels and channel systems aligned parallel with the direction of former ice movement and hence parallel with the slope of the ice surface. A smaller number of channels appear to have been cut during a later period of relatively free downslope movement of meltwater beneath the stagnating ice. Because the maximum depth of subglacial penetration by the meltwater streams appears rarely to have exceeded 350 to 400 feet, it is envisaged that an upper zone of the downwasting ice, approximately 400 feet deep, contained the majority of meltwater streams. As the ice surface lowered in height, commensurate lowering of the zone of meltwater drainage enabled a large number of englacial streams to become superimposed on to the ground beneath.

On the north-east and south-east flanks of the Cheviot massif there are conspicuous upper limits to the concentration of meltwater channels. In the north-east the limit descends south-eastwards from 1,100 feet in the north to 600 feet in the Breamish valley. In the south-east the limit descends north-eastwards from 1,225 feet in the south to 900 feet in the Breamish valley. While these heights are indicative only of the minimal level to which the relevant ice masses over-rode the Cheviot massif, it is considered that they are meaningful in establishing the approximate limits (within 350 to 400 feet) of incursion by the great streams of external ice flanking the massif. These limits are to some extent supported by the distribution of erratics, for the latter rarely occur above 1,000 feet. A small number of channels in the mid-

Breamish valley occur at heights between 800 and 1,300 feet, but do not conform with the alignments so clearly exhibited by the north-east and south-east systems. The Breamish channels are perhaps too few in number to form a valid theory concerning their location and distribution, but it is perhaps significant that they extend no farther eastwards than the upper limits of the north-east and south-east systems which converge in the Breamish valley. It has been suggested that the mid-Breamish channels were cut by meltwater streams associated with the Cheviot ice cap, whose eastern limit extended to approximately 900 feet in the Breamish valley. Since most of the meltwater derived from the Cheviot ice cap in this area would have been directed down the Breamish valley it is not surprising that relatively few meltwater channels occur on the adjacent hillsides.

During the period when meltwater streams were flowing at the 900 to 1,200 foot level, it is evident that drainage from the north-east, south-east and mid-Breamish area converged in the Breamish valley. The course then assumed by the drainage remains conjectural, but it seems reasonable to suggest that the ultimate route into which the combined meltwaters became directed was roughly south-eastwards over the Fell Sandstone cuesta which overlooks the Hedgeley basin. Had this drainage ever come in contact with the ground, enormous meltwater canyons would undoubtedly have been formed. The only channel cut through the cuesta is at a much lower level and was initiated by a later period of meltwater drainage. It is therefore concluded that the vast volumes of meltwater escaped south-eastwards englacially.

In the north-east Cheviots a considerable amount of deposition was accomplished by the great systems of fluvioglacial drainage. Prominent eskers and kame terraces occur liberally in areas where the hillsides slope less steeply and although occasional mounds of sand and gravel are present up to 950 feet, the vast majority lie below 650 feet. In many instances the eskers appear to have been deposited by streams issuing from subglacial channels and therefore must have been formed subglacially. Some eskers that are not obviously related to subglacial channels are covered by a thin veneer of ablation till or by large, sub-angular blocks; they too were presumably formed subglacially. The larger kame terraces and probably some eskers appear to have been deposited in subaerial cavities in the ice or at its margin, however, for the terraces in particular are often much too extensive to have been contained within subglacial caverns. It seems unlikely that caverns of the enormous dimensions that would be required could ever have existed beneath stagnant ice in this area without the roofs collapsing through lack of support. Detailed field mapping clearly resolved the apparent confusion of mounds and ridges into remarkable networks of eskers and kame terraces aligned generally south-eastwards in harmony with the meltwater channels on adjacent hillsides. The esker/kame terrace complexes were thus deposited by immense systems of fluvioglacial drainage also directed towards the south-east.

Apart from four small hummocks of sand and gravel, eskers and kame terraces are quite absent from the south-east Cheviots. This remarkable contrast with the north-east appears to have a threefold explanation (Chapter 9), the most important factors perhaps being: (a) the relatively free flow of meltwater drainage down the Aln valley (orientated in the same direction as meltwater drainage) enabling debris to be transported into a proglacial environment where it accumulated as deltaic sediments in a glacial lake, and (b) the fact that the amount of englacial detritus contained within the western ice mass seems to have been minimal, thereby severely restricting the source material for ice-contact deposits. The implications of this latter theory will be more fully discussed in the following chapter.

With the progress of downwastage the zone of meltwater flow in the ice utlimately impinged on the Fell Sandstone ridge that forms the Breamish/Aln watershed. As a consequence, the vast canyon of Crawley/Shawdon Dean began to be cut subglacially. It is remarkable that all of the meltwater directed into the Breamish valley from the north-west and south-west converged on this channel. The distinct up/down floor profile indicates that an enormous stream flowed through the channel under considerable hydrostatic pressure. Since channels entering the Breamish valley from the south-west terminate at approximately 450 to 425 feet (Middledean and Fawdon Dean), roughly the height at which Crawley/Shawdon Dean began to be cut, it may be suggested that the inception of the latter led to the abandonment of the former. Indeed, the channels at Great Ryle indicate that meltwater streams from the south-west began flowing down the Aln valley at that time. North of the Breamish, south-easterly directed meltwater continued in the same alignment and flowed subglacially through the Crawley/Shawdon channel.

As downwastage continued the strong control exerted by ice in directing meltwater drainage without regard to the slope of the ground appears to have been relaxed. This is particularly clear in the north-east where water from ice-free hillsides was able to penetrate directly downslope beneath the ice margin. A distinct group of channels was formed roughly at right-angles to the group cut during the period of south-easterly drainage on the hillsides. Since the majority of the second group occur below 600 feet, it is suggested that this marked change in the direction of fluvioglacial drainage took place when approximately 600 feet of glacier ice covered the lower slopes of the north-east Cheviots and the fringing basins. From the way in which the drainage was able to run freely downslope beneath the ice margin it may be inferred that the entire mass of glacier ice had become more or less stagnant and was slowly dissipating in situ. Despite this fact, it seems that meltwater streams were still unable to erode to a level below 300 feet, or to escape northwards down the Till valley. This suggests two things; (a) that thicker glacier ice lay to the north and still directed fluvioglacial drainage towards the south-east; (b) that below 300 feet the glacier ice was too compacted to allow the establishment of meltwater tunnels, or more probably, an englacial water-table had come into being, controlled by the level of outflow through the Crawley/Shawdon channel. Commensurate downwastage of the western ice mass enabled meltwaters to cut channels in the vicinity of Biddlestone before escaping englacially (presumably) down the valleys of the Netherton and Foxton Burns.

As the glacier ice encircling the Cheviot massif downwasted, the crests of the Fell Sandstone cuestas would have become ice-free relatively early since they rise to 1,000 feet in places. The progressive emergence of these cuestas probably caused the fragmentation of the "mer de glace" east of the Cheviot massif, and the lobe of ice occupying the basins north of the Breamish must have become isolated from glacier ice overlying the coastal province farther east. Similarly, the western ice mass presumably withdrew from its former confluence with the northern mass as the Breamish/Aln watershed became ice-free. The high escarpments between Rothbury and Alnwick must also have severed the western ice mass from glacier ice occupying the coastal lowland farther east. There is clear evidence indicating that a glacial lake formerly occupied the mid Aln basin following the withdrawal of glacier ice from it. For example, laminated clay occurs to the surface over much of the basin and the surfaces of three prominent deltas slope to a level of approximately 200 feet. The lake surface appears to have remained at about 200 feet, and the outflow drained down the Aln valley. The final drainage of the lake probably coincided with the further decay of ice in the Shipley-Alnwick area, which terminated the supply of debris that was being poured into the Aln basin. The lake overflow waters could then cut an outlet eastwards through the drift so that the channel occupied by the present river Aln became established. When the lake drained away, considerable volumes of water must still have been flowing through the Crawley/Shawdon channel because the Shawdon delta has been deeply dissected by a stream much larger than the tiny Shawdon Burn. It is therefore necessary to consider events occurring in the Hedgeley Basin at that time.

As the Fell Sandstone ridge, forming the Breamish/Aln watershed, emerged from the ice, a certain amount of marginal recession appears to have occurred in the Hedgeley Basin. This event must have taken place when melt-water from the north was still flowing south-eastwards because a considerable expanse of water became impounded as an ice-dammed lake between the northern ice margin and the Fell Sandstone escarpment. Up to 10 feet of laminated clay and silt accumulated in this lake and a large delta pitted with kettle holes and bounded on three sides by ice-contact slopes was constructed at the ice margin. Since the surface of the delta was built to a level of between 300 and 325 feet, the Crawley/Shawdon channel was clearly utilised as an outlet and thereby controlled the maximal lake level. The mid Aln lake must therefore have drained away before water from the Hedgeley lake ceased to flow through the Crawley/Shawdon channel. During the period of its existence, the Hedgeley lake appears to have extended northwards within the ice as an englacial "lake" or water-table, because the crests of a large number of eskers aligned south-eastwards have been built to roughly the 300-foot level. It seems likely that some were formed subglacially, but nearer the ice margin the streams probably flowed in open ice-walled canyons. The size and abundance of kettle holes and dead-ice hollows certainly indicate the highly fragmented nature of the glacier margin. The wide gravel plain in the Breamish valley between Ingram and Low Hedgeley probably dates from the Hedgeley ice-lake period, but this is by no means certain. The lake ultimately drained away rapidly when further downwastage of the northern ice opened a route northwards down the Till valley. The Crawley/Shawdon channel became abandoned.

