THE GLACIAL GEOMORPHOLOGY OF PART OF THE WESTERN GRAMPIANS OF SCOTLAND WITH ESPECIAL REFERENCE TO THE LIMITS OF THE LOCH LOMOND ADVANCE

by

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Geography Section
City of London Polytechnic

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CONTAINS

PULLOUTS

DECLARATION

The information contained within this thesis represents the results of original field investigations conducted by the author, between January,1979 and June,1983 while a part-time postgraduate student at the City of London Polytechnic. The thesis incorporates material used in submission in 1978 for a degree of Master of Science awarded by the Council for National Academic Awards.

Signature ....P. W. Thorp....

Date ....30 th January, 1984....

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Abstract

The limits of a large ice mass that built up in the western Grampians during the Loch Lomond Stadial were mapped using O.S.maps to scales of 1 : 10,000 and 1 : 25,000. The main forms of mapped glacial evidence comprised moraines, thick till, fluvioglacial landforms, erratics, boulder spreads, ice-moulded bedforms, striae and friction cracks. Outside the glacial limits the main types of mapped periglacial evidence included frost-riven bedrock, thick fossil screes, smooth debris-strewn slopes, tors, and solifluction lobes, terraces and sheets. 199 mapped trimlines, based on various forms of contrasting glacial and periglacial evidence, enabled the upper limits and form of the glaciers to be reconstructed, to varying degrees of accuracy, especially in the accumulation areas of the former glaciers. This information was supplemented by the evidence on 64 cols. The reconstructed form of the main ice mass indicates that it covered an area in excess of 2,000km^2, that its total volume was <u>ca</u> 460km^3, and that maximum ice-shed altitudes of <u>ca</u> 700 - 750m O.D. were attained in the Glen Nevis-Rannoch Moor-Glen Lyon areas.

Equilibrium firn lines calculated for the main ice mass and for 17 independent corrie, valley and plateau glaciers indicate that firn lines rose from <u>ca</u> 400m O.D. in the SW to +900m O.D. in the NE part of the study area. Trend surface analysis of corrie-floor altitudes, the spatial distribution of amounts of precipitation at the present time, and the equilibrium firn lines of the former Loch Lomond Advance glaciers, indicates a broad correspondence between these factors. Amounts of precipitation during the stadial on the mountains are estimated to have ranged from 3000 - 4000mm yr^{-1} in the SW to less than 1000mm yr^{-1} in the NE.

Glacial evidence outside the limits of the Loch Lomond Advance suggest that the pattern of build-up and directions of ice-flow in ice-sheet times were very similar to those that occurred in the stadial. This implies that the climatic parameters that operated during the stadial were broadly similar to those that operated in earlier glacial periods.

Acknowledgements

I owe much to Mr C. Squires who first introduced me to western Scotland and who aroused in me an enthusiasm for the Quaternary of the area that has remained undiminished over the last 25 years.

During the last seven years I have received a great deal of help and assistance from numerous individuals and organisations. In particular I am indebted to my supervisors, Dr John Lowe and Dr Murray Gray, for their encouragement, helpful advice and constructive criticisms since I first began the present research project in 1977. Since I first read his book 'The evolution of Scotland's scenery' in 1967 and his subsequent research papers the meticulous research work of Dr Brian Sissons has had a great influence on me in the way I have approached the fieldwork in western Scotland. In addition I have greatly appreciated his frank advice and critical comments at various stages in the research programme.

The trend surface program was written by Mick Riley who has been particularly patient with my many requests for help and advice. Kevin Jordan and John Main contributed much with their helpful discussions on the statistical content of the thesis. I am also grateful for useful discussions with Donald Sutherland, Richard Tipping and Stephen Lowe.

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CHAPTER 1

INTRODUCTION

1.1 Reasons for the selection of the study area and aims of the thesis

The author's interest in the study area was first aroused in 1959 when on a college field course in the Fort William area. Brief studies were made of the features of glaciation in the Glen Nevis, Glen Roy, Loch Leven and Glen Coe areas. This initial visit was succeeded by several brief reconnaissance visits during the early 1960's and by fieldwork undertaken during the summers of 1967 and 1968. The main purpose was to obtain information to expand the earlier work on glacial breaching in the area into an undergraduate thesis(Thorp,1968).

No further fieldwork was undertaken by the author in Scotland until 1977. Fieldwork was then carried out for the purpose of collecting morphological data for an M.Sc. thesis in Quaternary Studies(Thorp,1978). The field study area selected was bounded by the Ben Nevis Range in the N, Rannoch Moor in the E, the Aonach Eagach Range to the S and by Loch Linnhe to the W. This area was selected for several reasons:

i) The author had a long but intermittent interest in the glacial geomorphology of the area, spread over a period of nearly 20 years.

ii) Although a number of studies involving the mapping of the limits of a large ice mass in western Scotland, that was inferred to have formed during the Loch Lomond Stadial, had been made(Peacock,1970a,

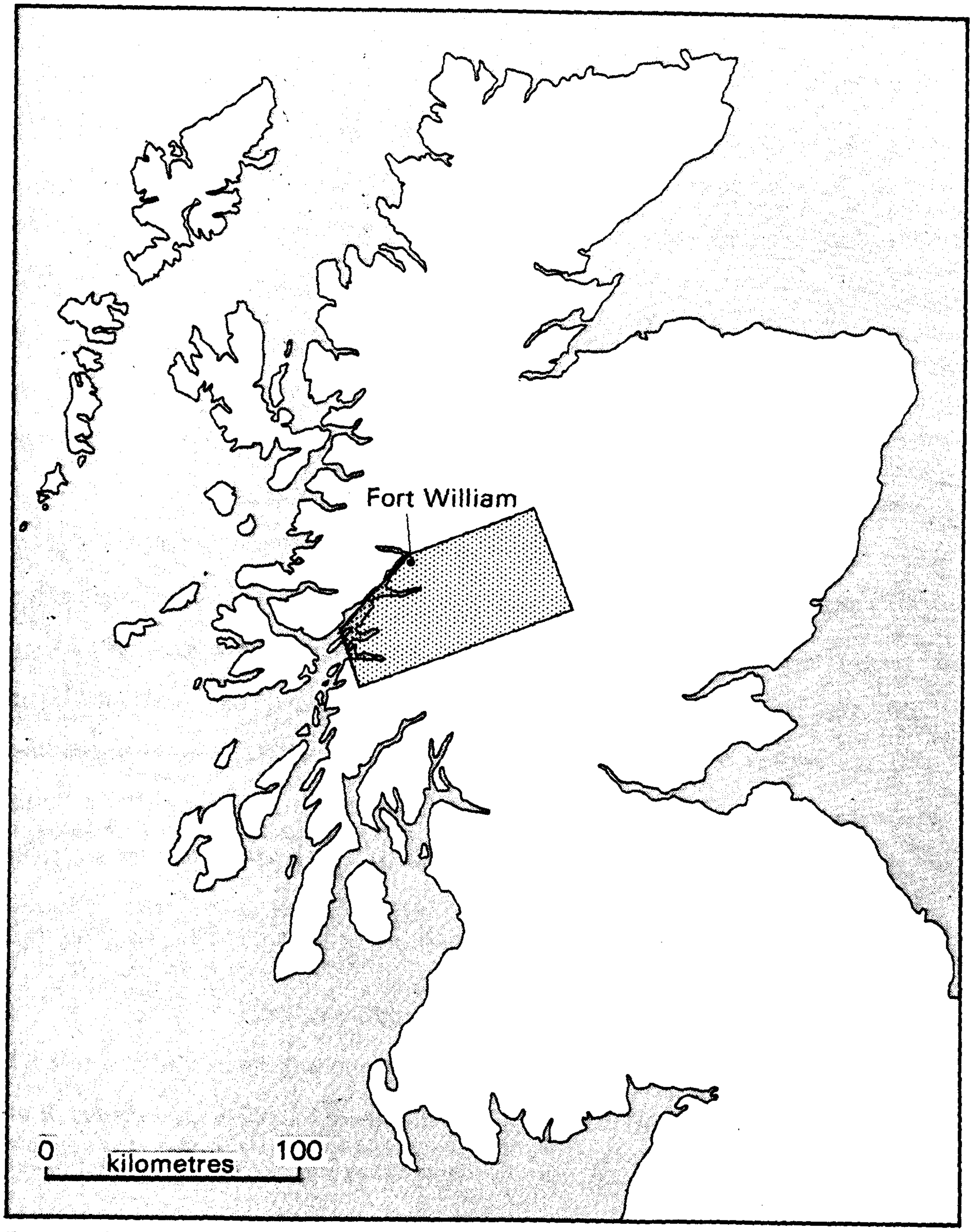


Figure 1.1 Location of the study area

1971a,1971b; Gray,1972,1975a; Thompson,1972; Sissons,1979b) none had dealt specifically with the problems and techniques of mapping glacier limits on the highly dissected mountains on the western side of the former ice mass. A major priority was clearly the mapping of glacial limits within the study area.

iii) To quote Gray(1972, p.6) the area selected was "small enough to permit detailed study but sufficiently large to allow the establishment of a pattern".

iv) The area had largely been neglected by glacial geomorphologists. During the 20 years prior to 1978 not a single research paper had been published on the glacial features in the Glen Nevis area. The only papers to be published relating to the northern part of Loch Linnhe and the Loch Leven areas were those by McCann(1961,1966) and Peacock(1970a,1970b,1971a). The main published work on the glaciation of the area as a whole was by Bailey _et al_(1960) and most of their conclusions were based on fieldwork carried out early in the present century. Thus there was an urgent need for up-to-date information on glacial features in the area.

The results of the fieldwork carried out during 1977 and 1978 formed the basis of the M.Sc thesis. Intensive fieldwork comprising about 26 weeks in aggregate was undertaken during the years 1979 to 1983 and provided much of the information contained in this thesis. Originally the study area extended from the Ben Nevis Range in the N to Ben Cruachan in the S and included the lower area to the W as far as Loch Linnhe. Later the area was increased in size to include Rannoch Moor and parts of the outlet valleys to the E of the Moor(Figures 1.1 and 1.2) to link up with work carried out by Thompson(1972) and Sissons(1979b). Aims additional to those outlined above are listed below:

i) The large outwash spreads at Corran Ferry, North Ballachulish, Benderloch and Connel were generally believed to represent the approximate maximum limits of the Loch Lomond Advance glaciers(McCann,1966; Peacock,1971a,1971b; Gray,1972; Sissons,1976). However, none of the limits had been related to their respective source areas by detailed mapping. Thus an important objective was considered to be an independent check on the proposed maximal glacier limits by mapping morphological evidence over a wide area and especially in the source areas.

ii) Mapping in the Glen Nevis and Loch Leven areas(Thorp,1978) had demonstrated that trimlines, based on contrasts between bedrock abraded and smoothed by glacial processes and bedrock shattered by severe frost-riving, had proved to be especially valuable in delimiting the former upper margin of the glaciers. Mapping in similar mountain terrain farther S would help to test the validity of trimlines based on such contrasts in bedrock surfaces. It would also help to assess whether there was sufficient consistency in the derived trimline values to enable a reasonable reconstruction to be made of the former glaciers, over a much wider area than was possible for the area dealt with in the M.Sc thesis.

iii) In parts of the study area large numbers of striae had been recorded by officers of the Geological Survey(Bailey,<u>et al</u>,1960). In other parts of the study area very few striae had been noted. Yet the underlying causative factors creating this variability had rarely been explained adequately. Nor had it been possible in Scotland to distinguish between sets of striae relating to different episodes of glaciation except in a few cases(Sissons,1977b; Ballantyne and Wain-Hobson,1980). Furthermore, although the author had mapped large numbers of friction cracks in the Glen Nevis and Loch Leven areas in 1977 and 1978, no reference to the mapping of these particular markings on glaciated bedrock in the field in Britain could be found in the literature. Thus it became apparent that the recording of the type,size,number and orientation of all observed glacial markings in relation to rock type and glacial limits would provide valuable information for the reconstruction of the former glaciers and eventually for the purpose of comparison with other areas.

iv) The Loch Linnhe area has several ice-limits which had in most cases been related to different sea levels at the time of their formation(Charlesworth,1955; Donner,1959; Synge and Stephens, 1966; McCann,1966; Gray,1972). Although it was not proposed to undertake any detailed quantitative work on former sea levels, such as that carried out by Gray(1972), nevertheless it would be instructive to map especially the inter-relationships between the Main Rock Platform and cliff, and the deposits and ice-limits relating to the Loch Lomond Advance. This was considered an important objective in view of the controversies surrounding the age and mode of formation of the Main Rock Platform(Gray,1972,1974a,1974b; Sissons,1974b; Peacock,1975).

v) Rannoch Moor is believed to have been a major source area for the successive ice-sheets that developed over the British Isles during the Quaternary period(Linton,1957; Bailey _et al_,1960; Sissons,1967). Thompson(1972) suggested that the altitude of the surface of the Rannoch Moor ice-cap may have reached _ca_ 915m O.D. in the Black Mount area during the Loch Lomond Advance. Sissons(1980) suggested a similar maximum altitude of _ca_ 850 - 900m O.D. for the ice-cap. Since both these values were estimates based on extrapolation a major aim was to undertake detailed mapping of the morphological features in the Rannoch Moor area to determine whether or not these estimates were correct. This could only be achieved by mapping the evidence relating to the upper limit of the ice on all the mountains within and encircling the Moor.

vi) Equilibrium firn lines have been calculated for a large number of independent corrie and valley glaciers and small ice-caps that formed during the Loch Lomond Stadial(Sissons,1979c) but, due to incomplete mapping, very few firn line calculations were available for the main ice mass in the western Highlands. Reconstructions of the main ice mass in the study area would allow the calculation of equilibrium firn lines and enable an approximate independent check to be made on the overall pattern of firn lines for the former Loch Lomond Advance glaciers in the Scottish Highlands, as proposed by Sissons(1979c).

How far the objectives, outlined above, were achieved will be discussed in the following chapters. A recent statement by Sissons(1976,p.131) is relevant here: "Thus in this respect, as with other aspects of the geomorphology of Scotland, a little is known and a vast amount remains to be discovered".One can only hope that this thesis goes one small step further in helping to fill the gaps in our knowledge of the Quaternary of Scotland.

1.2 Organisation of the thesis

The aims and objectives of this thesis have already been outlined in section 1.1 of the Introduction and need no further discussion. The remainder of this chapter includes a brief description of the main features of geological and topographical interest, with particular emphasis placed on the contrasts to be seen in a traverse from E to W across the study area. Such contrasts are of significance to many of the conclusions discussed in later chapters in the thesis. Section 1.4 outlines the main mapping methods used to obtain the data essential for reconstructing the ice-limits and form of the Loch Lomond Advance glaciers. Previous hypotheses for reconstructing former glacier limits in Scotland and in the study area are outlined in section 1.5. The biostratigraphic and radiocarbon dating evidence that has been published for dating the former glaciers in the study area is summarised in section 1.6.

The bulk of the thesis is divided into three main parts, namely (i) the basic field evidence;(ii) the interpretation and spatial implications of such mapped evidence; and (iii) the palaeoclimatic inferences based on the field evidence.