On the lower slopes of the north-east Cheviots below 300 feet small groups of eskers spread out in easterly and north-easterly directions from valleys tributary to the Till. The deposits are aligned approximately at right-angles to the south-easterly belt of eskers deposited earlier and clearly belong to the late phase of fluvioglacial drainage flowing down the Till valley. Indeed, eskers and kame terraces in the Chatton Basin and the Weetwood Gap trend down-valley. These phenomena represent the final evacuation of melt-water from the lower slopes of the north-east Cheviots and fringing basins. In the south-east it is presumed that the ice fragmented quite rapidly and that the broad stream valleys led meltwater drainage eastwards to the North Sea Basin.

The continued recession of the Tweed glacier should eventually have led to the isolation of a large mass of stagnant ice in the Milfield Basin north of Wooler. However, ice-contact landforms are totally absent from this broad basin, even although they abound in similar localities adjacent to the south. The anomaly is perhaps most satisfactorily accounted for in the following manner. Over much of the basin laminated clay and silt occur to the surface. Boreholes indicate that this deposit might be quite thick for between 40 and 70 feet of relatively stoneless clay has been proved. It has been suggested that the laminated sediments formed in a lake and additional evidence of the latter's existence is the enormous delta radiating out into the basin from an apex in the Glen valley to the west. Although the delta surface is entirely free from kettle holes, the series of eskers and kame terraces merging into the apex of the delta indicates that the latter was formed in a glacial environment. An extensive system of outwash streams emerging from a lobe of glacier ice in the Glen valley evidently constructed the delta. It is therefore possible that any ice-contact landforms previously occurring in the Milfield Basin were entirely over-run and obliterated by the delta and lake-floor deposits. Concerning the barrier to drainage which caused the formation of this extensive lake, the large system of enormous eskers and kame terraces extending across the Till valley at Etal suggests that the lake was impounded by the marginal zone of the Tweed glacier. From the distribution of laminated clays and the height of the delta edge, it may be estimated that the lake surface rose to approximately 145 feet. Since the lowest outlet from the Till valley, other than northwards to the Tweed, lies at 300 to 325 feet, any outflow from the lake must have gone into the Tweed glacier at Etal. Abandoned channels on the delta surface grading to a terrace at 125 feet north of Milfield indicate that a sudden drop in lake level occurred, the water flowing northwards and spreading gravel deposits round the end of the large esker belt. The lake appears to have finally drained when another fall in outlet level caused the further dissection of the terrace deposits, resulting in the channel now occupied by the river Till. Beginning at Etal, precisely where the gravel terrace terminates, an impressive rock-walled canyon meanders through drumlin topography to the Tweed. This gorge was probably cut subglacially by the large stream of water issuing from thelake when it stood at the 145-foot level; it has subsequently been occupied by the river Till.

Sissons (1964 and 1965) has suggested that the extensive belt of eskers and kame terraces between the rivers Tweed and Till, together with the

Milfield delta, represent the limit of the Aberdeen-Lammermuir Readvance in the Tweed valley. Evidence that might be put forward in support of that suggestion includes the following.

(1) The alignment of eskers and kame terraces between Cornhill and Etal is substantially different from that of similar landforms south of the Milfield Basin. Whereas the former are orientated chiefly from south-west to north-east and from west to east, the latter are aligned towards the south-east and south. Such evidence may well suggest different ice movements.

(2) Similarly, the orientation of drumlins in the lower Tweed valley is almost at right angles to that of striations and meltwater drainage features south of the Milfield area and would appear to be quite unrelated to the south-easterly movement of ice on the north-east flanks of the Cheviot massif indicated by the latter.

(3) The complete absence of kettle holes and fresh ice-contact landforms in the Milfield Basin could be taken as evidence that the delta and lake sediments have completely buried any such formations created during an earlier period of downwastage.

(4) The existence of the vast belt of sand and gravel deposits and its abrupt termination at Etal could suggest that the decaying marginal zone of a glacier lobe formerly occupied the territory between Cornhill and Etal while the adjacent Milfield Basin became filled with proglacial outwash deposits.

South of Edinburgh a drift 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 implies deglaciation followed by a readvance during which Southern Upland ice occupied a considerable area formerly occupied by Highland ice (Sissons 1965). The upper till has been interpreted as that deposited during the Aberdeen-Lammermuir Readvance. Drift sections are extremely poor in the Tweed-Till area, however, and information similar to that obtained from the drift south of Edinburgh cannot be put forward in support of the morphological evidence outlined earlier. For this reason it is perhaps necessary to examine the validity of that morphological evidence in more detail.

It has been suggested that the alignment of the Tweed drumlins and the fluvioglacial phenomena to the south indicates glacier flow towards the north-east and east and that this flow must be unrelated to the south-easterly flow indicated by meltwater channels, eskers and striations on the north-east slopes of the Cheviot massif, and adjacent areas. The implication is that the Tweed drumlins and the Tweed-Till belt of fluvioglacial deposits formed during a later readvance of the Tweed glacier that terminated near Milfield. The drumlins and adjacent fluvioglacial features are aligned parallel to glacially grooved and streamlined bedrock on the north-western and northern flanks of the Cheviot massif - features that suggest powerful ice movement and considerable erosion. The grooved and streamlined bedrock occurs either immediately adjacent to or at most one to two miles from the margins of the Aberdeen-Lammermuir Readvance limit suggested by Sissons. It seems unlikely, however, that such considerable ice erosion could have occurred only a short distance from the glacier snout (or margin) entirely during the suggested Readvance stage. It is more probable that the severe erosion was accomplished beneath a much thicker mass of ice. If this interpretation is correct then the basal layers of the Tweed glacier north of the Cheviot massif always flowed towards the north-east, closely controlled by topography, while the upper and marginal zones of the glacier moved north-eastwards, eastwards and south-eastwards as they rounded the periphery of the massif. Accordingly, the alignment of the

Tweed drumlins need not necessarily imply a readvance of ice even though that alignment differs from the movement of ice suggested by evidence south of Milfield. The drumlins possibly formed when downwastage had slowed down the forward movement of the Tweed glacier, and when the direction of flow in the basal layers closely corresponded with the alignment of the topography. At such a time fluvioglacial phenomena were probably forming in association with stagnating ice south of the Milfield Basin, where the slope of the ice surface was towards the south and south-east.

The absence of kettle holes and ice-contact deposits in the Milfield Basin is perhaps not surprising in view of the anomalous depression that floors the basin to an unknown depth below the present surface; a borehole has shown that it must lie near or below present sea level. The influence that this depression had on the events accompanying ice downwastage in the area remains conjectural, but the unusual depression was perhaps partly responsible for the formation of a lake in the Milfield Basin while areas adjacent to the north and west were still covered by stagnating glacier ice. Alternatively, the formation of the Milfield lake and outwash delta could be related to a period of stillstand during the downwastage of the last maximum ice sheet. Whether or not the ice readvanced from a position farther up-valley is of equal uncertainty.

Possible reasons for the existence of a vast esker-kame terrace belt between the Tweed and Till, and its abrupt termination at Etal, are discussed in the following chapter, and it is suggested that a readvance of ice need not necessarily be implied.

While the east Cheviot area should not be considered in isolation from other parts of Scotland where there is evidence that readvances possibly occurred towards the end of the last glaciation, the foregoing discussion gives some reason to believe that morphological evidence that might be taken in support of a readvance limit may be interpreted differently with equal validity. They could possibly result from a late stage in the period of glacier down-wastage during which the fluvioglacial features south of the Milfield Basin were formed.

The occurrence of morainic mounds in the Bizzle corrie, with an outwash apron extending down-valley from them, indicates that a very small glacier formed after the period of deglaciation discussed above. A broad ridge strewn with large boulders extends across the mouth of the corrie-like head of the College valley and an extensive outwash fan slopes steeply down-valley from it; it is suggested that the latter correlate in age with the Bizzle moraine. In view of the extensive plateau above 2,500 feet on The Cheviot, it is considered that a minor ice dome formed when the snowline had lowered during this later return to glacial conditions in parts of northern Britain. Valley heads on the western and northern flanks of The Cheviot, relatively sheltered in relation to maximal insolation, probably received considerable supplies of snow drifted off the summit snowfield. It is possible that much of the corrie development occurred at this time and under similar climatic conditions at earlier times. There is no direct evidence in the Cheviots with which to establish an absolute date for the corrie glacier stage and even its relative age is uncertain. The return to glacial conditions certainly followed the period of deglaciation responsible for fluvioglacial phenomena in the east Cheviot area because the snowline must have been very much higher during the latter period. Furthermore, distinct frost wedges occur in the Milfield delta and the extensive scree cones in some of the rock-cut meltwater channels probably date from the return to colder conditions. If the Aberdeen-Lammermuir Readvance limit suggested by Sissons for the Tweed valley is valid, then it seems logical to argue that the Cheviot corrie moraines represent the Perth Readvance because they are the next in sequence up-valley. However, since it has been suggested that the Aberdeen-Lammermuir Readvance limit is of uncertain validity, the relative age of the corrie moraines must also be doubtful. Indeed, the limits of the Perth Readvance in the eastern Borders have not yet been established satisfactorily. While Sissons (1965) placed the limit of the Perth Readvance on his map of "Stages of the last glaciation" in the upper Tweed valley, he did not discuss the evidence for this limit in the text. More than a century ago, A. Geikie (1863) and Young (1864) postulated two distinct phases of glaciation to account for the basal till and morainic mounds in the central Southern Uplands, conclusions recently confirmed by Price (1963). The latter mapped all the glacial phenomena of that area in detail and concluded that following a period of extensive ice sheet coverage, during which the till and a large number of meltwater channels formed, a readvance of valley glaciers descended to about 1,100 feet. Conspicuous hummocky moraines were formed by these glaciers. Admitting that the lack of datable deposits renders the absolute chronological relationships of the two glacial stages uncertain, Price tentatively suggested that the local valley glaciation occurred in the Zone 111 period (i.e. contemporaneous with the Loch Lomond Readvance of the Scottish Highlands). The limit of the Loch Lomond Readvance indicated by Sissons agrees with Price's suggestion, but the age of this limit has still not been established absolutely. If the moraines of the Tweedsmuir Hills do represent the Loch Lomond Readvance and if the meltwater channels and basal till mentioned by A. and J. Geikie, Young and Price resulted from an extensive ice sheet, then it is remarkable that the intervening Perth Readvance glaciers left no clear indication of their former presence in the same area. This is especially interesting in view of the well-marked limits of that glacial stage elsewhere in Scotland. Because of the uncertainty accompanying the correlation of late glacial limits in the

Tweed valley with those elsewhere in Scotland the age of the Cheviot corrie glaciation remains in doubt. However, it is perhaps significant that while valley glaciers in the Tweedsmuir Hills descended to about 1,100 feet, the small glaciers of the Cheviot corries appear to have extended down-valley no farther than 1,350 feet. Whether or not these valley moraines are correlative is unknown.

While it is generally recognised that the Pleistocene in north-west Europe has so far been characterised by several glacial periods separated by interglacial or interstadial periods, glacial landforms and the drift in the east Cheviot area may be satisfactorily interpreted with reference to only one ice sheet and subsequent valley glaciers. There are no discordant relationships within the great systems of meltwater channels, suggesting that some were formed at an earlier period. Similarly, the ice-contact deposits were clearly produced during one phase of glacier downwastage. It is perhaps possible that an earlier advance (or advances) of glacier ice over the east Cheviot area initiated some of the meltwater channels during its (their) retreat(s), and that during the last period of deglaciation the channels were simply re-occupied and perhaps further developed. No evidence has been found to confirm such a theory, however, or even to point in its direction. The drift sequence within and adajcent to the thesis area has also been interpreted as the deposit of one glaciation. It may therefore be concluded that the glacial sequence in the east Cheviot area can be reconstructed no further back in time than the last period when an ice sheet built up over the eastern Southern Uplands.

CHAPTER 9.

SOME THEORETICAL CONCLUSIONS

The detailed investigation of fluvioglacial phenomena in the east Cheviot area has enabled certain theoretical conclusions to be drawn concerning the origin and location of such phenomena. These conclusions may be valid in accounting for the form and distribution of fluvioglacial landforms in other areas formerly covered by an extensive ice sheet.

1. The Location of Meltwater Channels

One of the most striking contrasts in the morphology of the Cheviot massif is the unequal distribution of meltwater channels. Whereas hills below about 1,200 feet on the eastern fringe of the massif are furrowed by a great number of meltwater channels, such features are either few in number or totally absent from central parts of the massif and a vast area west of The Cheviot. It has also been observed in the field and from aerial photographs that a similar absence of meltwater channels characterises the entire range of hills extending south-west to the vicinity of Hawick. It is clear from the presence of till and glacially striated and grooved bedrock that these hillsides were formerly covered by glacier ice, so it may not be assumed that the paucity of meltwater channels indicates unglaciated areas. The unequal distribution of the channels must therefore be explained with reference to the meltwater drainage itself. In this connection, it is important to consider factors that influence the direction of meltwater flow. From maps and accounts of fluvioglacial phenomena elsewhere in northern Britain it is quite clear that the alignment of meltwater channels closely corresponds to the former direction of ice movement, as revealed by striations, grooved bedrock, crag and tail phenomena, drumlins and the transport of erratics. Since the slope of the ice surface also corresponds with the direction of ice movement (apart from perhaps local deviations due to irregularities in underlying topography) it seems reasonable to infer that the broad alignment of fluvioglacial drainage is probably controlled by that slope. It has already been indicated that the higher summits of the Cheviot Hills range and the hills south of Hawick supported ice domes. The movement of ice from these accumulation centres was down the pre-existing valleys radiating from the watershed (subsequently ice-shed). South of Hawick, ice movement was predominantly north-eastwards down the valleys of the Teviot, Allan Water, upper Rule Water and upper Jed Water; farther east, the movement varied between northwards and north-westwards until the ice became tributary to the larger stream moving north-eastwards down the Teviot/Tweed lowland. On the opposite side of the water/ice-shed, ice movement was predominantly south-eastwards in the upper valleys, but swung round into an easterly and north-easterly alignment on being influenced by ice movement from the Solway basin over the low Fell country north of the Tyne Gap. When the period of deglaciation began, meltwater drainage released by the ice domes would have followed the slope of the ice surface, which was roughly normal to underlying topography. With a rising snowline, the interfluves between the broad, but deeply incised, valleys would have become ice-free, and the meltwater drainage would have become constricted in the valleys themselves. For these reasons fluvioglacial drainage was able to flow relatively freely away from the area, either through or beneath ice occupying the valleys. In this way relatively few glacial meltwater streams became established on the hillsides because access to the valley bottoms appears always to have been available. Vast quantities of the till infill have been removed from the valleys, and although the present streams actively undercut cliffs of till in a few places, it seems probable that much of the infill was removed by meltwater drainage.

Round the eastern perimeter of the Cheviot massif, the former direction of ice movement was mostly at right-angles across the spurs and valleys. The alignment of meltwater drainage during the earlier periods of downwastage was similarly across the trend of the spurs and valleys. It was therefore inevitable that when englacial streams became superimposed to the ground beneath they cut channels across the spurs and in some valley heads. Evidence in support of this concept also occurs in the upper Tweed area, mapped and described by Price (1960, 1963). From his maps it is striking how the majority of meltwater channels occurring in the area are located on the north-western side of the Tweed valley. They are cut across spurs in a similar fashion to those in the east Cheviot area and their alignment coincides with the direction of former ice movement. On the south-eastern side of the Tweed valley, a large number of deeply incised tributary valleys enter from the south-east, but there is a marked absence of meltwater channels cut at right-angles across the intervening spurs. The southern tributary valleys lead north-west from the Tweedsmuir massif which formed an important centre of glacier dispersion during the last ice sheet glaciation. Consequently, the movement of ice was down these tributary valleys into the Tweed valley where the ice turned sharply north-eastwards. It is therefore to be expected that during the dissolution of the ice sheet meltwater drainage flowed down the tributary valleys, chiefly over or upon the valley bottoms, but on the north-eastern side of the Tweed valley, the meltwater streams were aligned across the spurs and cut prominent channels as they became superimposed.

In many areas formerly covered by an ice sheet the sides of valleys orientated in the direction of former ice movement are frequently furrowed by meltwater channels, and it may be thought that the conclusions pertaining to the Cheviot Hills are invalid elsewhere. It should therefore be emphasised that the Cheviot area is deeply dissected by steep-sided valleys and that extensive fragmentation of the glacier ice early in the deglaciation probably resulted from this underlying topography. In this event the meltwater streams were probably directed to positions over or upon the valley floors. Where broad valleys with relatively gentle slopes occur, there is much more scope for meltwater streams to become established in channels on the hillsides; there would be less tendency for lateral migration to locate the streams principally on the valley floors.

2. <u>The Influence of Topography on the Superimposition of Meltwater Streams</u>

Theoretically, superimposed englacial streams are liable to erode channels almost anywhere on a hillslope with little regard to the form of the ground. Price (1963) reported from the upper Tweed valley that, "The meltwater drainage system, at least at the higher levels, ignored the trends of the underlying relief." While Sissons (1963) observed that some subglacial drainage systems show a strong correlation with the relief of the areas in which they occur, he also noted that "In other instances, however, channels at least show little relation to the main relief features, for they cross spur crests indiscriminantly, ignoring cols and even being cut into the tops of hills." However, detailed mapping of all glacial meltwater features in the east Cheviot area and observations in north-east Scotland have led the writer to conclude that while small channels may sometimes show no relationship to topography, it is remarkable how the largest channels and complex channel systems are consistently located in pre-existing cols and valley heads. Accounts of some glacial meltwater channels in previous literature tend to support this conclusion.