In chapter 2(Part 1) the types of glacial and periglacial evidence that were mapped in the field are described and discussed, in relation to similar studies undertaken in Scotland. The mapping of striae and friction cracks is dealt with in a separate chapter since much quantitative information on striae and types of friction crack, including azimuths of the features and their approximate numbers in relation to rock type, were recorded in the field. Trimlines are similarly dealt with in a separate chapter since they proved to be of considerable importance for reconstructing the form of the main ice mass, particularly in the main source areas of the glaciers. In addition much information was obtained in the field on the types and clarity of the trimlines(Appendix A),in view of the relative neglect of these features in the literature. The field evidence for delimiting the former glaciers in the study area is presented in chapter 5. This is dealt with at some length since it constitutes the basic data on which many of the interpretations and hypotheses in the later chapters depend.

In Part 11 discussion centres on the methods used to reconstruct the form of the Loch Lomond Advance ice mass in the western Grampians and on the significance of the spatial distributions of the different types of field evidence. Attention is drawn in chapter 6 to the usefulness of making glaciological comparisons between present day glaciers in Spitsbergen and the former glaciers,reconstructed from morphological evidence,in the western Grampians.

Part 111 deals with the reconstruction of the climate of the Loch Lomond Stadial at ca 10,500 yrs B.P. and attempts to assess the implication of this for understanding developments during the Devensian and earlier glaciations. The methodological problems of constructing a relatively accurate map of present day precipitation in the western Grampians are outlined in chapter 10. The parameters of 271 corries in the study area are analysed in chapter 11, with particular emphasis placed on the relationships of corrie aspect and corrie-floor altitude to palaeoclimatic influences. Various parameters of the reconstructed glaciers, including firn lines, are used in chapter 12 to make palaeoclimatic inferences about summer temperatures and the spatial distribution of precipitation during the Loch Lomond Stadial. In chapter 13 the evidence provided by erratics,striae,friction cracks and ice-moulded bedforms is used to reconstruct ice-sheet source areas, ice-divides,patterns of growth and ice-flow directions during the Devensian. Comparisons are made between such reconstructions and the form and extent of the ice mass that built up in the western Grampians during the Loch Lomond Stadial.

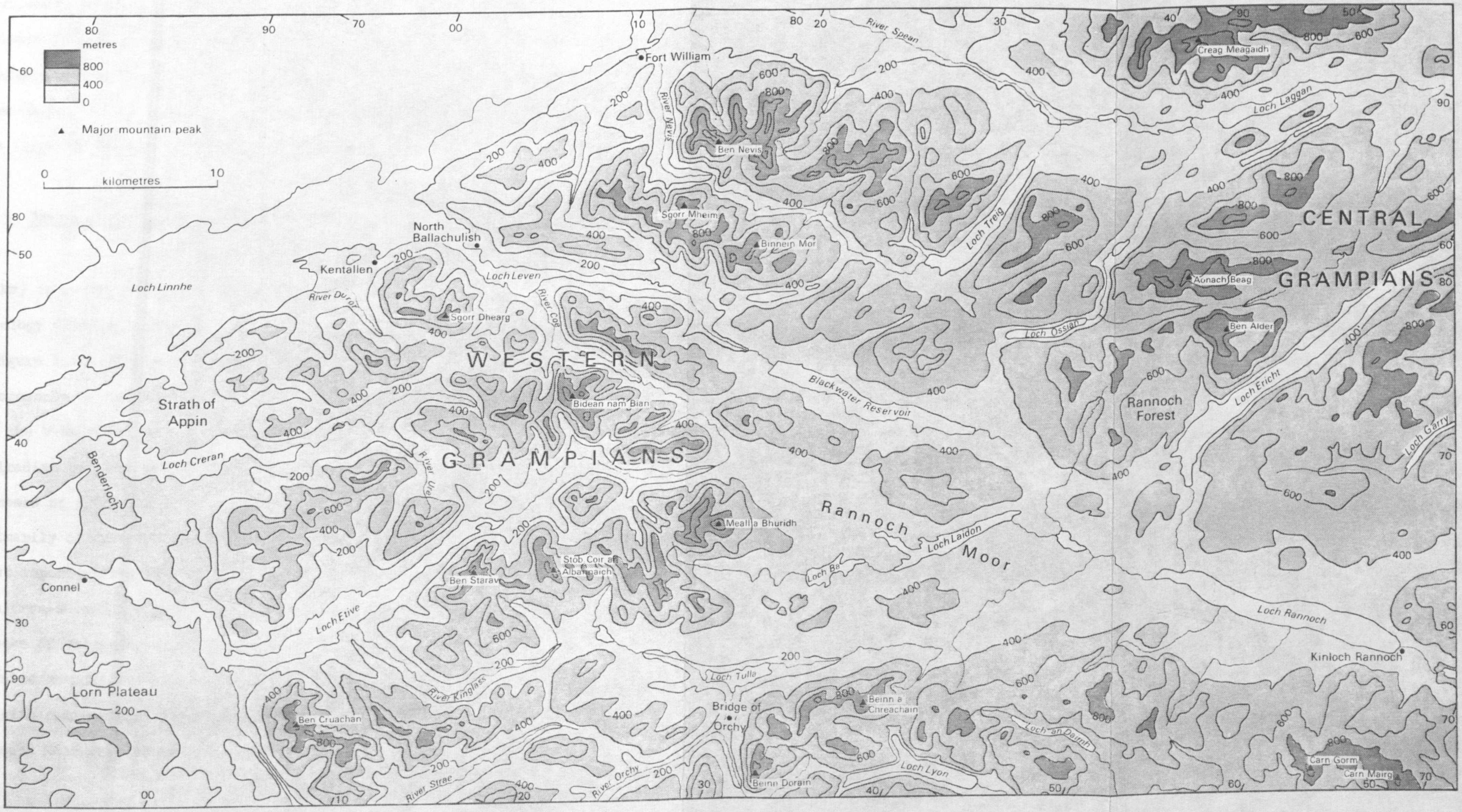


Figure 1.2 Relief and major lochs in the study area and some of the localities mentioned in the text.

In chapter 14 the validity of using morphological evidence, and in particular the use of trimlines, for delimiting former ice-limits is assessed. Particular stress is placed on the value of recording striae and friction cracks, especially crescentic fractures, for determining former ice-flow directions over an area as large as the one described in this thesis.

1.3 Geology and relief of the study area

The study area is sufficiently large enough(85 X 45km) to represent considerable changes in solid and superficial geology (Figure 1.3) and in topography in a traverse from E to W (Figure 1.2). The area mirrors similar changes to be seen on a macroscale across the Scottish Highlands(Linton,1959; Sissons,1967). In the W the solid geology is highly complex with intensely folded Dalradian pelites,psammites and quartzites intruded into by great numbers of dykes and by several major igneous masses, composed primarily of granites,diorites and granodiorites(Figure 1.3). The most important intrusions comprise the Ben Nevis and Glen Coe cauldron-subsidence complexes, that are associated with downfaulted blocks of volcanic rocks, and the Etive,Rannoch Moor and Ballachulish granite/granodiorite intrusions. Farther E the geology is less complex with pelites and psammites predominating and, except for the Rannoch Moor granite intrusion, igneous rocks are relatively sparse.

In the W the landscape is deeply dissected and large numbers of glacial breaches have created low-level gaps through the mountains. Farther E the dissection is less severe; many of the mountains are broader and areas of dissected plateau begin to appear.

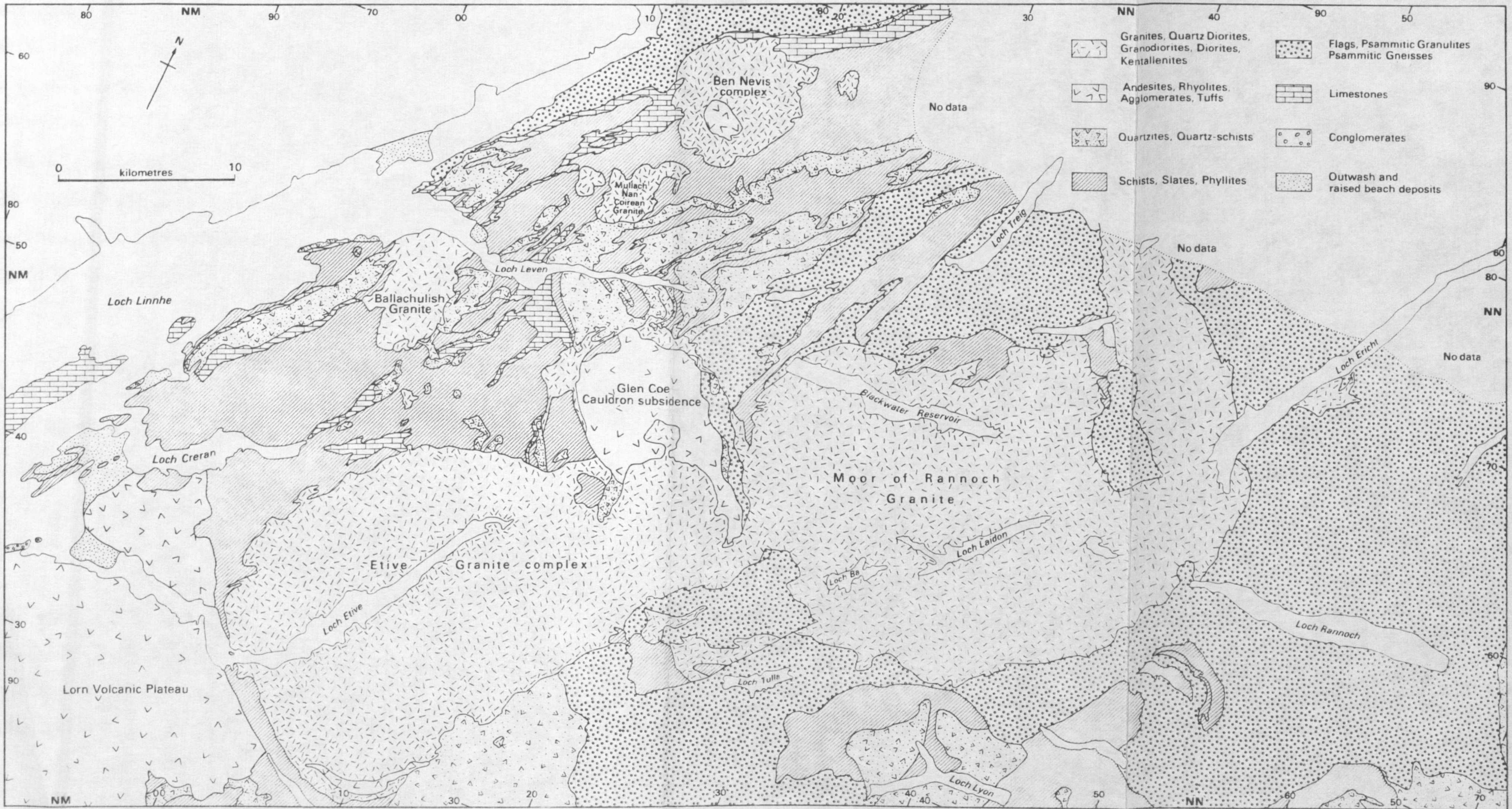


Figure 1.3 Generalised solid geology of the study area based on sheets 45, 46, 53, 54W, 54E, 55, 62E, 63.

(a) <u>The western glens and mountains</u>

Perhaps the most strikingly scenic part of the whole study area is the zone of mountains extending from Ben Cruachan(1126m) in the S to Ben Nevis(1344m) in the N and referred to hereafter in this thesis as the Western Mountain zone. It is an area characterised by deep,penetrating glens,steep slopes,vertical crags,smooth rock walls,pointed peaks and numerous corries. Some idea of the considerable dissection of the terrain by ice is afforded by the numerous glacial breaches and by the fact that <u>ca</u> 70% of the corries in the study area occur within this mountain zone (Figure 11.1).

The Ben Nevis Range, although dominated by the broad mass of Ben Nevis itself, contains 14 other peaks over 1,000m O.D. Only one major breach at the eastern end of the range at <u>ca</u> 500m O.D. interrupts the continuity of the range. Especially noteworthy are the glistening, quartzite screes below the summits of Stob Coire Easain(1080m) and Stob Choire Claurigh(1177m).

The Mamore Forest Range is noted for its graceful, pointed peaks including the two highest of Binnein Mor(1128m) and Sgurr Mhaim(1098m) fashioned from schist and quartzite. These contrast with the rounded, granite summits of Meall a' Chaorain (910m) and Mullach nan Coirean(939m) at the western extremity of the range.

Between the two mountain ranges lies Glen Nevis, noted for its rapid change in morphology from E to W (Bailey,<u>et al</u>, 1960),with a broad,but shallow,U-shaped valley profile in the E contrasting with a very deep, steep-sided, U-shaped cross-profile in the W. These contrasting valley segments are separated by a high rock bar cut through by an impressive gorge. These contrasts in morphology may relate to glacial breaching of the original watershed of western Scotland(just S of Ben Nevis) and a consequent shift of the watershed eastwards to the head of Glen Nevis(Thorp,1968).

The sharp,serrated ridge of the Aonach Eagach Range is formed mainly from the resistant rocks of the Glen Coe Volcanic Series. Cols below 760m O.D. are only to be found at the western and eastern extremities of the range. Impressive crags overlook the corries on the northern side of the range. The steep, south-facing slopes of the range overlook the precipitous Glen Coe carved directly out of the resistant volcanic rocks. Its steep crags, truncated spurs,hanging valleys,corries,U-shaped cross-profile and rock bar, midway along the valley, have rightly earned for Glen Coe numerous superlatives in the travel guides and in the popular literature.

The mountains on the S side of Glen Coe culminate in Bidean nam Bian(1141m), composed of resistant andesites and rhyolites. Here the mountains have been carved into separate blocks by glacial breaching. In contrast, the mountains on the E side of upper and central Glen Etive form a continuous, but irregular ridge, bitten into by numerous corries and containing several peaks exceeding 1,000m O.D., including Ben Starav(1078m) and Stob Coir'an Albannaich(1044m),Stob Ghabhar(1087m),Clach Leathad(1098m) and Meall a' Bhuiridh(1108m). The southern part of this mountain range is characterised by many bare,rock surfaces relating to the sheeting structure of the Starav granite, as exemplified by the upper slopes of Stob Coir' an Albannaich.

Several major morphological contrasts characterise Glen Etive. In the upper part of the glen, leading down from Rannoch Moor, it is narrow and deep with a classic U-shaped profile and flanked by the steep mountains of Buachaille Etive Mor and Sron na Creise. The valley widens to approximately 5km just N of the head of Loch Etive and narrows again to only 2km, where steep slopes sweep down from Ben Starav and Beinn Trilleachan. Beyond these mountains the glen widens out to attain a width of about 7km, before narrowing to $2\frac{1}{2}$km at the constriction created by the Ben Cruachan massif (1126m).

The mountain ridge separating Glen Etive from Glen Creran is generally lower than elsewhere in the Western Mountain zone with just two peaks exceeding 900m O.D. Moreover there are numerous cols and breaches at low altitudes of ca 220 to 550m that slope from E to W across the ridge. The deepest breach is located where the River Ure flows from Glen Etive into Glen Creran along a steep-sided, rock-walled defile.