Channels 2, 3, 5, 6, 10, 11, 12 (Map 5), are cut mainly in bedrock and vary in maximum depth from 50 to over 100 feet; all occupy relatively narrow, pre-existing cols and valley heads, and there is a marked absence of channels similar in size on the same spurs. Only one major channel appears to dominate a col if it is relatively narrow, and any adjacent signs of melt-water erosion are usually only minor benches and small abandoned branches occasionally seen higher up the col slopes or "hanging" on the sides of the main channel. The large col which cuts through the Tinto Hill ridge in Lanarkshire exhibits similar properties (Sissons 1961b). The writer has also observed many examples in the eastern Grampians and the Cairngorm Mountains where the dominant channels across spurs and even major watersheds invariably occupy narrow col floors.

Where broader cols are found one channel may still predominate, but branches and tributaries are more numerous and better developed; channels 1, 7, 8, 9, 15, 17 (Map 5), in the Cheviots, the Deuchrie Dod channels in East Lothian (Sissons 1958c), and a branch of the Carlops system south-west of Carlops village in Peeblesshire (Sissons 1963) may be quoted as examples.

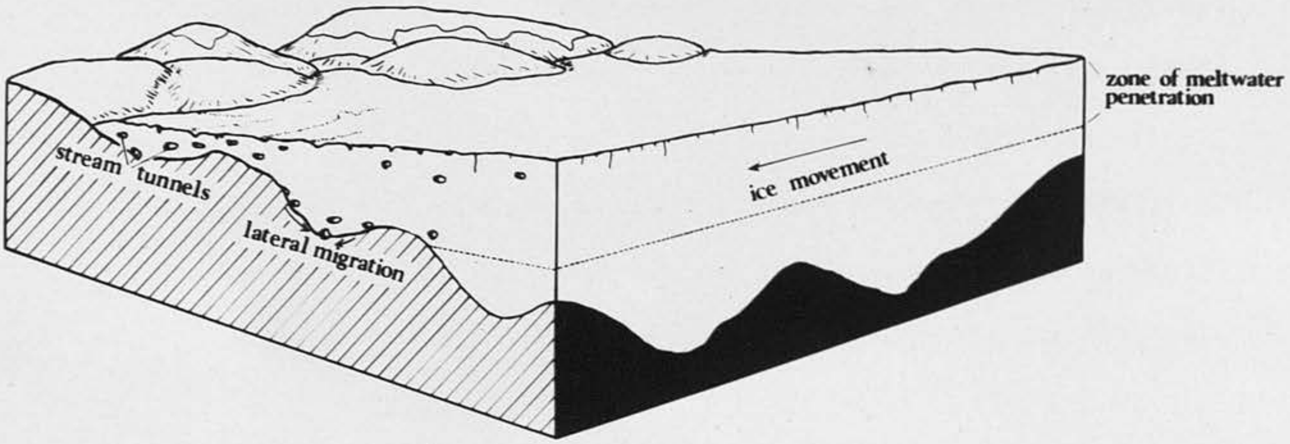
In the eastern Border hills, wide, shallow cols frequently lead into broad valley heads that probably experienced only minor modifications during the Pleistocene. Such a feature exists in the north-east Cheviots where the valley now occupied by the tiny Humbleton Burn heads in a broad embayment about 1,600 yards across; the width of this valley near its outlet is only about 550 yards. In this instance, a very intricate system of deep, anastomosing meltwater channels has been incised into the floor and gentle lower slopes of the valley head (Map 5). An even more composite channel system is described from an embayment on the east side of the Eddleston valley, Peeblesshire (Sissons 1961a, Figure 10, p. 25), and equally complicated channelling has been observed by the writer on aerial photographs covering two broad valleys opening northwards from the slopes of Morven, north of Ballater in Aberdeenshire.

The form and development of many mapped and observed meltwater channels and channel systems therefore suggest a strong correlation with the relief of the areas in which they occur. A subglacial origin has been indicated for those described in the Cheviots (Chapters 2 and 4), while the majority quoted from the literature have also been interpreted with the subglacial hypothesis. Since the channels are nearly all located in areas of broken and considerable relief, and in view of the discussion in Chapter 2, the superimposition of englacial streams most satisfactorily explains the means by which the eroding meltwaters became subglacial. It now remains to consider the mechanism of superimposition.

The magnitude and complexity of glacial meltwater channels and channel systems in many areas testify that vast volumes of meltwater were released by the dissipation of the various stages of the last glaciation in Britain. It is therefore reasonable to assume that dense networks of englacial (and probably supraglacial) streams existed in (and on) the downwasting ice at certain times. In the north-east Cheviots, the last major stream of ice to cover the peripheral hillsides swept eastwards and south-eastwards round the slopes of the massif as it spread down and out of the Tweed valley (Kendall and Muff 1902). A similar body of ice, penetrating contemporaneously from the west, swept over the south-east flanks of the Cheviots. Pre-existing valleys and spurs radiating from the east Cheviots were therefore orientated approximately at right-angles to the direction of ice movement and subsequently to that of meltwater flow upon deglaciation. If one imagines a system of englacial streams being superimposed on to such an area of short, deep valleys separated by spurs with well-developed cols in their crests, then the following events might be expected.

Stages in the Superimposition of Englacial Meltwater Streams

a. Early stage of downwastage



b. Late stage of downwastage

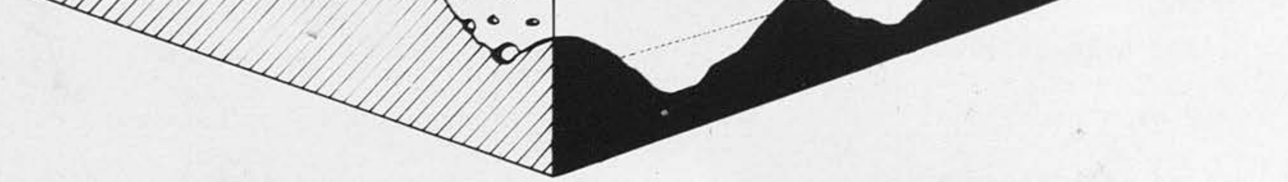


Figure 9.1

Streams which find themselves superimposed on to the steep slopes of narrow cols and spur ends may cut faint benches or notches in the bedrock, but are more likely to slip laterally downwards eroding mainly in the ice at the ice/ground contact rather than into the steep hillsides. In the meantime, adjacent streams located over the site of col floors will probably cut down more quickly through the ice as suggested by Price (1963) and demonstrated by Sissons (1963). If connections existed between such adjacent streams and those over the col floors (and anastomosing channel patterns in many areas certainly suggest the likelihood), then the lower streams may be expected to tap the waters of those at higher levels, thereby concentrating the flow of meltwater more and more over the sites of the col floors. Eventually, incision into a col floor may be made with the combined volume of a previously disseminated englacial channel system. Multiple intakes to channels 5, 6, 12 (Map 5), and abandoned sections at higher levels in channels 10, 15 (Map 5), demonstrate the possibility of this hypothesis in the Cheviots. (Figure 9.1).

Where wider cols with less steep slopes and broader floors lay beneath the ice it was possible for perhaps a few superimposed streams to erode more definite channels tributary to the main feature before they became abandoned on the col slopes, for example, channels 7, 8, 9, 15, 17 (Map 5).

The Trows system (Map 5) suggests that a broad, shallow, embayment type of valley head allowed latitude for the superimposition of englacial streams more as a complex system with many branches; several streams became deeply incised independently before uniting as the valley narrowed downstream. Even so, frequent abandoned loops and segments on the sides of and above the main branches testify to the continual concentration of water into the lowest route available in the valley floor.

In conclusion, it may be said that the foregoing observations suggest that in some upland areas of Britain (particularly the east Cheviot area) underlying topography considerably influenced the form and nature of glacial meltwater channels cut by meltwater streams superimposed from downwasting ice. Channels appear to be deepest in col floors because of the concentration of meltwater here by lateral migration downslope and capture of streams superimposed on the slopes of the cols. Wide valley heads allowed the establishment of complex channel systems, although abandoned loops and branches higher up the valley sides indicate that a similar concentration of meltwater was occurring in the valley floors.

An alternative possibility to the foregoing suggestion is that certain flow structures may have been induced within those parts of the glacier ice that became slightly concentrated as they flowed through the pre-existing cols, and such structures ultimately influenced the direction and location of meltwater drainage. Consequently, the majority of meltwater streams became directed into the cols as the ice began to downwaste, and the alignment of their abandoned channels reflects the direction of former ice movement.

It is also possible that the downwasting ice surface was itself considerably influenced by underlying relief, perhaps causing the concentration of supraglacial and englacial drainage in certain zones of the ice. Only observations from areas where ice masses are presently downwasting over irregular underlying topography can shed more light in this direction.

3. The Location of Meltwater Deposits

There is not only a marked contrast in the distribution of meltwater channels between the east Cheviot area and hills to the west, but also in the distribution of landforms composed of fluvioglacial sand and gravel. The western flanks of the Cheviot massif and the valleys leading north from the Cheviot Hills range are almost entirely free of ice contact phenomena such as eskers and kame terraces. Similarly, the south-eastern part of the massif is relatively devoid of these features. Great zones of eskers and kame terraces, liberally strewn with kettle holes and dead-ice hollows, occur in adjacent areas however, and it seems necessary to try and accountfor their presence in some localities and their virtual absence from others.