A similar area of low mountains and many cols at low altitudes exists in the SE part of the zone between Glen Orchy and Glen Kinglass. Here the rocks are mainly pelites, quartzites and psammites that tend to give rise to lower mountains than those formed on the resistant Etive granites farther W. However, over-riding by ice during the early stages of each glacial period is probably an additional factor, since the mountain summits show extensive rounding and smoothing by ice.

(b) Rannoch Moor

Rannoch Moor occupies a basin approximately 400km^2 in area that averages 300 - 400m O.D. in altitude. The basin largely coincides with the outcrop of Rannoch Moor granite, with the great majority of the surrounding mountains composed of metamorphic or volcanic rocks. In a few areas the granite forms ground higher than 700m O.D., as along the ridge from Beinn a' Chrulaiste to Stob na Cruaiche across the centre of the Moor and in the Rannoch Forest area.

Radiating out from Rannoch Moor and cutting through the encircling mountains is a series of impressive glacial breaches(Linton,1957). The fiord of Loch Leven and the glens of Coe, Etive and Orchy breach the western rim of the mountains. To the NE the deep breaches of Treig,Ossian and Ericht cut through mountains largely composed of pelites and psammites. By far the largest outlet from the Moor is by way of Loch Rannoch. This loch occupies a valley that achieves a maximum width of about 12km at the western end of the loch. Thereafter the valley narrows eastwards to only 2 -3km at Kinlochrannoch at a constriction resulting from resistant quartzites.

(c) <u>The eastern glens and mountains</u>

A number of important changes in the morphology of the landscape becomes apparent when compared with the Western Mountain zone. These can be generalised as follows:

i) An increase in the proportion of land between 200 and 500m O.D.

ii) The landscape is more subdued with many more mountains with rounded summits and gentler slopes,except where corries occur and selective erosion has taken place(as along the fault-guided Loch Ericht).

iii) Areas of poorly dissected plateau begin to appear as in the area between lochs Ericht and Rannoch.

iv) The number of clearly defined corries declines to 84 compared with 187 in the Western Mountain zone(Figure 11.1).

North of Loch Laggan the land rises to culminate at the broad,flat-topped summit of Creag Meagaldh(1130m). High corries bite into the serrated edges of the massif, especially on its eastern side.

Loch Laggan occupies a broad NE to SW orientated area of lower ground that is diversified by low hills and mountains rising to 600m O.D. in places, as along the S side of the loch.Many of the valleys and mountain ridges between lochs Laggan and Ericht are controlled in their orientation by strong NE to SW Caledonian structural trends in the bedrock. Several mountains exceed 1,000m O.D. including Geal Charn(1049m) and Beinn a' Chlachair(1088m), but physically dominating the area by virtue of their sheer bulk are Ben Alder(1148m) and adjoining mountains where seven peaks exceed 1,000m O.D. The flat-topped massifs here are only exceeded in the area of land above 1,000m O.D by the Ben Nevis Range(Figure 1.2).

To the S the mountains of the Rannoch Forest, formed from Rannoch Moor granite and psammitic granulites, comprise a series of narrow ridges with mainly poorly-developed corries along their eastern edges. The majority of peaks here lie between altitudes of 800 and 950m O.D.

The mountains that lie between lochs Ericht and Rannoch tend to be subdued in outline and are separated by only shallow glens, except for those descending to the Pass of Drumochter, such as the glen occupied by Loch Garry. Only three well developed corries could be identified in this area.

Similar mountains with subdued and rounded summits, averaging 700 to 1,000m in height,occur along the southern edge of the Rannoch valley. Corries are largely absent except below the summits of Carn Gorm(1029m) and Carn Mairg(1042m). Farther W toward the head of Glen Lyon and S of Loch Daimh slopes become more precipitous, with deeply incised glens and numerous corries.

1.4 Mapping methods

Initially, vertical aerial photographs to a scale of ca 1 : 24,000 were used to identify major forms such as moraines, limits of thick till and erosional bedforms before commencing fieldwork. As air-photograph coverage was only available for the NW quadrant of the study area this was only undertaken for about one-sixth of the total area. Most of the basic data were obtained from field mapping alone. In the Western Mountain zone maps at scales of 1:10,000 and 1:10,560 were used, mainly because at the time of the survey maps at a scale of 1:25,000 were not then available for some areas. In the the remainder of the study area maps at a scale of 1:25,000 were used. All the field data were eventually transferred to Ordnance Survey maps at a scale of 1:50,000(sheet numbers 41,42,49,50 and 51).

Exposures of superficial deposits were examined wherever possible. Important differences in stone clast shape and lithology were noted and the maximum thicknesses of till and fluvioglacial deposits measured or estimated. Slope angles were measured with the aid of an Abney level. The distribution of erratics was mapped by noting and recording the location of free-standing boulders on maps to scales of 1:10,000 or 1:25,000. In some areas boulders revealed by exposures in till or moraines were recorded where boulders

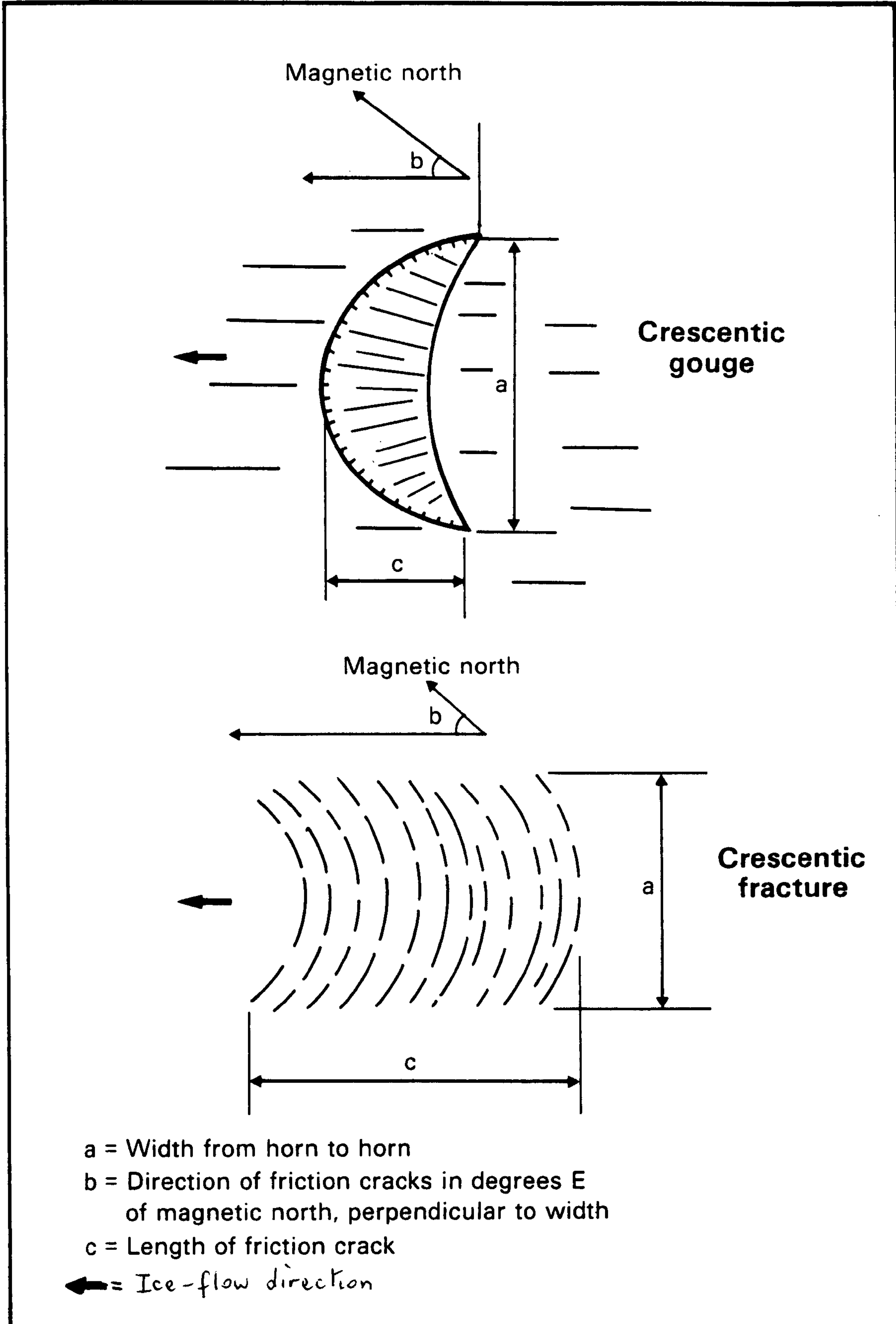


Figure 1.4 Measured parameters of crescentic gouges and crescentic fractures

on the surface were few in number.

No attempt was made to measure the height and slope of raised shorelines or of fluvioglacial landforms by means of instrumental levelling as this technique was considered to lie outside the scope of this study. The primary considerations were to map the spatial distribution of glacial,fluvioglacial and periglacial features over a wide area and to determine any former glacial limits.

Areas of exposed bedrock were noted and their surfaces examined for microscale features such as striae and friction cracks, termed glacial markings in this thesis. Where striae were located their approximate number,orientation and the rock type on which they occurred were noted. The main parameters measured on crescentic fractures and crescentic gouges were as follows(Figure 1.4):

i) distance from horn to horn;

ii) maximum length of the marking;

iii) orientation of concavity measured perpendicularly to i) in degrees from magnetic north; and the

iv) rock type on which they occurred.

Crescentic gouges clearly influenced by planes of weakness in the bedrock were not measured. In some locations the dip and orientation of the bedrock surface was recorded. This was not done systematically for all glacial markings on bedrock surfaces as the primary aims were to obtain former ice-flow directions and to distinguish between those markings inside the glacial limits from those outside the limits. A more detailed study of the several thousand glacial markings identified in the study area would have required time and effort beyond the scope of this study(For a fuller discussion of the methods and problems

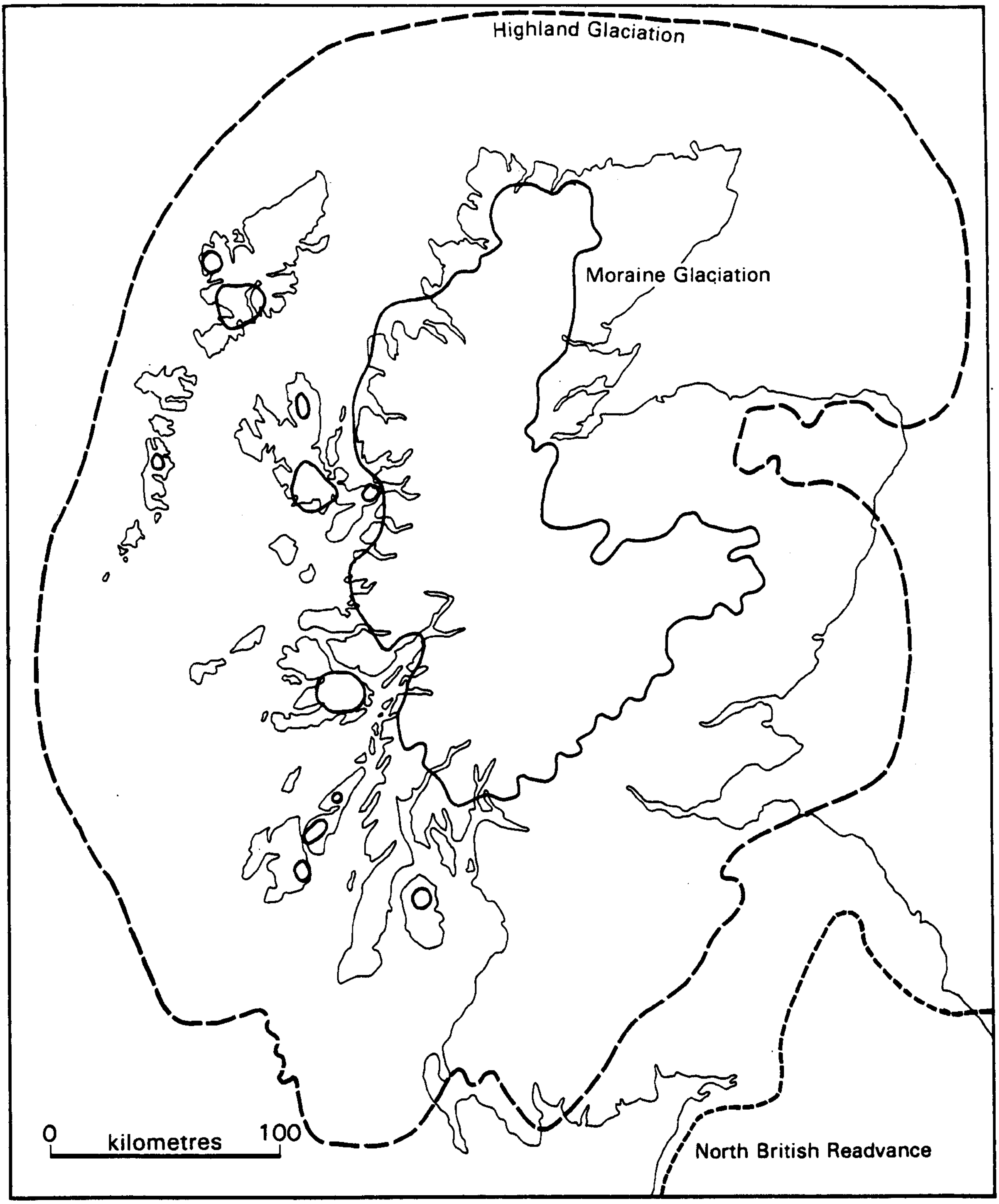


Figure 1.5 Scottish late-glacial ice-limits of Charlesworth (1955)

of mapping glacial markings see Thorp(1981a) and chapter 3 in this thesis).

The methods used to map former glacial trimlines and the problems of interpreting such evidence requires fairly lengthy discussion and is therefore dealt with in chapter 4.

1.5 Ice-limits in the field area: previous hypotheses in the literature

It had long been recognised in Scotland that the recession of the last ice-sheet was succeeded by an advance of valley glaciers(Chambers,1853; Geikie,1863; Simpson,1933). Charlesworth (1955), however, was the first to attempt to reconstruct detailed ice-limits in the Scottish Highlands and Islands(Figure 1.5), although many of his ice-limits have subsequently proved to be incorrect. Donner (1957) investigated pollen stratigraphy in relation to Charlesworth's limits and found that 'Allerod'(Zone 11) deposits were absent within his Stage M limit(now equated with the Loch Lomond Advance), but were present outside the limit. He concluded that Charlesworth's 'Moraine glaciation' took place during Younger Dryas(Zone 111) times. Charlesworth's contention that the ice-limits of the 'Moraine glaciation' could be linked to a so-called 100ft sea level was also shown to be incorrect and instead they have been related to a lower sea level (McCann,1961,1966).

Sissons(1967) reinterpreted the evidence and published a map of the Loch Lomond Readvance ice-limits based on available data(Figure 1.6). A modified outline(Figure 1.7) based on more detailed mapping has been published more recently(Sissons,1979d).

Some detailed ice-limits for the study area,based on detailed mapping, are shown on Figure 1.8. Several ice-limits are

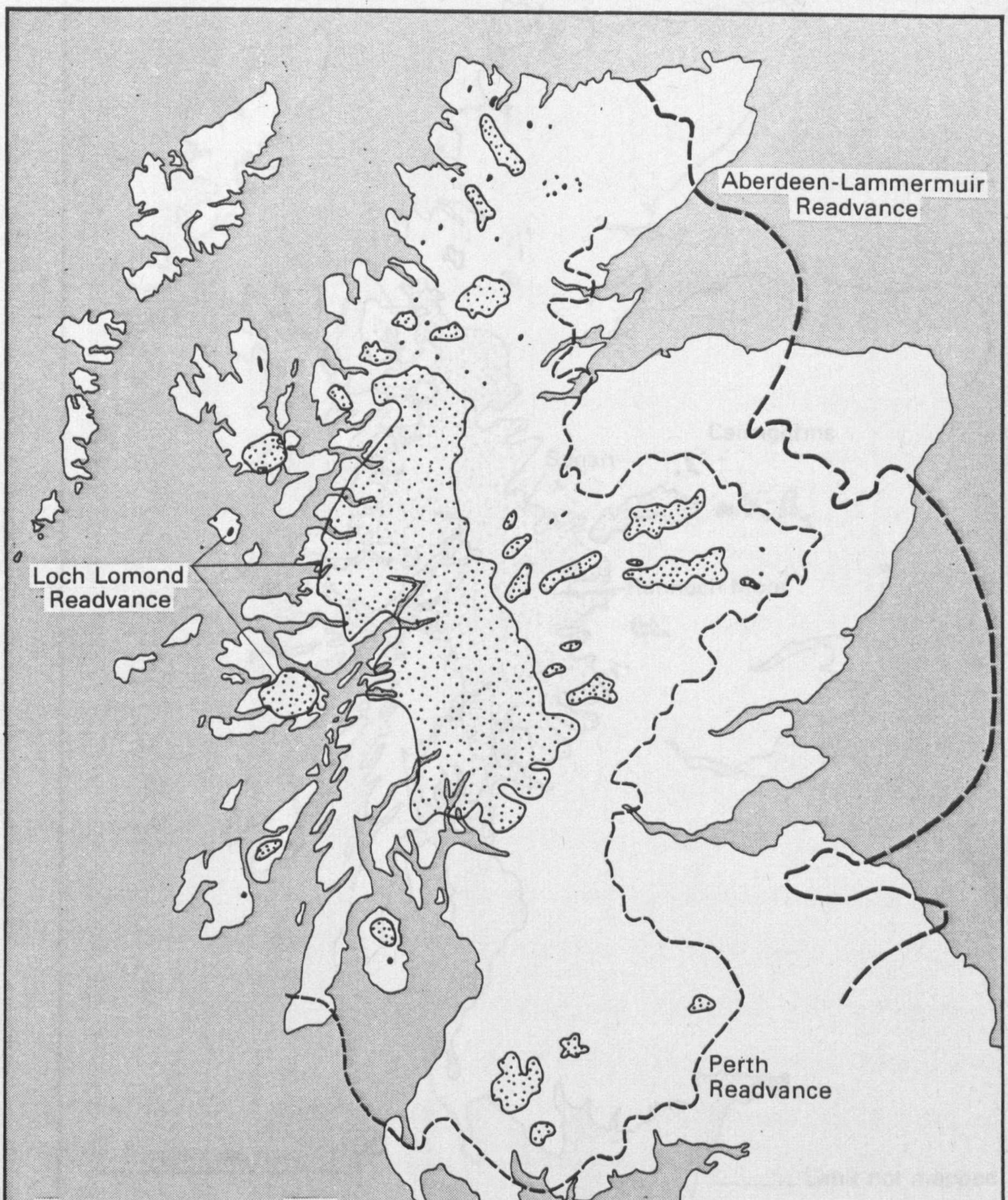


Figure 1.6 Scottish late-glacial ice-limits of Sissons (1967)

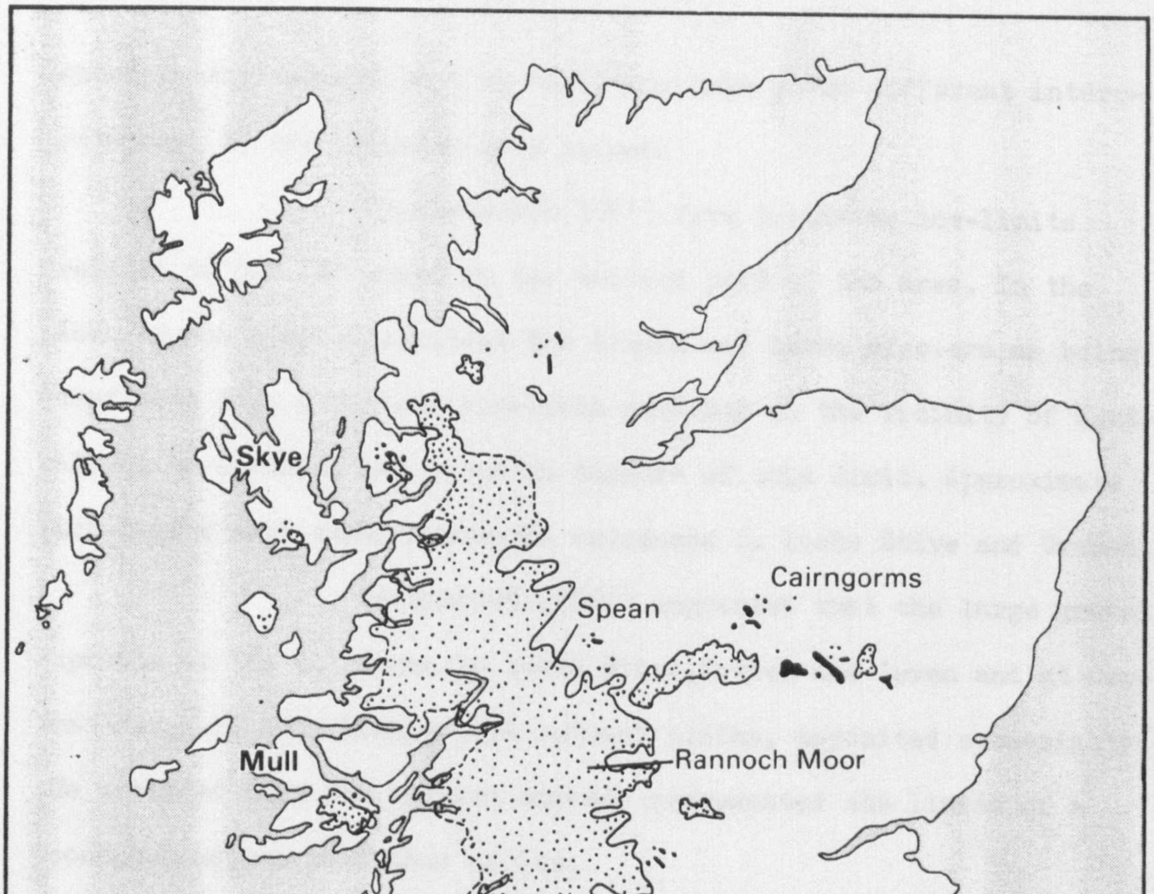


Figure 1.7 Loch Lomond Advance ice-limits of Sissons (1979d)

shown in the western part of the study area where different interpretations of the evidence have arisen.

Charlesworth(1955) drew tentative ice-limits related to his 'M' stage in the western part of the area. In the Loch Linnhe area he depicted the Linnhe and Leven glaciers as being confluent with a maximum ice-limit existing in the vicinity of Kentallen. No evidence was cited in support of this limit. Approximate ice-limits were shown across the entrances to lochs Etive and Creran.

McCann(1961,1966) suggested that the large gravel spreads at the entrances to lochs Etive,Creran and Leven and at Corran Ferry in Loch Linnhe were outwash plains, deposited subaerially. He proposed that the outwash spreads represented the limits of a contemporaneous readvance of ice.

Modifications to some of McCann's limits were proposed by Peacock(1971a,1977). He considered that most of the outwash deposits related to retreat stages within the Loch Lomond Advance limit and postulated ice-limits beyond the proximal margins of the outwash spreads. Gray(1972,1975a) also proposed minor modifications to McCann's ice-limits. For example, he suggested that the maximum extent of the Etive glacier was possibly represented by deposits S of Saulmore farm(NM893334), 2km SW of Connel Bridge. He further suggested, on the basis of identification of a broad, inferred end moraine SW of Lochan Dubh(NM905396), that the Creran glacier extended farther W than was proposed by McCann.

In the eastern half of the study area glacial evidence and ice-limits were mapped by Thompson(1972) in the Glen Lyon and Loch Rannoch areas. A large outwash plain at the eastern end of Loch Rannoch was considered to represent the maximum extent

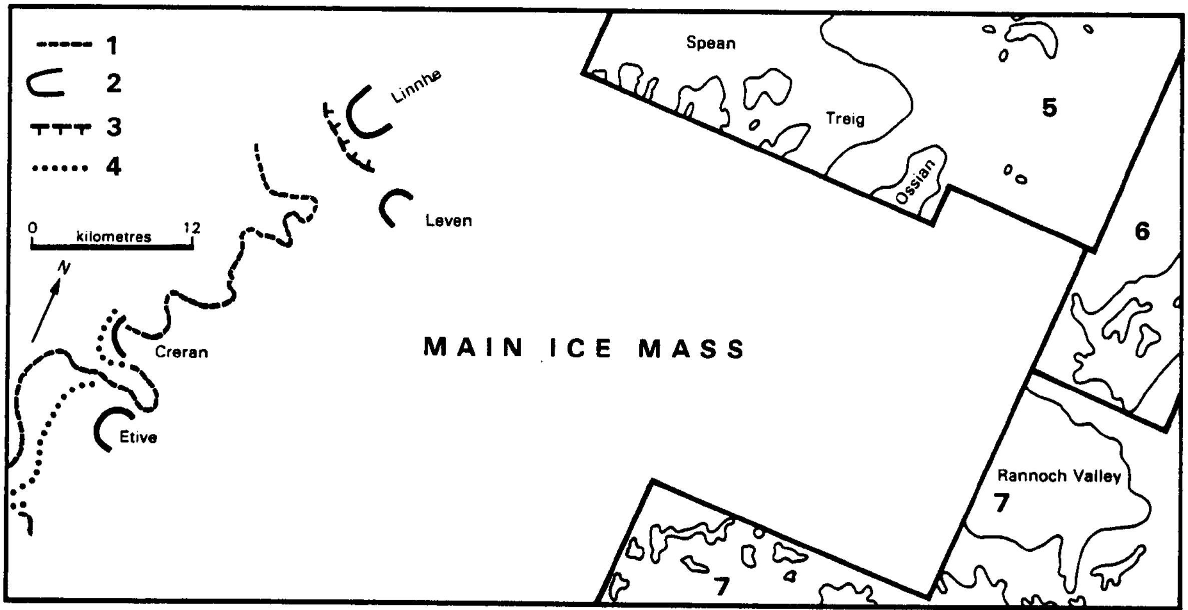


Figure 1.8 Limits of the Loch Lomond Advance in the study area mapped prior to 1982 according to 1 Charlesworth (1955), 2McCann (1966), 3Peacock (1971a), 4 Gray (1975a), 5 Sissons (1979b), 6Sissons (1980), 7 Thompson (1972)

of the Rannoch glacier, while well-developed lateral moraines were used to reconstruct its former margins. Extensive areas of hummocky moraine were depicted within the proposed limits.

Farther N a wide range of morphological evidence, including many pronounced end and lateral moraines, was used by Sissons (1979b) to map ice-limits from Glen Roy to the Strath of Ossian. In addition the limits of a number of small independent glaciers were mapped. Parts of the area have been discussed in more detail in other papers(Sissons,1978,1979e,1981a), particularly in relation to the features associated with the ice-dammed lakes of the Roy-Laggan area.

The ice-limits shown on Figure 1.8 on the high ground between lochs Ericht and Rannoch are derived from Sissons(1980, Figure 2, p.33), but no evidence was cited for the reconstruction of these particular ice-limits.

Except for the work undertaken by Officers of the Geological Survey(Hinxman et al, 1923: Bailey et al, 1960), Ballantyne's (1979) study of glacial Loch Tulla and its associated shorelines and work carried out in the vicinity of lochs Creran and Etive by Gray(1972,1975a), little work of a geomorphological nature has been undertaken over the remainder of the area. It is this area that is largely described and discussed in the following chapters.

1.6 Dating the former glaciers

Radiocarbon dates of 11,430$\pm$ 220, 11,530$\pm$ 210 and 11,805$\pm$ 180 B.P have been obtained from molluscan shells in glacially-disturbed marine clays at South Shian(NM908422) by Loch Creran (Peacock,1971b). Since these deposits lie beneath glacial materials, the ice advance must be more recent than ca 11,500 B.P and it is

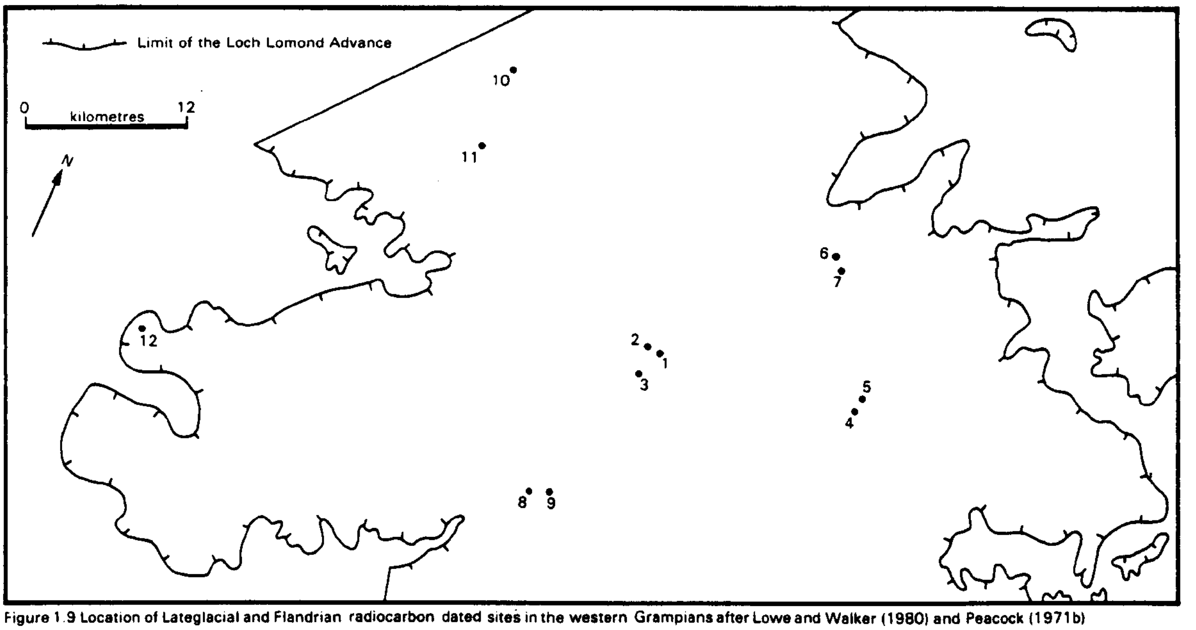


Figure 1.9 Location of Lateglacial and Flandrian radiocarbon dated sites in the western Grampians after Lowe and Walker (1980) and Peacock (1971b)

1 Kingshouse 1	7 Corrour 2
2 Kingshouse 2	8 Clashgour 1
3 Kingshouse 3	9 Clashgour 2
4 Rannoch Station 1	10 Lairigmor 1
5 Rannoch Station 2	11 Lairigmor 2
6 Corrour 1	12 Creran 1

generally assumed that the only period of time cold enough to develop glaciers is the Loch Lomond Stadial. Hence the ice-limits of the Creran glacier are generally assigned to this period. As the evidence presented in this thesis will show that the former Creran glacier was contiguous with the large ice mass shown in Figure 6.1 such an ice mass clearly relates to the Loch Lomond Stadial. In addition the maximum limits on the E side of the ice mass have been shown to be related to the same stadial by Sissons(1979b).