In the Teviot valley, ice-contact deposits first appear downstream from Hawick, but thereafter occur quite continuously as eskers and kame terraces on either side of the valley down to the Teviot/Tweed confluence. In the vicinity of Eckford, the eskers are orientated north-eastwards and can be readily traced into the valley of the Kale Water. The present river flows westwards, but it is evident that meltwater streams flowed eastwards, following the direction of former ice movement. A belt of ice contact deposits continues intermittently round the sinuous valley formerly occupied by the Kale and into that of the Bowmont Water; it terminates at the apex of the Milfield delta (Chapter 6). The deposits are located chiefly on the floor of the valley, but in places, kame terraces appear to cling precariously onto the precipitous sides. Beyond the Teviot/Tweed confluence, the low ground peripheral to the Cheviot Hills is thickly covered by the vast system of eskers and kame terraces extending eastwards to Etal in the Till valley. Partially submerging the southern margin of the Tweed drumlin field, these fluvioglacial deposits evidently post-date the drumlins, a few of which protrude above the vast sea of sand and gravel. The latter occurs as a triangular-shaped belt of deposits, extending roughly 4 miles from the Tweed,where it is 2 miles broad, to its abrupt terminus on the river Till, where it is less than a mile broad. South of Wooler and Chatton, the eastern foothills of the Cheviot massif and basins lying west of the Fell Sandstone escarpments are thickly covered by ice-contact fluvio-glacial deposits as far south as the river Breamish. While similar landforms are virtually absent in the territory south of the Breamish and Aln, striking eskers and kame terraces occur east of the Fell Sandstone dip-slope, for example, the Shipley system (Chapter 5), the Bradford Kaims (Dinham 1927) and those east of Alnwick (Parsons 1966).

It is therefore apparent that in the broad belt of country curving generally eastwards from lower Teviotdale to the east Cheviot area, ice-contact deposits occur only in certain places, and in order to account for this uneven distribution of the phenomena, the characteristics of the localities in which they occur must be considered. The following points emerge:

(a) The ice-contact deposits occur predominantly in low-lying areas, although eskers occur up to 600 feet on gentle hillsides in the north-east Cheviots.

(b) The most extensive systems of eskers and kame terraces occur particularly where valleys broaden into basins; for example, the west-east section of the lower Kale valley, the (suggested) former valley of the Till between Cornhill and Etal, and the Cementstone basins south of Wooler and Chatton.

(c) The deposits are located predominantly where considerable volumes of melt-water and water from ice-free areas were channelled into localities such as those mentioned above; for example,

i. The alignment of eskers in the vicinity of Eckford indicates that substantial volumes of meltwater flowing down Teviotdale turned eastwards into the Kale valley; the route presently followed by the Teviot between Eckford and Kelso was presumably established at a later period.

ii. Similarly, much of the meltwater drainage from the mid-Tweed area (including the Ettrick, Yarrow, Gala and Leader Waters) possibly entered the depression between Cornhill and Etal; the route to the sea presently occupied by the lower Tweed was probably formed much later.

iii. The Kale/Bowmont valley also received several streams coming from the Cheviot Hills to the south.

iv. In the north-east Cheviots, the great number of meltwater channels testifies to the abundance of meltwater drainage from the north-west and in addition, considerable volumes of water would have flowed down the valleys radiating from the heart of the massif.

(d) The ice-contact deposits are concentrated in areas where the topography is such that extensive masses of glacier ice would have become relatively detached from the main body of active ice early in the downwastage; during the ensuing period of prolonged stagnation, they would have provided ideal environments for the establishment of fluvioglacial deposition.

In accounting for the irregular distribution of ice-contact deposits, it is considered that the foregoing points all contribute significantly, but perhaps the most important factor involved in this discussion concerns the actual origin of the sands and gravels. Eskers and kame terraces lead away from the outlets of large rock-cut meltwater channels in some parts of the Cheviots and it seems likely that much of the sand and gravel was derived from the channels. However, the majority of ice-contact deposits are located in areas somewhat remote from meltwater channels, and in any case, the total volume of these deposits is greatly in excess of the combined volume of the nearest systems of meltwater channels. Since there seems to be some relationship between the occurrence of areas in which ice-contact deposits are concentrated and the pre-existing river valleys, it is reasonable to suggest that much of the debris composing the eskers and kame terraces was derived from these valleys. It has already been suggested that substantial quantities of the till infill within the valleys was probably removed by meltwater drainage flowing down the valleys. It is therefore concluded that a proportion of the fluvioglacial deposits was derived from such sources. In this connection, the fact that shallower layers of till appear to have been deposited in valleys on the south and south-east sides of the Cheviot area than in those to the north may be of great consequence. Since smaller amounts of debris were available for meltwater streams flowing down the southern valleys, it is possible that the relative paucity of ice-contact deposits in these areas is partly explained in this way.

Accounts of eskers by Chamberlin (1895) and Upham (1894), led Charlesworth (1957) to conclude that like drumlins, eskers form in the lower layers of ice and gather their material "from englacial sources". Flint (1957) observed that the sediments of which an esker is built are closely similar to the constituents of the till in the vicinity, and stated, "The similarity suggests that they have a common origin in the drift carried in the ice." Although Tarr and Martin (1914) described 5 to 10 feet of ablation moraine on the surfaces of many Alaskan glaciers, they rarely mentioned material incorporated within the ice. However, discussing the sediment load of subglacial streams, they inferred that "much the greater proportion is derived from the lower layers of the glacier". While detailed discussions concerning the relationships between eskers and englacial detritus appear not to have been written, it seems to be generally agreed that in modern glaciers, eskers derive most of their material from englacial moraine. Such material appears to be very common in all modern glacier-regions, although it may be absent in places and masked by supraglacial debris in others (Charlesworth 1957). The amount and distribution of morainic debris once contained by glacier ice in areas formerly glaciated is often difficult to assess, because of the absence of bore-hole data and sections. Although these difficulties are encountered in the eastern Borders, certain generalisations can be made concerning the former drift content of the ice that covered that territory.

(1) In the lower Tweed valley, the remarkable field of closely spaced drumlins indicates that the Tweed glacier contained immense quantities of drift and considerable deposition was enforced. Several of the more prominent drumlins rise over 100 feet above adjacent depressions and Gunn (1895) recorded a boring which was made to a depth of 102 feet in till in a hollow between two drumlins. It is therefore possible that more than 200 feet of till occurs in places in the Tweed drumlin field. Much of this debris was undoubtedly derived from severe glacial erosion in higher areas to the south-west. For example, the softer sedimentary strata of the highly inclined Silurian beds have been gouged into a remarkable landscape of parallel ridges and furrows in the country between the Ettrick and Rule valleys. The Upper Old Red sediments between Denholm and Roxburgh (north-east of Hawick) have been deeply moulded into massive ridges and depressions principally by ice action, and the multitude of volcanic outcrops protrude as prominent stosses in crag and tail formations. These streamlined landforms gradually merge into the Tweed drumlin field in the vicinity of Kelso. Since the drumlin field also merges laterally into streamlined rock ridges and/or crag and tail phenomena, it is probable that the bedrock buried beneath the drumlins may have been grooved in a similar fashion before the subsequent accretion of till round these nuclei led to the formation of drumlins. This phase of drumlin deposition presumably resulted partly from lower layers of the Tweed glacier becoming heavily charged with debris eroded from the soft sedimentary rocks of the locality. It may also have coincided with an early stage of downwastage, when the ice was moving with decreased velocity and volume. Drumlins in the Tweed valley appear to be composed chiefly of lodgement till, built up by accretion in the lower layers of glacier ice

that was flowing quite rapidly, for many are extremely elongated forms (particularly in the centre of the valley where movement may have been most rapid). So long as powerful forward movement was able to mould the vast thicknesses of till into streamlined forms, most of the till was probably carried along chiefly in horizontal stream-lines in the ice before being deposited as drumlins or as till sheets plugging valleys and other depressions in the topography. When the powerful forward motion ceased, presumably with the onset of downwastage, the immense quantities of detritus caught up in the lower layers of the ice would have caused considerable friction as the ice continued to move weakly in a forward direction prior to complete stagnation. Under such conditions it seems likely that extensive shear-planes developed, curving upwards from the base of the ice, particularly in areas where drumlins or transverse ridges (such as the Fell Sandstone escarpments) acted as prominent obstacles to ice movement. Debris lodged in the basal layers of the ice would have been carried upwards along such shear-planes - as observed in some modern glaciers (Carruthers 1939, Charlesworth 1957) - and would have formed a more or less continuous zone of dirt-laden ice extending to the surface of the glacier. Depressions aligned somewhat transversely or obliquely to the principal direction of ice movement, such as the lower Kale valley, the Cornhill-Etal area and the Cementstone basins east of the Cheviot massif, would have been sites particularly favourable for impeding the efficient flow of ice and thereby became localities in which shearing and the upward movement of till were characteristic. With the progress of downwastage, immense volumes of water began to thread through the stagnating ice, frequently as complex systems of interconnecting tunnels, and concentrated particularly in the lowland areas just referred to. Whether or not the stream courses were subaerial, englacial or subglacial, it is clear that substantial quantities of debris must have been washed from the dirt-laden ice by these networks of meltwater rivers. The selective action of running water subsequently arranged the detritus into beds of material varying in calibre from fine sand to large boulders, and the majority of stones became rounded or sub-rounded. It is therefore suggested that many of the eskers and kame terraces in the belt of country curving round the Cheviot Hills from lower Teviotdale to the Breamish are composed of deposits derived in the above fashion from glacier ice that was heavily charged with englacial moraine.