No radiocarbon dated lateglacial site has yet been found in the study area,,although Donner(1957) suggested a lateglacial age for sediments at Pulpit Hill near Oban that lie a few kilometres outside the study area. A detailed analysis of the pollen stratigraphy in the sediments at this site confirms this view(Tipping,pers.comm.).

Deposits in eleven enclosed basins within the Advance limits in the study area(Figure 1.9) have been investigated by Lowe and Walker(1976,1980). The biostratigraphic evidence indicates only Flandrian pollen while the radiocarbon dates obtained from the basal limnic sediments in the kettle holes support the view that the area was occupied by ice during the Loch Lomond Stadial.

The smaller independent glaciers shown in the study area are believed to have formed contemporaneously with the main ice mass. This is based on abundant evidence(summarised in Sissons,1979d) for the existence of only one glacial advance during the lateglacial period. Thus all the ice-limits shown on the figures in this thesis are believed to belong to the same advance, namely the Loch Lomond Advance.

The types of evidence used to delimit such glaciers are presented in Part 1 while the spatial implications of the evidence are discussed in Part 11.

PART 1

DESCRIPTIONS OF TYPES OF FIELD EVIDENCE

CHAPTER 2

FIELD MAPPING OF THE GLACIAL AND PERIGLACIAL EVIDENCE

2.1 Introduction

The primary aim of this chapter is to describe the main forms of morphological evidence that have been used to reconstruct former ice-limits and ice-flow directions in Scotland and to discuss the main problems associated with the mapping of such evidence.

A wide range of morphological evidence relating to the Loch Lomond Advance has been used to reconstruct former glacier limits and surfaces(Thompson,1972; Gray,1972,1982a; Gray and Brooks,1972; Sissons,1972,1974a,1977a,1977c,1979a,1979b; Sissons and Grant,1972; Robinson,1977; Ballantyne and Wain-Hobson,1980; Cornish, 1981; Thorp,1981b). Such studies have strongly emphasised that it is the distribution of the total assemblage of the different forms of glacial and periglacial evidence that is significant in delimiting the areas covered by the Loch Lomond Advance glaciers. Although the different types of evidence are largely dealt with separately in this chapter it is to be stressed that such a structure is only adopted for convenience.In chapter 5 the evidence will be dealt with on a collective basis, where the glacial limits that have been recognised in the study area will be discussed in detail.

The main forms of glacial evidence include end, lateral,medial,fluted and hummocky moraines,thick till,boulder spreads, erratics,striae,friction cracks,ice-moulded bedrock and fluvioglacial landforms. This type of evidence has been complemented by the mapping of periglacial evidence outside the Loch Lomond Advance limits. Such evidence has consisted primarily of fossil periglacial lobes, terraces and sheets, strongly frost-riven bedrock, fossil screes, blockfields and fossil sorted polygons and stripes.

Examples of such forms of evidence and the problems associated with their interpretation in the study area are discussed below in sections 2.2 to 2.14. Discussion of the significance of the spatial distribution of many of the forms of evidence will, however, be deferred until chapters 7 and 8.

2.2 Moraines

End and lateral moraines have been utilised extensively to delimit the former maximal extent of many Loch Lomond Advance glaciers in Scotland. Many such examples have been mapped in the NW Highlands(Sissons,1977c,1979a), on Skye(Sissons,1977b), on Rhum(Ballantyne and Wain-Hobson,1980), on Mull(Gray and Brooks, 1972) and in the Southern Uplands(Cornish,1981).

Within the study area end moraines, and in some cases lateral moraines, have been identified at or near the maximal limits of the Treig and Ossian glaciers(Sissons,1979b), along the southern margin of the Rannoch glacier(Thompson,1972) and at the terminus of a glacier flowing E from Ben Alder(Sissons,pers.comm.). In the W end moraines have been located in the Strath of Appin(Gray, 1972) and at the seaward ends of Loch Creran(Peacock,1971a; Gray, 1975a) and Loch Etive(Gray,1972).

Only a few end and lateral moraines, additional to those quoted above, have been identified by the writer in the study area. On the east side of the main ice mass end moraines formed at three localities, where the Rannoch glacier flowed northwards into tributary valleys(Figure 7.1). In each case the evidence demonstrates that the main glacier failed to push very far into the tributary valley. Instead the ice-margin formed a very steeply inclined surface that ran straight across the tributary valley. On the S side of the Rannoch glacier,end and lateral moraines indicate that the ice managed to reach the col at NN537512, but failed to penetrate into the valley to the E. Two other possible end moraines were identified during the present survey, one at the western end of Glen Duror(NN000554) and the other S of Loch Leven in Gleann an Fhiodh (NN072552). NW of Loch Ossian lateral moraines, believed to delimit the upper surface of the former Ossian glacier, were mapped at three places on the mountain sides at altitudes between ca 600 and 670m O.D. (Figure 7.1).

A few medial moraines were identified in the study area, sometimes in the form of a steep-sided ridge(e.g.on the S side of Glen Creran at NN014446), or more frequently as a complex series of ridges and mounds(e.g. E of Loch Etive at NN100400). Their location in the lee of a spur between two valleys or between two corries, together with their generally low altitude in relation to inferred trimlines and other field evidence in the immediate vicinity, enabled them to be distinguished from end moraines(Figure 2.1). Moreover their alignment,in relation to the spur and the orientation of the valleys, indicates the confluence of two glaciers rather than the margin of a glacier at the entrance to a tributary valley that lacked a glacier (cf Figures 2.1(a) and 2.1(c)).

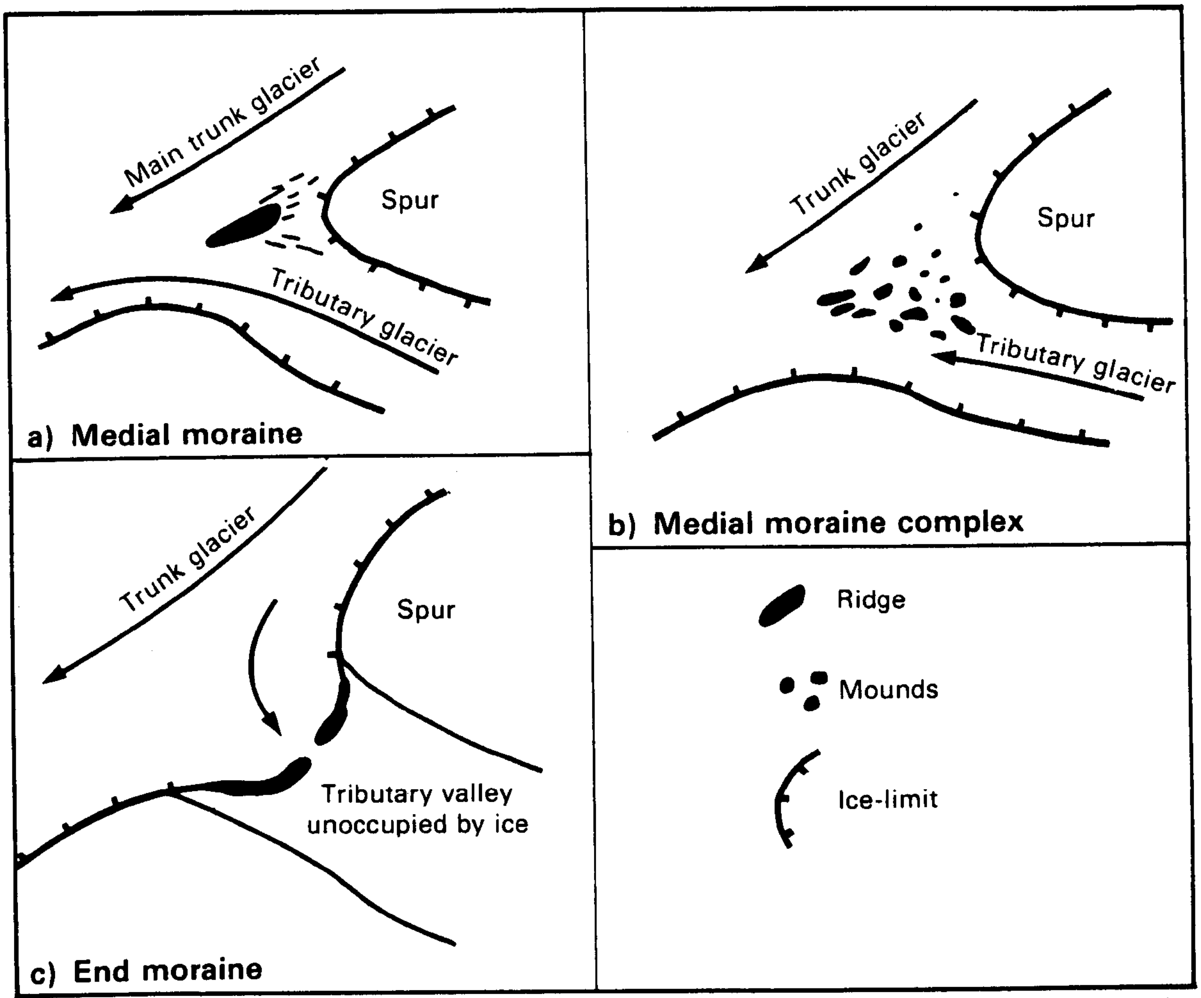


Figure 2.1 Schematic illustrations of the location of end and medial moraines at the confluence of two valleys

2.3 Hummocky moraines

Hummocky moraines comprising steep-sided mounds, usually strewn with boulders, characterise many localities in the study area. Sections in the mounds often indicate a wide range of materials and structures that reflect varying glaciological conditions. The most common materials comprise boulders of varying sizes and shapes within a gritty matrix. A crude stratification may be present in some mounds while in others clear sorting of sands and gravels suggest a fluvioglacial origin. Since it is impossible, in most cases, to differentiate between kames and hummocky moraines on the basis of morphological criteria,as most mounds lack sections,the general term hummocky moraine is applied to all such mounds in this thesis unless specified otherwise.

Various studies(Thompson,1972; Gray and Brooks,1972; Sissons et al,1973; Sissons,1974a,1977a,1977b,1977c,1979a,1979b) have suggested that hummocky moraine is only found within the Loch Lomond Advance limit as defined by end moraines and such evidence can therefore be used to map former glacier limits where end moraines are absent or feebly developed. However, doubts have been expressed on the validity of using hummocky moraines as a means to delimit former glacier margins(Sugden and Clapperton,1975). Furthermore a few areas of supposedly hummocky moraine have been mapped outside Loch Lomond Advance limits,as defined by end moraines and associated landforms (Sissons,1977c). A primary aim of the mapping programme was,therefore, the mapping of all clearly defined mounds in the study area in order to assess their relationships with all other forms of mapped evidence and with the inferred ice-limits.

2.4 Till

Gray and Brooks(1972) used the upper limit of deeply gullied drift to partly delimit the margins of former glaciers on Mull, although this particular work was based mainly on the interpretation of aerial photographs and not on field evidence. Sissons(1974a,1977c,1979b) has used the upper limit of gullied drift to delimit the margins of a number of former glaciers in the eastern and central Grampians, the NW Highlands and in the Glen Roy area. For example, Sissons(1974a, p.97) states that thick drift plastered on a valley side has a sharp upper limit that rises up-valley and appears as a small step on the valley side. In Glen Roy a massive drift accumulation, roughly 80m thick, occurs at the northern limit of the inferred tongue of ice that pushed into Glen Roy(Sissons, 1979b).

Substantial deposits of till,ranging from ca 5m to 40m in thickness,occur at many locations in the western Grampians. In a number of localities the abrupt termination of thick till delimits or helps to delimit the former glacier margin. For example, such limits are found on the N side of Loch Creran(NM965456), N of Loch Etive(NM98736), N of Glen Lyon(NN431440, NN503480 and NN492483), E of Loch Treig(NN353723) and N of Loch Rannoch(NN538655 and NN543640).

2.5 Fluvioglacial landforms

Outwash spreads are found at or a short distance within the Loch Lomond Advance limit at a number of places in Scotland(McCann,1966; Peacock,1970a; Gray,1975a; Sissons,1974a,1977c, 1979b), while kettled,kame terraces are particularly notable in the Callander area(Smith et al,1978).

In the study area large outwash spreads occur at or near the maximal limits of the former Rannoch(Thompson,1972), Creran(Peacock,1971a),Etive(Gray,1975a),Linnhe and Leven(McCann, 1966) glaciers. Fine examples of kettled kame terraces have been mapped by Gray(1975a) along both sides of Loch Etive,

Very few additional fluvioglacial outwash landforms were mapped during the present study. Only in two areas were these outwash landforms used to delimit former glacier margins. For example, a series of outwash fans occurs along the southern edge of Loch Linnhe near Kentallen(NN006573), while a dissected outwash train declines in altitude towards the SW from the exit of Glen Duror(NN000551 to NM985545). These outwash landforms are believed to represent the maximal limits of the Linnhe and Duror glaciers respectively. SW of Loch Ericht a series of kame terraces flanks the edges of a shallow valley(NN471629), but these are small features rarely more than 5m broad and only 2 - 3m high.

Eskers have been mapped on the slopes above Loch Rannoch(Thompson,1972) in Strath Ossian and N of Loch Treig(Sissons, 1979d) and near lochs Creran and Etive(Gray,1975a). The only examples of eskers mapped additional to those above were SW of Loch Ericht(NN471629 and vicinity) and along the southern edge of Rannoch Moor(NN412482 and vicinity).

2.6 Meltwater channels

Numerous examples of meltwater channels have been identified in many parts of Scotland. Fine examples of large and complex anastomosing channel systems have been mapped in the Southern Uplands(Sissons,1960,1961). Meltwater channels are also well displayed in the Cairngorms(Sugden,1970) and in NE Scotland(Clapperton and Sugden,1977). In contrast to these areas the study area contains comparatively few meltwater channels.

Meltwater channels are primarily useful for indicating regional ice-flow trends,since their alignment often reflects the pressure conditions existing at the time of their formation,that direct meltwater streams toward the glacier margin(Clapperton and Sugden,1977). Care,however,needs to be exercised with their interpretation since the hydraulic pressure gradients will change during thinning of an ice mass and the orientation of the channels will no longer be directed by the regional ice flowlines. The value of meltwater channels in determining former glacier limits is generally less. Exceptions include Sissons'(1974a) mapping of numerous small and delicate meltwater channels within the limits of a plateau ice-cap in the Gaick area,that contrast with the much larger channels that occur outside the ice-limits. In this thesis their main use has been to supplement the information derived from ice-direction indicators such as striae,friction cracks and roches moutonnées.

2.7 Boulder spreads

A number of studies(Sissons,1977c,1979a; Ballantyne and Wain-Hobson,1980; Cornish,1981) have stressed the usefulness of boulder spreads in helping to infer glacier limits, since in many areas such spreads frequently terminate abruptly at limits defined by lateral and end moraines.

Boulders are abundant in many parts of the study area, but in some areas the number of boulders increases dramatically.

Examples of such extensive boulder spreads are to be found W of Loch Etive(NN050410 and vicinity), where huge numbers of large boulders of Starav granite to 3m in length litter the ground, and in upper Glen Creran(NN078512 and vicinity), where many large schist boulders were derived from the corries at the head of the glen. Other notable examples occur SW of Loch Treig(NN280680 and vicinity) and W of Loch Ericht (NN470650 and vicinity).

In many areas extensive boulder spreads are closely associated with hummocky moraines and in some places the moraines may be largely composed of heaps of boulders(e.g. NW of Ben Nevis at NN143738). In numerous localities the distribution of boulders clearly relates to source areas such as the rock walls of corries and steep valley-side crags; numbers of boulders quickly diminish in a down-glacier direction from such source areas.

The number of free-standing boulders on a valley side may often diminish abruptly on the higher part of the slope, above the inferred ice-limit,even though the angle of slope remains the same. Such approximate boulder limits can provide useful supplementary evidence for ice-limits derived from other forms of ice-marginal evidence. For example, numerous large boulders of schist (ca 1 - 1½m) are spread across the floor and the lower slopes of the valley immediately to the N of the head of Loch Etive(NN114500).Yet they are largely absent on the upper slope and ridge summit on the E side of the valley above the inferred ice-limit at ca 550 - 600m O.D. However, in some areas the reverse may occur where large numbers occur outside the inferred ice-limits, as for example, on the ridge N of Ben Nevis(NN164740) where many large andesite boulders occur in solifluction lobes. These examples are generally exceptions since areas outside the glacier limits described in this thesis are usually characterised by relatively low numbers of free-standing boulders.

2.8 Erratics

Previous studies in Scotland that used erratics, as distinct from boulder spreads, to delimit the former margins of glaciers are relatively few. Exceptions include Sissons'(1974a) use of the presence of Rannoch Moor granite erratics on parts of the Gaick plateau to suggest the absence of locally nourished ice in those areas during the Loch Lomond Stadial. In contrast Cornish(1981) used the presence of greywacke erratics,in association with hummocky moraine,to identify some of the Loch Lomond Advance limits in the western Southern Uplands. Such an interpretation was only possible because ice-flow movement during the Loch Lomond Stadial took place into the Loch Doon basin whereas under ice-sheet conditions a radial flow of ice occurred out of the basin(Cornish,1982). Few such clear limits to erratics that corresponded with inferred glacier limits were located in the study area and, therefore, their primary use was to determine ice-flow directions(see chapters 7 and 13).

Erratics have long been used to determine the direction of former ice-movements in different parts of Scotland(*e.g.* the reports of the Boulder Committee of the Royal Society of Edinburgh from 1878 to 1884; Bailey *et al*, 1960). Their continuing usefulness is demonstrated by recent studies involving the mapping of erratics(Peacock,1970a,1970b: Sugden,1970; Shakesby,1976; Coward,1977; Flinn,1978; Cornish,1981,1982). Yet in view of their comparative usefulness the total number of detailed studies of the distribution of erratics is still quite small and indeed as recently as 1974 this led Sissons(1974c, p.319) to state that in Scotland "studies of erratics are still preliminary".

In the study area former ice-flow directions were determined by mapping the distribution of erratics from a number of different bedrock sources. These comprised the Ben Nevis granite, Mullach nan Coirean granite, Rannoch Moor granite, Cruachan granite, Starav granite, Ben Nevis volcanics, Glen Coe volcanics and various Moinian rocks(Figure 1.3).

However, the usefulness of erratics can be diminished and their distribution open to mis-interpretation due to two main problems. Firstly, in a region such as Scotland, that has undergone multiple glaciation,erratics of an earlier glaciation may be redistributed by a later one. Secondly, mis-identification of a rock type can occur and can result in an erratic being assigned to the incorrect source rock. For example, facies variations can occur within a source rock, as in some igneous intrusions such as the Etive and Ben Nevis granite complexes, which show concentric zones of granite with differing compositions and textures, emplaced at different times. It is thus possible to obtain two very similar hand specimens of igneous rock from what are supposedly two differing intrusions with dissimilar compositions. Very careful petrological analysis is therefore required to distinguish between the two specimens and this is rarely practicable except over a limited area.

Examples of the first type of problem occur at several places in the study area. Abundant boulders of Rannoch Moor granite occur in the valley of Gleann a' Chaolois(NN153600) yet a wide range of evidence in this area enables a reconstruction of the former Loch Lomond Advance glaciers and demonstrates that ice did not reach this valley from the Rannoch Moor area during the Loch Lomond Stadial(see p.96). This suggests that the granite boulders were initially deposited by the ice-sheet, but were redirected during the Loch Lomond Stadial. Such a sequence of events was hinted at by Bailey et al(1960, p.277), "The crossing of boulder trails... is well illustrated in Gleann a' Chaolais... This glen, even at its head, retains much of the material brought by the main westward ice-flow, including great numbers of Moor of Rannoch 'granite': and yet right down to its mouth it is strewn with blocks of andesite carried down-valley from the crags of Aonach Eagach during the later stages of glaciation'.

A similar example of possibly redirected erratics occurs along the northern flanks of the Ben Nevis massif, but since the interpretation of these erratics is intimately linked to ice-limits a full discussion is given in section 5.5.

An example of the second type of problem occurs in upper Glen Etive leading down from Rannoch Moor. Many granite boulders in the glen,that were originally thought to be Rannoch Moor granite in origin(Thorp, 1981b),are now thought to be more likely derived from local outcrops of Cruachan granite; in places the two granodiorites are very similar in hand specimens.

Other areas where the second problem was encountered include the area S and SW of Loch Tulla, parts of Glen Creran and along the shores of Loch Linnhe. This was due to the mingling of granite erratics from two or more different intrusions and correct identification,.with any degree of certainty, became impossible in the field.

2.9 Ice-moulded bedrock

The presence of strongly mamillated rock outcrops, and especially the presence of roches moutonnées, is usually regarded as classic evidence for the existence of former glaciers since the early acceptance of the Glacial Theory (MaClaren, 1849; Jamieson, 1862). Many studies have used such criteria for helping to delimit the extent of former glaciers and ice-sheets in many parts of the world(see general summaries in Flint, 1971; Embleton and King, 1975; Sugden and John, 1976). In Scotland the presence of ice-moulded bedrock has been used, in conjunction with other forms of evidence, to infer the former extent and ice-flow directions of the last ice-sheet (Bailey, et al, 1960; Clapperton and Sugden, 1977; Flinn, 1978; Cornish, 1982). More recently a number of studies have suggested that the mapping of areas of strongly ice-moulded bedrock, in conjunction with the mapping of frost-riven bedrock, can be used as a valid method for delimiting the former ice-limits of the Loch Lomond Advance glaciers(Sissons, 1977a, 1977b; Thorp, 1978, 1981a; Ballantyne and Wain-Hobson, 1980). Thus the testing of such a hypothesis was one of the primary aims of this thesis(see chapter 4).

Many fine examples of ice-moulded bedrock exist in the study area. Microforms include rock surfaces that still retain their glacial polish, grooves, channels, striae and friction cracks (the latter two features will be discussed in chapter 3). Macroforms include roches moutonnées, extensive glacial pavements and rock knobs and hills streamlined by ice.

On most rock outcrops the glacial polish has been weathered away. However, in special circumstances excellent examples of 'fossil' freshly deglaciated rock surfaces can be preserved(cf Gray and Lowe,1982 and Gray,1982b for the Snowdonia area). In the study area such examples occur at the S end of Loch Treig, where the use of the loch for supplies of water for hydro-electric power since the 1930's has led to considerable fluctuations in the loch surface. During periods of low water roches moutonnées composed of psammitic granulite are exposed whose surfaces have been washed free of drift and which display numerous grooves,crescentic gouges and polished surfaces that are virtually unaffected by subaerial weathering.

Other excellent examples of polished rock surfaces can be found along a newly-created track leading to Loch Ericht(especially in the vicinity of NN481611). The lack of subaerial.weathering can be attributed to the protection afforded by a covering of drift and peat until very recently. Worthy of note is the fact that the surfaces of aplite veins in the granite are 'flush' with the surface of the main granite body. Elsewhere in the study area aplite veins almost always project above the main bedrock,due to their high resistance to subaerial weathering processes. For example, reddish-tinged aplite veins are especially abundant in the porphyritic Starav granite. Invariably the veins, where exposed to the atmosphere, project above the surrounding granite surface to maximum heights of 6-8cms. The glacial polish has been weathered away entirely or partly from many of the veins, especially where the veins are traversed by joints or fine cracks, although rough,pitted remnants may still be present in places. In other places the glacial polish may be completely preserved as indicated by numerous striae on a very smooth surface. In contrast the surface of the granite is rough and pitted where the felspars and mafic minerals have been weathered away leaving a rough surface of quartz grains that are eventually loosened and washed away.

Roches moutonnées and rock outcrops streamlined by the ice abound in the study area from sea level up to altitudes of ca 900m O.D. and especially in the Western Mountain zone. Particularly fine examples of roches moutonnées are to be found in schist in Glen Nevis(NN159685 and NN176670), in rhyolite in upper Glen Coe (NN188562) and in schist in Glen Creran(NN034475).

Roches moutonnées are particularly useful for helping to determine former ice-flow directions, although care is needed in the field as the orientation of roches moutonnées can be strongly controlled by structures within the bedrock(Rastas and Seppala,1981). Such structures include bedding planes,joints and lineations(e.g. schistosity and cleavage planes).

An important observation noted by Demorest(1937) and which has been given little attention in the literature is that where the ice-flow direction more-or-less coincides with the direction of dip of the bedrock then it possible for 'reversed' roches moutonnées to form. On the basis of the glacier flow directions inferred in this study such examples of 'reversed' roches moutonnées occur in psammitic-type rock NW of Loch Tulla at NN257443.

Thus where the orientation of roches moutonnées was measured care was taken to record the strike and dip of the bedrock, wherever possible, and to check the orientation data with the ice-flow directions derived from evidence such as striae and friction cracks. In areas where striae or friction cracks were absent or poorly represented as large a number of roches moutonnées as possible were measured for their orientation. Such conditions apply to areas such as Glen Lyon,Glen Daimh,the western part of Rannoch Moor, the area W of Loch Tulla,the central part of Glen Creran and the area SW of Loch Treig(Figure 8.1).

It is interesting to note that a number of hills and ridges with summit altitudes of ca400 to 700m O.D. in the study area have a pronounced roche moutonnée form. Good examples occur S of Loch Ossian(NN381666) and in Glen Nevis(NN179698), but they are most abundant in the area W and SW of Loch Treig(e.g. at NN300691 amd NN315685). The features strongly suggest that roches moutonnées form a continuum from microscale features less than a metre in length and less than half a metre high to macroscale features that can be a kilometre or more in length and hundreds of metres high. Whatever their size it appears that the glaciological processes involved are basically the same.

2.10 Periglacial evidence - a discussion

Many active and fossil periglacial features can be found on the upper slopes of Scottish mountains. The evidence for present day periglacial activity in the form of minor periglacial features such as small stone polygons and stone stripes,ploughing blocks,solifluction terraces and some solifluction lobes has been summarised by Sissons(1976, p.108-109). However, there are abundant examples of periglacial features that are difficult to ascribe to the present time. These consist of severely frost-riven bedrock, blockfields,thick stable screes,large stone-banked solifluction lobes and terraces, and slopes strewn by angular debris and possibly smoothed by former solifluction processes. Reviews of studies of periglacial features in Scotland by Gray and Lowe(1977) and Sissons(1979d) and recent detailed quantitative research by Ballantyne(1981) on periglacial landforms in the Northern Highlands of Scotland favour the view that the formation of these features largely took place under the severe periglacial conditions of the Loch Lomond Stadial, although the initiation of some or all of the features during ice-sheet decay could be possible.

Thus a primary aim of the field mapping programme was to map as many relic periglacial features in the study area as possible. If the views outlined above were correct such relic features should only occur <u>outside</u> the limits of the former Loch Lomond Advance glaciers. Such mapping would also clearly complement the work to be undertaken in mapping the distribution of ice-moulded bedrock to determine trimlines(chapter 4).

The main types of periglacial features mapped in the western Grampians are described in sections 2.11 to 2.14 while the implications of their spatial distribution are discussed in chapter 7(section 7.9).

2.11 Frost-riven bedrock

Many bedrock exposures in the study area show evidence of being affected by severe frost-riving, or macrogelivation, particularly on the upper slopes and summits of the mountains. Many exposures display angular edges, deep open joints and loose angular blocks, partly prised away from the rock *in situ* by frost-wedging along joints and bedding planes. The conditions that produced such frost-riving are largely inoperative at the present time as many of the rock exposures are covered or partly covered by mosses, lichens or by a vegetation mat; freshly frost-riven debris rarely occurs.

Nevertheless, the evidence afforded by the appearance of rock outcrops is not always unambiguous. One important reason is that bedrock composition,texture and structure strongly influence resistance to frost-riving(Wiman,1963; Potts,1970; Thorp, 1981a,1981b; Ballantyne,1981). A qualitative assessment of the degree to which a particular rock type is susceptible to frost-riving is shown in Figure 2.2. This is based on observations of a wide range of rock types in the western Grampians. The figure must only be regarded as a crude approximation to reality since considerable variations can occur within one so-called rock type e.g. schists can be massive or highly fissile while volcanic rocks can vary tremendously in texture,crystal size and jointing characteristics and these can greatly influence the effectiveness of frost-riving.