As the ice slowly disappeared, considerable amounts of supraglacial and englacial debris were gradually superimposed onto the ground beneath, onto ice-contact deposits and onto areas free of such phenomena. Thin layers of such ablation till overlie sands and gravels south of Wooler, but up to 7 feet occurs in the Shipley area. A threefold division of the glacial drift of eastern Northumberland and Durham has long been recognised, and it is possible that the upper layer consists of ablation till. For example, the lower till is blue-grey in colour, extremely tenacious and normally overlies broken rock-head; the upper till, somewhat reddish-brown in colour and containing a much larger proportion of far-travelled stones (Carruthers 1930) is more loosely compacted. In some instances the upper till rests directly on top of the lower, with a sharp line indicating the division, but in many places, varying thicknesses of bedded sand, gravel and laminated clay separate the two tills. The upper surface of these "middle sands and gravels" shows no evidence of having been overridden by moving glacier ice. For this reason and because of its loose compaction, Carruthers (1927, 1930, 1932, 1939) interpreted the upper till as ablation moraine. He concluded, "One may suggest that this upper clay represents englacial detritus, a heterogeneous collection of dirt and stones, often of distant origin. All this material, once scattered through an ice-sheet several hundred feet thick would be slowly melted out and left behind as the ice dwindled away at the close of glaciation." The present writer firmly agrees with Carruthers that the drift phenomena of the lower Tweed valley and north Northumberland are the product of one ice sheet, and that the upper till represents englacial detritus which slowly settled in situ as ablation moraine. Indeed, in the absence of such material it would be extremely difficult to account for the immense volume of fluvioglacial deposits that occur in certain parts of the area.

(2) The territory affected by the western ice mass from the Solway basin lies south of the Cheviot Hills range and includes the south and south-east flanks of the Cheviot massif. The relative absence of ice-contact deposits in this area (Chapter 5) appears to be related to two principal factors:

(a) Firstly, there are no deposits of till comparable in depth and extent to those in the north, suggesting that much less englacial debris would have been available for reworking by meltwater streams. Since only one till has been observed in the south Cheviot area, it seems that englacial moraine was either completely lacking or else too sparse to be significant in the drift sequence. In the absence of such material, meltwater streams would not have contained much load and so the formation of ice-contact deposits would have been severely restricted.

(b) Secondly, whereas admirable sites for the stagnation of large masses of glacier ice were present in the northern area, similar environments were lacking in the south, where the ground is considerably higher. The upper-mid Aln valley is certainly at a similar elevation to the Cementstone basins farther north, but while the latter are aligned rather obliquely or transverse to former glacier flow, the orientation of the Aln valley coincides with the former movement of the western ice mass and would have permitted uninterrupted flow. Consequently, there was probably less shearing to carry detritus into englacial positions, and during downwastage, the relatively clean ice would almost certainly have disappeared quickly.

For the above reasons the environments and conditions conducive to the deposition of ice-contact landforms were either entirely absent or severely restricted in the south Cheviot area.

4. The Problem of Multiple Glaciation

Throughout this thesis it has been assumed that all the fluvioglacial phenomena are related to the last period of maximum glaciation in Britain. While it is reasonably certain that the ice-contact deposits and meltwater channels cut in drift were formed as the last ice sheet in the east Cheviot area downwasted, it is doubtful if the large-rock-cut meltwater channels formed entirely during the same period. If such conspicuous landforms were created as a result of the last glaciation, then it seems reasonable to expect that similar features formed during a previous period of widespread glaciation. If the ice sheets of separate glaciations had differed radically in their extent and directions of movement, then it is possible that rock-cut meltwater channel systems related to different glaciations would have been aligned discordantly to one another. Furthermore, some evidence of drift-filled channels or degraded channels might also be expected. Yet there is no such evidence in the Cheviot Hills. In view of this, two alternative explanations may be suggested; either (a) the Cheviot Hills were ice-covered during only one period of glaciation, or (b) the Cheviot Hills were covered by similar ice masses during separate glaciations and that phenomena such as deep rock-cut meltwater channels are the result of more than one period of glacier downwastage. If the second alternative is valid, then all of the meltwater channels in the Cheviot massif were certainly utilised during the last time that glacier ice covered the entire massif.

Additional Drift Information

Chapter 1.

(a) [illegible] Drift Sections.

Site 1. [illegible]

[illegible]

The gravel layer between [illegible] and contains many [illegible]. The upper till layer [illegible] eroded by the stream that deposited the [illegible] gravels. The fabric analysis was made [illegible].

Site 2. Located immediately above Site 1, [illegible] inches above the layer of laminated clay and sand and [illegible] inches below the top of the upper till bed.

The upper till is fairly compact, but not nearly as much as the lower till. This seems to be so because the matrix of the former is chiefly silt, whereas the latter is a tough clayey till.

Site 3. Located on the right bank of the stream approximately [illegible] yards up-stream from the two previous sites.

APPENDIX

Additional Drift Information

Chapter 1.

(a) Shipley Burn Drift Sections:

Site 1. Grid Ref. 4155/6192. Left bank of the stream.

4 feet ..... soil, coarse water-worn gravel and cobbles.

3 to 4 feet ..... reddish-brown till.

3 to 4 inches ..... laminated clay, fine sand.

2 to 12 inches ..... water-worn gravel and cobbles.

6 feet ..... grey-brown till to water level.

The gravel layer between the two tills thickens in a downstream direction and contains many blocks over 1 foot in size and some up to 3 feet in diameter. The upper till layer thins in places as though its upper surface had been eroded by the stream that deposited the overlying gravels.

The fabric analysis was made 4 feet above river level.

Site 2. Located immediately above Site 1, 14 inches above the layer of laminated clay and sand and 26 inches below the top of the upper till bed.

The upper till is fairly compact, but not nearly as much as the lower till. This seems to be so because the matrix of the former is chiefly silt, whereas the latter is a tough clayey till.

Site 3. Located on the right bank of the stream approximately 40 yards upstream from the two previous sites.

15 feet ..... bedded gravels, laminated clay and silt.

18 feet ..... grey-brown till to water level.

The fabric analysis was made 15 feet above river level.

Site 4. Located approximately 100 yards upstream from Site 3, on the left bank of the stream.

2 to 4 feet ..... bedded sand, gravel and cobbles.

7 to 9 feet ..... grey-brown till to river level.

The till is clayey and tenacious and contains a high number of stones.

The fabric analysis was made 3 feet 4 inches above river level.

Site 5. Grid Ref. 4152/6180. Right bank of the stream.

$6\frac{1}{2}$ feet ..... reddish-brown till.

15 to 17 feet ..... bedded sands and gravels.

15 to 20 feet ..... grey-brown till to river level.

The upper till is very silty in places but also contains blocks up to 2 feet in size. Sub-rounded, sub-angular and shattered stones all occur in the upper till.

The fabric analysis was made in the upper till.

Site 6. Grid Ref. 4152/6176. Right bank of the stream.

7 feet ..... reddish-brown till.

c.45 feet ..... bedded sands and gravels to river level.

The till layer is very silty and contains many large blocks up to 2 feet in size. Some of the blocks appear to be in a shattered state.

(b) Eglingham Burn site. Grid Ref. 4116/6108. Right bank of the stream.

2 feet 4 inches ..... soil and weathered alluvium.

1 foot 4 inches ..... water-worn gravel, $\frac{1}{4}$ to 12 inches in size.

6 to 12 inches ..... grey stoneless clay.

3 feet 6 inches ..... purply-brown till to river level.

The till in this section is not only slightly different in colour from that of the previous sites, but also it contains fewer stones; the stones are generally more angular than those in the tills, at the other sites.

(c) Thrunton Brick Pit. Grid Ref. 4092/6097.

The till is variable in colour, stone content and matrix. In places it is very clayey and relatively free from stones; in others, it is silty and frequently contains pockets of sand and stoneless clay. The fabric analyses were made at three different sites in the pit; one in an 8-foot vertical section at the top of the pit, and the others in holes dug into the sloping bank of till scraped by the machines of the brick works.

(d) Netherton Burn site. Grid Ref. 3998/6072. Right bank of the stream.

1 foot 4 inches ..... sandy soil.

8 feet 7 inches ..... grey-brown till to river level.

The till is extremely stoney, the stones being contained in a silty-gritty matrix.

<u>Chapter 3</u>.

(a) Exposures of lake floor deposits in the Hedgeley basin.

i. Grid Ref. 4075/6186. Right bank of the river Till.

The river has undercut the bank and caused considerable slumping, but up to $5\frac{1}{2}$ feet of laminated clay and silt was exposed at this site in 1963. The sediments are stoneless and are red, brown and grey in colour.

ii. Grid Ref. 4076/6193. Right bank of the river Till.

1 foot 6 inches ..... soil and weathered material.

6 inches ..... weathered gravel.

2 feet 2 inches ..... brown mottled/grey clay.

1 foot 4 inches ..... sand, grit and small gravel.

7 inches ..... well laminated clay; grey and light-brown.