The rock types most resistant to freeze-thaw action are those with widely spaced joints such as the coarse-grained granites and some of the massive schists. Undoubtedly the most resistant of all the rock types encountered in the study area is the Starav granite. This granite forms extensive areas of huge curvilinear sheets of bare rock on the mountain sides as a result of pressure-release jointing parallel to the slope. Many loose boulders of Starav granite average 1-3m in length and their size relates to the sparse jointing sets in the bedrock. Even at altitudes greater than 700-800m O.D. scree is frequently minimal,ice-moulded

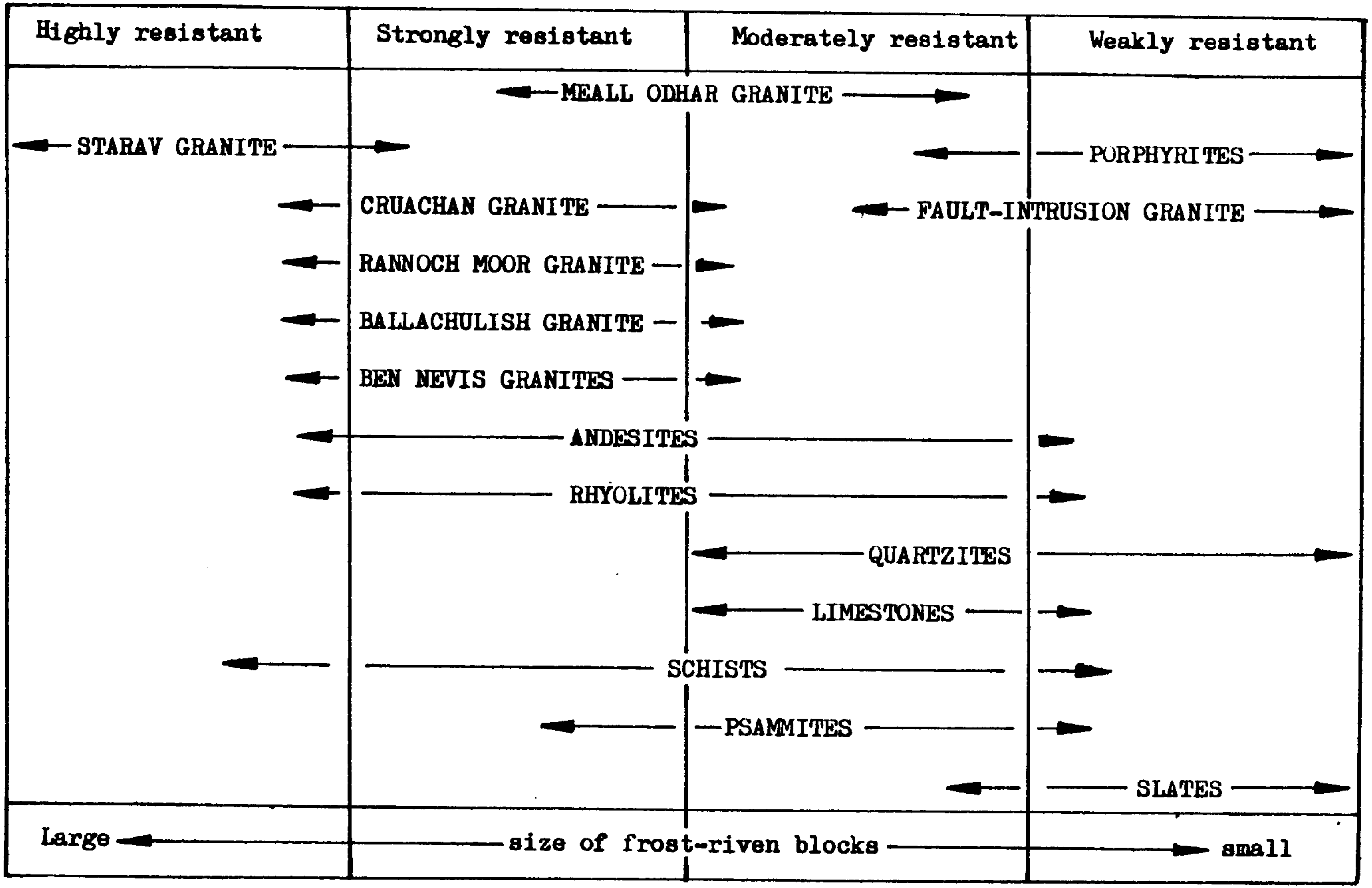


FIGURE 2.2 Qualitative ranking of rock types in study area according to susceptibility to frost riving

surfaces are common and solifluction lobes,terraces and aprons are poorly represented on the Starav granite. The general lack of solifluction forms can clearly be related to the lack of sufficient quantities of frost-riven debris e.g. along the summit ridges of Beinn nan Lus(NN 1237) and Beinn Trilleachan(NN 0843) in the Loch Etive area.

Conversely, rock types that contain numerous closely-spaced joints include some of the quartzites,porphyrites (mainly a dyke rock) and the fault-intrusion granites related to the Glen Coe cauldron-subsidence igneous complex. These shatter very easily and many examples of strongly frost-riven rock can be located on valley sides and floors to relatively low altitudes. Good examples occur in the valley NE of Stob Coire nan Lochan (NN 157556) and on the south-facing slope of Garbh Bheinn(NN 166599). These rock types are usually associated with great quantities of frost-riven debris and numerous solifluction forms.

It is interesting to note that Ballantyne(1981) has stressed the importance of lithology in determining the nature of past and present periglacial activity. He concluded that where macrogelivation was dominant, as on quartzite and siliceous schists, blockfields,blockslopes and massive boulder sheets were the dominant features. He contrasted this activity with the deflation surfaces, sand sheets, non-vegetated debris-mantled slopes and turf-banked terraces that tended to form on granitic rocks.

Thus any visual assessment of the degree to which a given rock type has been affected by frost-riving must take into account the nature of the rock.

2.12 Screes

Screes are especially abundant in the study area because of the altitude and steepness of the deeply dissected mountains. They range from altitudes as low as 10-15m O.D. as along the cliff edge of the Main Rock Platform at NM 863331 to altitudes greater than 1000m O.D. near some of the highest summits in the study area.

Scree is actively forming at the present time below many steep corrie walls,gullies,rock chutes and crags, especially if the bedrock is easily frost-shattered. Yet in many cases much of the present day activity, inferred from the fresh light colour of the scree fragments, appears to be related to the redistribution of fossil scree material from the upper slopes to the lower slopes by means of slope wash and stream flow. Furthermore, thick screes, either covered by mosses and lichens or partly vegetated, are characteristic of many localities on the upper slopes of the mountains and are clearly fossil.

Therefore, during the mapping programme a simple classification was adopted with the screes assigned to three categories, namely active,partially-active and inactive in order to detect any significant spatial distributions(see chapter 7).

2.13 Solifluction terraces,lobes and sheets

Although the present writer has not carried out any systematic and detailed quantitative work on features formed by solifluction in the study area, observations and a few measurements confirm that some of the features are active today. For example, extensive solifluction sheets and terraces with risers averaging ½-1m mantle much of the broad summit ridge of Beinn Bhreac-Liath (NN 303339) above an altitude of ca 700m O.D. This ridge lies just S of the study area. In some places the edges of two turf-banked solifluction sheets have merged together(indicated by a shallow linear depression and by a change in vegetation) while farther along the risers of the two sheets are separated by enclosed linear depressions that may vary in depth from ½-2m and in which water may be trapped. One such depression was measured in July,1980 that was 1-1½m deep, 6m in length and had a maximum width of ca 1m. When the depression was measured again in August,1981 its length had been reduced to 3½m and its maximum width was only ca ¾m demonstrating that movement had occurred in the intervening period of time.

In some areas the steep edges of turf-banked terraces show turf and stones disrupted by recent movements. Good examples of this recent activity were observed at the head of Glen Nevis(NN 267733) at altitudes of ca 900-950m O.D.

Ploughing blocks occur on a number of mountains in the study area and linear furrows in their wake have been measured to over 4m in length. The bow-wave of material on their down-slope and the clarity of many of the furrows strongly suggest active movement at the present time. Excellent examples of ploughing blocks are to be seen N of Loch Treig at NN 312731, although numerous examples occur elsewhere.

Nevertheless, many of the larger boulder-banked solifluction lobes,terraces and sheets appear to be inactive at the present time, as indicated by the general stability of the soil and vegetation cover on many of the features.

All examples of both active and inactive solifluction features were recorded in the field and marked on the base maps since their distribution was considered to be of critical importance in determining their precise relationships with the bedrock and with former glacial limits.

2.14 Smooth debris-strewn slopes and tor-like summits

Many upper mountain slopes, especially those facing W and SW, are characterised by smooth slopes that are littered with angular frost-riven fragments within a gritty matrix. Angles of slope vary considerably from possibly 7-8^{o} to 30^{o} or more. Often severely frost-riven outcrops of rock occur on the upslope side of the deposits, generally along ridges and interfluves; these would appear to be the main source of the rock debris. Whether the debris is undergoing substantial movement today is unknown because of a lack of sufficient quantitative data at the present time.

French(1976, p.152) describes similar landforms from high Arctic areas and his Figure 7.8(p.153) is remarkably similar to the appearance of a number of mountain areas in the study area e.g. N of Loch Etive at NN 165525 and on the slopes of Beinn a' Bhric(NN 319642) although numerous examples occur at other localities.

A number of summit ridges and valley sides have upstanding rock outcrops generally only 1-4m high that could be termed tor-like summits after Ball and Goodier(1970). The term is used purely in a descriptive sense and in no way implies a genetic significance. Some of the tor-like features still bear traces of ice-moulding on their surfaces but the majority have sharp angular outlines and deep open joints and many show no signs of former glacial activity. They are frequently surrounded by aprons of frost-riven debris or by the smooth debris-strewn slopes described above.

Since both types of periglacial feature would appear to be basically relic landforms relating to the lateglacial period(Ballantyne,1981) they were included in the mapping programme.

2.15 Conclusions

It is necessary to map a very wide range of glacial and periglacial evidence in order to minimise errors in the reconstruction of former glacier limits and ice-flow directions. Unfortunately, not all areas in the western Grampians contain clear morphological evidence since spatial variations can occur within the glacial limits, presumably due to differing geological,topographical.and glaciological factors. For example, the floor of the upper part of Glen Creran contains numerous pronounced hummocky moraines and till ridges(e.g. in vicinity of NN 080514) and many large,free-standing boulders of schist and kentallenite, whereas the central and lower part of the glen, between the glacial breach of Glen Ure(NN 070475) and the terminal moraine at S.Shian contains relatively few hummocky moraines, free-standing boulders or thick deposits of till except at the junction of some of the tributary valleys(_e.g._ at NN 040463, and NN 013445). Changes in the velocity of the Creran glacier created by ice spilling through the Glen Ure breach from Glen Etive may account for such strong spatial variations in the morphological evidence. On the other hand topography may have had some influence since the central part of Glen Creran narrows and in addition several large ridges, composed of schist, rise to summit altitudes of _ca_ 160m O.D. above the floor of the glen. These factors are likely to have induced increases in the velocity of the glacier and led to the scouring away of much of the glacial debris in this part of the glen.

The mapping of the glacial morphological evidence is strongly complemented by the mapping of the periglacial evidence, especially on the upper slopes of the mountains, although as with the glacial evidence considerable variations exist in the clarity and abundance of the evidence from area to area. Such variations relate particularly to the inter-relationships between lithology, altitude and the degree of the exposure of the bedrock to macrogelivation.

Some of the clearest periglacial evidence is to be seen on mountain slopes composed of quartzite above _ca_ 800m O.D. as along parts of the Mamore Forest Range and the Ben Nevis Range. Nevertheless, the periglacial evidence is not always so clearly discernible since many mountains are covered by peat or grassed surfaces that lack any clear morphological evidence or comprise types of schist or granite bedrock that at altitudes below _ca_ 500m O.D.

display a surface morphology that is difficult to interpret and that can lead to errors of judgement being made. In addition minor periglacial activity has continued through the post-glacial period(Ballantyne,1981) and this can lead to mis-interpretation of the evidence in the field.

CHAPTER 3

STRIAE AND FRICTION CRACKS : METHODS AND PROBLEMS

3.1 Introduction

The main purpose of mapping striae and friction cracks in the study area was to obtain information on former ice-flow directions. The methods used to identify and map striae and friction cracks together with some of the problems of interpretation are discussed below.

Glacial striae have been recognised and mapped in the British Isles since the mid-nineteenth century(Maclaren,1849; Jamieson,1862). Their continuing usefulness is demonstrated by recent studies in Scotland(Peacock,1970a; Sissons,1977b; Cornish,1982) and in Scandinavia(Aarseth and Mangerud,1974). In contrast, although friction cracks have long been described and mapped in North America(Gilbert,1906; Lahee,1912; Harris,1943; Dreimanis,1953) and Scandinavia(Okko,1950; Virkkala,1951,1960; Anderson and Sollid,1971) very little attention has been paid to such small-scale features in Britain. Exceptions include studies by Thorp(1981a),Gray and Lowe (1982) and Gray(1982b).

The known types of friction crack and fractured bedrock markings described in the literature are shown in Figure 3.1 and comprise crescentic gouges(Gilbert,1906), crescentic fractures (Lahee,1912), conchoidal fractures(Ljungner,1930), lunate fractures (Harris,1943), chattermarks(Harris,1943), jagged groove(Harris,1943), reversed crescentic gouges(Anderson and Sollid,1971) and reversed

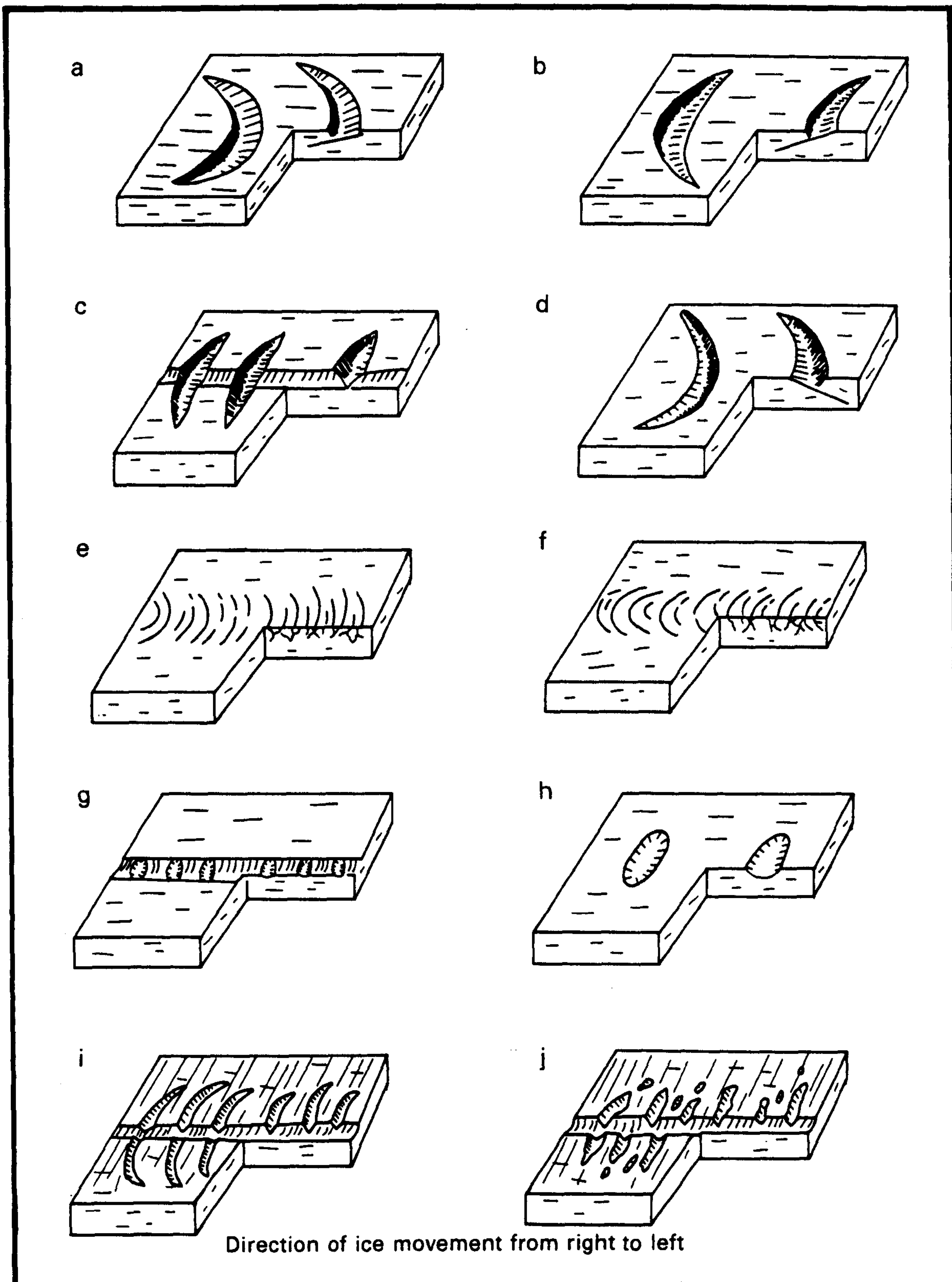


Figure 3.1 Types of friction crack and fractured bedrock markings found on abraded bedrock.

a Lunate fracture (Harris, 1943)
b Crescentic gouge (Harris, 1943)
c Crescentic gouges with groove (Harris, 1943)
d Reversed crescentic gouge (Anderson and Sollid, 1971)
e Crescentic fractures (Lahee, 1912)
f Reversed crescentic fractures
g Chattermarks (Harris, 1943)
h Conchoidal fracture (Ljungner, 1930)
i Transverse 'swallow wing' fracture (Gray, 1982b)
j Transverse friction marks (Gray, 1982b)

crescentic fractures(MacClintock,1953). More recently, Gray(1982b) has suggested, from observations of fracture markings on newly-exposed, glacially-abraded bedrock on the floor of Llyn Peris in Snowdonia that two new types of fractured bedrock markings should be added to those already known. He refers to these as 'swallow-wing' fractures and transverse friction marks(Figure 3.1).