6 inches ..... grit and small gravel; one slab of sandstone, 1 foot x 9 inches x 1 foot.

4 feet ..... purply-brown till to river level.

iii. Grid Ref. 4077/6194. Right bank of the river Till.

8 to 10 feet ..... laminated silt and clay.

9 to 10 feet ..... purply-brown till to river level.

iv. Grid Ref. 4079/6196. Right bank of the river Till.

9 to 10 feet ..... laminated silt and clay (upper foot weathered).

20 to 25 feet ..... purply-brown till to river level.

The till exposed at the last three sections is extremely tenacious and clayey. It contains a relatively sparse scattering of stones, but large blocks over 12 inches in size occur.

(b) Auger results from the Hedgeley basin.

i. Grid Ref. 4075/6203. Floodplain of the river Till.

2 feet ..... sand.

1 foot ..... gravel.

ii. Grid Ref. 4072/6203. Sloping surface approximately 15 to 20 feet above the river floodplain.

6 inches ..... soil.

3 inches ..... sandy soil.

3 inches ..... clayey soil.

2 feet ..... brown clay.

iii. Grid Ref. 4071/6203. Morphology as above.

6 inches ..... soil.

2 feet 6 inches ..... yellow-brown silt and fine sand.

iv. Grid Ref. 4069/6203. Morphology as above.

7 inches ..... clayey soil.

2 feet 5 inches ..... laminated clay.

v. Grid Ref. 4071/6199. Floodplain of the river Till.

3 feet ..... gravel.

vi. Grid Ref. 4068/6200. Sloping surface above floodplain.

3 feet ..... clay.

vii. Grid Ref. 4067/6197. Frontal slope of delta.

6 inches ..... clayey soil.

4 inches ..... silt and clay.

1 foot 2 inches ..... laminated clay.

viii. Grid Ref. 4065/6196. Higher up frontal slope of delta.

7 inches ..... sandy soil.

1 foot 3 inches ..... fine silt.

5 inches ..... silt and clay.

9 inches ..... laminated clay.

ix. Grid Ref. 4064/6197. High up frontal slope of delta.

1 foot 9 inches ..... sand.

1 foot ..... silt.

3 inches ..... silt and clay.

x. Grid Ref. 4063/6196. Same level as ix.

8 inches ..... soil and fine gravel.

6 inches ..... sand.

1 foot 10 inches ..... fine micaceous silt.

xi. Grid Ref. 4062/6197. Floor of kettle hole.

1 foot 10 inches ..... yellow-brown silt.

1 foot 2 inches ..... light grey silt.

xii. Grid Ref. 4062/6198. Near the surface of the delta.

1 foot ..... soil, sand, gravel.

10 inches ..... sand.

1 foot 2 inches ..... silt.

xiii. Grid Ref. 4061/6198. Edge of the delta surface.

3 feet ..... sand (ground surface littered with fine gravel).

Chapter 6.

(a) Exposures of laminated clay in the Chatton Basin.

1. Along the Hollow Burn.

Grid Ref. 4059/6273 ..... few inches of grey and red laminated clay at stream level.

2. Along the Mill Burn.

Grid Ref. 4064/6289 ..... small exposure of grey and red laminated clay near stream level.

3. Left bank of the river Till, north-north-east of Newton.

Grid Ref. 4044/6257	9 inches	soil.
	2 feet 6 inches	red-brown clay with stones = solifluction debris.
	1 foot	gravel (fine).
	2 to 3 inches	sand.
	1 foot	gravel.
	3 feet	grey and red laminated clay to river level.
Grid Ref. 4045/6261	3 feet	weathered silt and laminated clay.
	2 feet	fine gravel.
	3 to 4 feet	red laminated clay then obscured to river level.

4. Holes dug for drainage tiles in a field south of the Till, c.500 yards west of Chatton.

Grid Ref. 4049/6284 ..... 4 feet 6 inches to unknown depth ..... grey and red laminated clay.

5. Left bank of small tributary stream on right bank of the Till, c.1,500 yards north-north-east of Chatton.

Grid Ref. 4063/6298 ..... 1 foot 8 inches to unknown depth below stream level ..... grey and red laminated clay.

(b) Exposures of laminated clay in the Milfield Basin.

1. Old clay pit c.600 yards south-east of Nesbit.

Grid Ref. 3988/6333 ..... small exposures of grey and red laminated clay.

2. Left bank of the river Till at First Linthaugh (1L. Map 2).

Grid Ref. 3934/6369 ..... 2 to 3 feet ..... fine gravel.
6 inches ..... red-brown laminated clay.
$2\frac{1}{2}$ inches ..... grey-green silt.
1 inch ..... red laminated clay.
$2\frac{1}{2}$ inches ..... grey-green silt.
$\frac{1}{4}$ inch ..... red laminated clay.

etc. to total depth below the overlying gravel of 6 feet. The remaining 2 feet above river level are obscured by slumping.

3. Left bank of the river Till at Fourth Linthaugh (4L. Map 2).

Grid Ref. 3932/6363 ..... 4 feet ..... fine gravel.
2 feet ..... red laminated clay to unknown depth.

Grid Ref. 3932/6362 ..... 2 feet 6 inches ..... well bedded fine gravel.
8 feet 6 inches ..... grey-green and red laminated clay.
to unknown depth

4. Left bank of the river Till opposite Flodden Tile Works.

Grid Ref. 3933/6358 ..... 2 feet 4 inches ..... fine gravel.
2 feet 8 inches ..... red laminated clay.

BIBLIOGRAPHY

Anderson, W. 1927 See Carruthers, R. G., et al.

1930 See Carruthers, R. G., et al.

1932 See Carruthers, R. G., et al.

Bowen, D. Q., and Gregory, K. J. 1965 A glacial drainage system near Fishguard, Pembrokeshire. Proc. Geol. Assoc. 76, 275-281.

Buckland, D. D. 1839 A Memoir on the Evidence of Glaciers in Scotland and the North of England. Proc. Geol. Soc. IV, p. 29.

Bullerwell, R. G. A. 1910 On the superficial deposits at the foot of the Cheviot Hills between Wooler and Glanton. Geol. Mag. 7, 452-458.

Bulman, G. W. 1891 On the Origin of the Upper Drift Sands and Gravels of Northumberland. The Naturalist. 43-45.

Burnett, G. A. 1927 See Carruthers, R. G., et al.

1930 See Carruthers, R. G., et al.

1932 See Carruthers, R. G., et al.

Butler, G. G. 1907 Anniversary Address. Proc. Berwickshire Nat. Club. 19, 97-107.

Carey, S. W., and Ahmad, N. 1961 Glacial Marine Sedimentation. Arctic Geology. vol. 2, 865-894.

Carruthers, R. G., et al. 1927 The Geology of Belford, Holy Island and the Farne Islands. Mem. Geol. Surv. Eng. and Wales.

1930 The Geology of the Alnwick District. Mem. Geol. Surv. Eng. and Wales.

1932 The Geology of the Cheviot Hills. Mem. Geol. Surv. Eng. and Wales.

Carruthers, R. G. 1931 The Cheviot Hills (in) Contributions to the Geology of Northumberland and Durham. Proc. Geol. Assoc. vol. XLII, Pt. 3, 228-232.

1939 On Northern Glacial Drifts: Some Peculiarities and their Significance. Quart. J. Geol. Soc. Lond. 95, 299-328.

Chamberlain, T. C. 1895 Recent Glacial Studies in Greenland. Bull. Geol. Soc. Amer. 5, p. 71.

Charlesworth, J. K. 1957 The Quaternary Era. vol. 1. Arnold, London.

Clapperton, C. M. 1960 Glacial and post-glacial geomorphology in an area south of Hawick. M.A. dissertation, Edinburgh University.

1963 The Geological History of the Cheviot Massif with Particular Reference to its Glaciation. Trans. Hawick Archeol. Soc. 18-27.

1966 The influence of topography on the superimposition of glacial meltwater streams. In: Deglaciation. ed. R. J. Price, (Brit. Geomorph. Research Grp.), Occasional Paper 3, 13-18.

Clough, C. T. 1887 See Miller, H., and Clough, C. T.

1888 The Geology of the Cheviot Hills (English Side). Mem. Geol. Surv. Eng. and Wales.

1889 The Geology of Plashetts and Kielder. Mem. Geol. Surv. Eng. and Wales.

Clough, C. T., and Gunn, W. 1895 The Geology of part of Northumberland, including the country between Wooler and Coldstream. Mem. Geol. Surv. Eng. and Wales.

Common, R. 1953 The Geomorphology of the East Cheviot Area. Ph.D. thesis, Edinburgh University.

1954 The Geomorphology of the East Cheviot Area. Scot. Geogr. Mag. 70, 124-138.

1957 Variations in the Cheviot Meltwater Channels. Geogr. Studies. 4, 90-103.

Curry, J. 1867 On the Drift of the North of England (Abstract). Quart. J. Geol. Soc. Lond. 23, p. 40.

Derbyshire, E. 1961 Subglacial Col Gullies and the Deglaciation of the North-East Cheviots. Trans. Inst. Brit. Geogr. 29, 31-45.

Dinham, C. H. 1927 See Carruthers, R. G., et al.

1930 See Carruthers, R. G., et al.

Dwerryhouse, A. R. 1902 The Glaciation of Teesdale, Weardale and the Tyne Valley and their Tributary Valleys. Quart. J. Geol. Soc. Lond. 58, 572-607.

Eastwood, T. 1953 British Regional Geology, Northern England. (Third Edition), 63-66.

Embleton, C. 1961 The geomorphology of the Vale of Conway, North Wales, with particular reference to its deglaciation. Trans. Inst. Brit. Geogr. 29, 47-70.

Fitzpatrick, E. A. 1963 Deeply weathered rock in Scotland, its occurrence, age and contribution to the soils. J. Soil Sci. 14, 33-42.

Flint R. F. 1957 Glacial and Pleistocene Geology. Wiley & Sons, New York. p. 156.

Garwood, E. J. 1899 Additional notes on the glacial phenomena of Spitsbergen. Quart. J. Geol. Soc. Lond. 55, p. 691.

1910 The Geology of Northumberland and Durham. Geol. Assoc. Jubilee vol. 661-697.

1922 Geology. (in) A History of Northumberland. vol. XI, 2-11. Newcastle-upon-Tyne.

Geikie, A. 1863 On the glacial drift of Scotland. Trans. Geol. Soc. Glasgow. 1, 185 pp.

Geikie, J. 1876 The Cheviot Hills. Good Words. XVII, 11-15, 81-86, 264-270, 331-337, 550-556.

1894 The Great Ice Age. (Third Edition) London.

Gjessing, J. 1960 Isavs meltningstidens drenering dens forløp og formdannende virkning i nordre Atnedalen. Ad. Novas 3 Oslo.

1965 Deglaciation of southeast and east-central south Norway. Norsk Geografisk Tids. 20, 133-147.

Glen, J. 1954 The stability of ice-dammed lakes and other water-filled holes in glaciers. J. Glaciol. 15, 316-318.

Godard, A. 1965 Recherches de geomorphologie en Ecosse du nord-ouest. Paris.

Gunn, W. 1895 See Clough, C. T., and Gunn, W.

1897 The Geology of the Country between Norham and Tweedmouth. Mem. Geol. Surv. Eng. and Wales.

1897 The Geology of the Coast South of Berwick-on-Tweed. Mem. Geol. Surv. Eng. and Wales.

Gunn, W. (cont.) 1900 The Geology of Belford, Holy Island and the Farne Islands. Mem. Geol. Surv. Eng. and Wales.

1927 See Carruthers, R. G., et al.

Hardy, J. 1872 Langleyford Vale and the Cheviots. Proc. Berwickshire Nat. Club.

Hickling, A. 1931 Structure and Topography, with a General History of the Carboniferous Deposits. (in) Contributions to the Geology of Northumberland and Durham. Proc. Geol. Assoc. vol. XLII, Pt. 3, 219-228.

Howse, R. 1862 On the Glaciation of the Counties of Durham and Northumberland. North of England Inst. Min. Eng. vol. XIII.

Kendall, P. F., and Muff, H. B. 1901 Evidences of Ancient Glacier-dammed Lakes in the Cheviots. Geol. Mag. N.S. VIII, 513-515.

1902 The Evidence for Glacier-dammed Lakes in the Cheviot Hills. Trans. Edinb. Geol. Soc. 8, 226-230.

Kirby, R. P. 1961 Movements of ice in central Labrador - Ungava. Cahiers de Géographie de Quebec. 10, 141-217.

Lebour, G. A. 1878 Outlines of the Geology of Northumberland. Newcastle.

1891 On Certain Surface Features of the Glacial Deposits of the Tyne Valley. Trans. Nat. Hist. Soc. of Northumb., Durham and Newcastle.

Linton, D. L. 1949 Unglaciated Areas in Scandinavia and Great Britain. Irish Geog. 2, 30-31.

1955 The Problem of Tors. Geogr. J. 121, 470-487.

1959 Morphological Contrasts of Eastern and Western Scotland. Geographical Essays in Memory of Alan G. Ogilvie. Nelson, Edinburgh. 16-45.

Manley, G. 1951 The range of variation of the British climate. Geogr. J. 117, 43-65.

1955 On the occurrence of ice domes and permanently snow-covered summits. J. Glaciol. 2, p. 453.

1959 The late-glacial climate of north-west England. The Liverpool and Manchester Geol. J. 2, 188-213.

Mannerfelt, C. M. 1945 Några glacial morfologiska formelement. Geogr. Ann., Stockh. 27, 1-239.

Merrick, E. 1909 On the Superficial Deposits around Newcastle-on-Tyne. Proc. Univ. Durham Phil. Soc. 3, 141-153.

Miller, H., and Clough, C. T. 1887 The Geology of the Country around Otterburn and Elsdon. Mem. Geol. Surv. Eng. and Wales.

Milne-Home, D. 1876 Notices of high water marks on the banks of the river Tweed, and some of its tributaries. Trans. Roy. Soc. Edinb. 27, p. 513.

Parsons, A. 1966 Some aspects of the glacial geomorphology of north-eastern Northumberland. M.Sc. Thesis, Leicester University.

Peel, R. F. 1951 A study of two Northumbrian spillways. Trans. Inst. Brit. Geogr. for 1949. 15, 73-89.

Price, R. J. 1960 Glacial meltwater channels in the upper Tweed drainage basin. Geogr. J. 126, 483-489.

1963 The glaciation of a part of Peeblesshire, Scotland. Trans. Edinb. Geol. Soc. 19, 325-347.

1964 Landforms produced by the wastage of the Casement Glacier, Southeast Alaska. Inst. Polar Stud. (Ohio State Univ.) Rep. No. 9.

1966 Eskers near the Casement Glacier, Alaska. Geogr. Ann. Stockh. 48, Ser. A, 111-125.

Raistrick, A. 1931 Glaciation. (in) Contributions to the Geology of Northumberland and Durham. Proc. Geol. Assoc. 42, 281-291.

Sissons, J. B. 1958a Sub-glacial Stream Erosion in Southern Northumberland. Scot. Geogr. Mag. 74, 163-174.

b Supposed Ice-dammed Lakes in Britain with Particular Reference to the Eddleston Valley, Southern Scotland. Geogr. Ann., Stockh. 40, 159-187.

c The deglaciation of part of East Lothian. Trans. Inst. Brit. Geogr. 25, 59-76.

1960 Some Aspects of Glacial Drainage Channels in Britain. Part I. Scot. Geogr. Mag. 76, 131-146.

1961a Part II. Scot. Geogr. Mag. 77, 15-36.

b A subglacial drainage system by the Tinto Hills, Lanarkshire. Trans. Edinb. Geol. Soc. 18, 175-192.

Sissons, J. B. (cont.) 1963 The glacial drainage system around Carlops, Peeblesshire. Trans. Inst. Brit. Geogr. 32, 95-111.

1964 The Glacial Period. The British Isles, A Systematic Geography. Nelson, Edinburgh. 131-149.

1965 Quaternary. The Geology of Scotland. Oliver & Boyd, Edinburgh. 467-496.

Smythe, J. A. 1908 The Glacial Phenomena of the Country between the Tyne and the Wansbeck. Trans. Nat. Hist. Soc. of Northumb. and Durham, N.S. 3, 79-109.

1912 The Glacial Geology of Northumberland. Trans. Nat. Hist. Soc. of Northumb. and Durham, N.S. 4, 86-116.

Sugden, D. 1965 Aspects of the glaciation of the Cairngorm mountains. D.Phil. thesis, Oxon.

Tarr, R. S., and Martin, L. 1914 Alaskan Glacier Studies. The National Geographic Society, Washington.

Tate, G. 1849 The Polished and Scratched Rocks in the Neighbourhood of Alnwick. Trans. Tyneside Nat. Field Club. 1, p. 38.

1866a Records of Glaciated Rocks in the E. Borders. Proc. Berwickshire Nat. Club. 5, 236-240.

b The Cheviots. Proc. Berwickshire Nat. Club. 5, 236-240.

Trechmann, C. T. 1915 Scandinavian Drift on the Durham Coast. Quart. J. Geol. Soc. Lond. 71, 53-82.

1920 On a Deposit of Interglacial Loess. Quart. J. Geol. Soc. Lond. 75, 173-203.

1931 The Scandinavian Drift or Basement Clay on the Durham Coast. Proc. Geol. Assoc. 42, 292-294.

Upham, W. 1894 Bull. Geol. Soc. Amer. 5, p. 71.

Walton, K. 1966 In personal communication with the writer.

Woolacott, D. 1905 The Superficial Deposits and pre-glacial Valleys of the Northumberland and Durham Coalfield. Quart. J. Geol. Soc. Lond. 59, 64-96.

1920 On an exposure of Sands and Gravels containing Marine Shells at Easington, Co. Durham. Geol. Mag. 57, 307-310.

Woolacott, D. (cont.) 1921 The Interglacial Problem in Northumberland and Durham. Geol. Mag. 58, 21-32, 60-69.

Young, J. 1864 On the former existence of glaciers in the high grounds of the South of Scotland. Quart. J. Geol. Soc. Lond. 20, 452-462.

Young, J. 1966 Analysis of glacial deposits near Fala, Midlothian. Ph.D. thesis, Edinburgh University.