Gray and Lowe(1982) have suggested that the term 'friction crack' could be used to include many of the small-scale erosional forms described in the literature. However, since very few lunate fractures and chattermarks and no conchoidal fractures were mapped in the study area the term 'friction crack' as used in this thesis will be applied in the more restricted sense suggested by Harris(1943) and Embleton and King(1975) and will refer only to crescentic fractures, crescentic gouges and reversed examples of each type. The term 'glacial marking' will be used to refer to all types of marks on bedrock created by ice-flow movements, viz. striae,friction cracks and fractured bedrock markings.

3.2 Mapping striae

Care needs to be exercised in the mapping of striae as agencies other than glaciers can produce striae, as Embleton and King(1975, p.184) have pointed out. These include (i) surface markings resulting from differential chemical weathering; (ii) avalanches; (iii) icebergs and drifting shore-ice; (iv) mudflows and landslides.

The most difficult problem encountered in the field when interpreting inferred striae on rock surfaces were those that occurred on strongly-foliated rocks such as mica-schists,schistose quartzites and, to a lesser extent, some of the psammites. Differential chemical weathering has generally etched out shallow depressions along the less resistant folia and if these have a similar alignment to the inferred ice-flow direction their resemblance to striae can present problems of interpretation.

The problem was generally resolved in the field by noting the micro-morphology of the depressions. Pseudo-striae were identified on the basis of several criteria including (i) their often greater depth in comparison with true glacial striae; (ii) the varying depths they display as a result of varying depths to which weathering has occurred; (iii) the sinuosity of some of the depressions; (iv) their close conformity with minor structures in the bedrock unlike the cross-cutting relationships shown by the true striae. In situations where there was still some doubt on the true origin of markings on a rock surface no record was made in the field.

Many striae have been mapped by officers of the Geological Survey(viz. sheets 45,46,53,54W,54E,55,62E and 63) in the western Grampians. A high proportion of the locations where striae have been recorded by officers of the Geological Survey were inspected in order to ascertain the following: (i) the type of glacial markings present; (ii) the approximate number of markings at any one locality; (iii) the sense of direction of the glacial markings; and (iv) the rock type on which the markings occur.

This enabled the writer to conclude firstly, that an arrow on the Geological Survey map invariably represents a group or set of striae and/or friction cracks and not just an individual marking; and secondly, that the markings mapped were primarily striae and chattermarks.

Ice-flow directions different to those indicated by Geological Survey striae are suggested for four localities(see Figure 8.1). Geological Survey striae 4km to the W of Loch Tulla indicate a SW - directed flow of ice. In this locality there are a number of heavily ice-scoured rock knobs and these were examined very thoroughly for striae, but none was located. However, numbers of excellent roches moutonnées and stoss-and-lee forms exist at this locality and in the vicinity and these provide consistent evidence of an ice-flow directed towards Glen Orchy to the ESE from the mountains at the head of Glen Kinglass. This direction is likely to relate mainly to a stage in the build-up of ice mass in the western Grampians with ice flowing towards the outlet of Glen Orchy before the three cols to the S(see Figure 8.1) were overtopped by ice and began to operate as outlets for the ice-congested basin of Loch Tulla.

The Geological Survey striae at the other localities (e.g. SW of Loch Treig and in upper Glen Lyon) are believed to show the wrong sense of direction and have been reversed on Figure 8.1. This conclusion is based on the orientation of roches moutonnées in the Glen Lyon area and on a wide range of evidence in the Glen Nevis-Loch Treig area including friction cracks,roches moutonnées and the distribution of erratics.

Few striae additional to those mapped by the officers of the Geological Survey were found in the study area. Hence the great majority of non-Geological Survey arrows shown on Figure 8.1 relate only to friction cracks.

3.3 Mapping friction cracks

The widths of friction cracks were obtained by measuring the distance between each horn tip. Widths of crescentic fractures range from a minimum of 2cm to a maximum of 40cm, with the majority less than 10cm. Crescentic gouges and reversed crescentic gouges range from 3cm to 50cm with modal values of 10 - 20cm. Crescentic fractures always occur in series with the number of individual fractures ranging from five to as many as thirty or more. A set of crescentic fractures can extend for distances greater than one metre, although average lengths are generally less than 30 - 40cm. Crescentic gouges can occur singly, in rows commonly two to six in number, or they may be scattered randomly across a bedrock surface, although most will still display a similar orientation.

Approximately two-thirds of all friction cracks observed are comprised of crescentic fractures, with a further quarter accounted for by crescentic gouges. Relatively few reversed crescentic gouges, lunate fractures and chattermarks were observed. These observations on the relative proportions of glacial markings(excluding striae) are similar to those recorded by Harris(1943) and Dreimanis(1953), although Gray and Lowe(1982) found that reversed crescentic gouges were more numerous than 'normal' crescentic gouges in the Llyn Llydaw area of Snowdonia. No 'swallow-wing' fractures or transverse friction marks(Gray,1982b) were observed, although this may relate to the relatively small outcrop areas of slate, and especially to the lack of newly-exposed, abraded slate surfaces in the study area.

Various properties of friction cracks have been suggested as providing a reliable means of determining former ice-flow directions. Gilbert(1906) described crescentic gouges with the horns always pointing up-glacier, but subsequent studies by Harris (1943) indicated the presence of lunate fractures with horns pointing down-glacier, while Anderson and Sollid(1971) mapped examples of 'reversed' crescentic gouges whose horns also pointed in a down-glacier direction(Figure 3.1). Hence the direction in which the horns pointed could not be used as the sole criterion in determining the ice-flow direction.

Harris argued that the principal fracture in friction cracks always pointed forward in the direction of ice-flow. This claim was refuted by Dreimanis(1953) who found that the principal fracture,in the majority of friction cracks in the Cirrus Lake area of Ontario,dipped against the direction of ice-flow.

Field observations by Lahee(1912), Harris(1943), Dreimanis(1953) and Okko(1960) emphasise the consistency of crescentic fractures whereby the horns always point in the direction of ice movement, that is, each fracture is concave forward. However, laboratory experiments by MacClintock(1953), in which a steel ball was used to score optical glass, produced 'normal' crescentic fractures(with horns pointing forward) if the ball was allowed to slide, whereas 'reversed' crescentic fractures were produced by rolling the steel ball. This has raised doubts on the reliability of using the direction in which the horns of crescentic fractures point as a means of determining ice-flow movements, and led Flint(1971) to comment that " possibly therefore all known crescentic fractures were made by sliding tools". Moreover, in a study of contact-induced stress fractures Johnson(1975) argued that crescentic fractures were produced by block sliding whereas crescentic gouges were the product of block rolling. Conversely,Gray and Lowe(1982) and Gray(1982b) suggested that such a mode of formation may not be correct since the 'jagged groove' of Harris(1943) comprises a groove formed by a sliding block that is closely associated with a series of crescentic gouges formed at the same time (Figure 3.1). Such disagreements emphasise the lack of information known about the precise relationships between the movements of debris in basal ice and the ways by which the bedrock is fractured. Until such positive relationships are found inferred examples of reversed crescentic gouges and reversed crescentic fractures will continue to pose problems for determining correct ice-flow directions in the field(see section 3.4).

In view of these considerations, sense of direction of friction cracks was determined according to a number of criteria. Where striae and friction cracks occur on the same outcrop of rock, the crescentic fractures and striae generally show a strong parallelism with only minor deviations from the mean orientation, unless crossing sets are present. Crescentic gouges and reversed crescentic gouges often show a wider dispersion of orientations, exceeding 100° in extreme cases(cf Gray and Lowe,1982), when compared with striae and crescentic fractures on the same rock exposure. This can be attributed to the greater control exerted on these particular friction cracks by schistosity,lineations and joints within the bedrock. These can influence the direction taken by the principal and secondary fractures and the subsequent shape of the rock fragment that is removed(cf Harris,1943,p.246). Conversely no examples of minor rock structures influencing the general form and orientation of crescentic fractures were found; these appear to form independently of minor rock structures if not of rock texture. Thus crescentic fractures are believed to provide the most accurate record of ice-flow directions, and represent a reference datum against which the orientations of other types of friction cracks can be compared. Thus, where present, the orientations of crescentic fractures were measured and recorded in preference to other types of friction cracks. Inferred direction of ice-flow was determined from the direction in which the horns pointed. Discussion of the problems posed by possible reversed crescentic fractures is deferred until section 3.4.

At the relatively few locations where crescentic fractures were absent but other types of friction crack were present, direction was obtained by comparing the measured orientations with the macro-topography, roches moutonnées, the plucked lee-side of rock outcrops and the distribution of erratics.

When friction cracks were located, a series of short traverses was undertaken to ascertain the areal extent of the friction cracks on exposed rock surfaces. Where numbers of friction cracks totalled less than ca 15 - 20 the orientations of all markings were measured. Unless there were widely divergent results the dominant direction was recorded as a single arrow on the base map. Where directions were widely divergent(i.e. $> 90^{\circ}$) a mean value was calculated and entered on the base map. At a few localities crossing sets of crescentic fractures were identified. Mean values were obtained for the two differing directions and are represented in Figure 8.1 by two crossing arrows.

If friction cracks were abundant(i.e. > 20) only a sample selected subjectively, but as representatively as possible, was measured(1 in 2 to 1 in 4 depending on the numbers present). The subjective element is considered unlikely to introduce any unconscious bias into the results, since at the great majority of locations, parallelism of the friction cracks was the norm rather than the exception. Thus in some areas where friction cracks are especially abundant, as at the head of Glen Nevis and on the slopes above Loch Leven, an arrow drawn on maps depicting field evidence in this thesis may represent as many as ca 50 individual or sets of friction cracks, although average values are generally much less than this. Arrows were not drawn on the base map unless sets or individual friction cracks numbered at least five in any one locality. Therefore, in this thesis, each arrow represents a group of glacial markings that can be termed a cluster.

3.4 Problems of reversed crescentic fractures

Opposing sets of crescentic fractures have been observed by the writer at six localities in the western Grampians. Five of the localities occur within the inferred limit of the Loch Lomond Advance. The details are shown in Table 3.1.

Nat.Grid Ref.	Rock type	Altitude of bedrock (m)	Azimuth Normal (Degrees from N)	Azimuth Reversed (Degrees from N)	Thickness of ice (m)
NN 242683	Quartzite	650	70	260	150
NN 143417	Starav granite	600	260	90	70
NN 168394	Starav granite	380	120	300	160
NN 033409	Meall Odhar granite	330	270	90	200

NN 179344	Quartzite	300	180	360	100
NN 059465*	Meall Odhar granite	670	320	140	-

* Outside the limit of the Loch Lomond Advance

TABLE 3.1 Examples of inferred reversed crescentic fractures

At all six locations 'normal' crescentic fractures are far more numerous than the inferred reversed crescentic fractures. The examples at NN 242683 are located on a spur at the western extremity of the Mamore Forest Range and since this is a critical area close to a major ice-shed some discussion is required. The opposing sets of crescentic fractures suggest three possibilities concerning their origin:

(i) The inferred 'reversed' crescentic fractures are in reality 'normal' crescentic fractures, and Loch Lomond Advance ice flowed from the NE to the SW, contrary to the direction proposed in this thesis.

(ii) The markings indicate two different glacial episodes with an earlier phase of markings surviving a later ice advance(_i.e._ Late Devensian ice-sheet markings have survived the Loch Lomond Advance). Gray and Lowe(1982) proposed such an explanation to account for crossing sets of striae in the Llyn Llydaw area.

(iii) The markings indicating a supposed flow of ice toward the SW are true 'reversed' crescentic fractures.

The first hypothesis can be rejected very strongly on the basis of a comprehensive range of evidence(Thorp,1978 and p.125in this thesis) that indicates that a powerful flow of ice occurred toward the NE. The second hypothesis is again difficult to reconcile with the field evidence. Neither of the opposing sets of crescentic fractures show any difference in clarity, although one would expect the inferred earlier markings to be less distinct than the presumed later markings. Furthermore, the thickness of the Loch Lomond Advance ice at this particular locality is estimated at ca 150m which, together with the strongly abraded and fresh nature of the quartzite bedrock surface, suggests that it would be unlikely that the earlier markings could survive such intense glacial abrasion. Therefore the third possibility is the one that is favoured.

If the crescentic fractures described above are true 'reversed' crescentic fractures the implication is that some of the crescentic fractures mapped as 'normal' types at other localities may in fact be reversed types. However, this is considered unlikely because of the strong constraints imposed by the steep terrain and the presence of clear supporting evidence of ice-flow direction at the great majority of the locations(over one hundred) where crescentic fractures have been observed. To infer a diametrically opposed direction of ice movement at most of these locations would be to reconstruct totally implausible glaciers and ice-flow directions. For example, at NN 168394 and NN 179344 the inferred reversed crescentic fractures are concave upslope toward a corrie in each case. To regard this direction as the true ice-flow direction would be to imply that ice was flowing into a corrie up a reverse slope and since no example has yet been mapped of ice flowing toward and into a corrie during the Loch Lomond Stadial this direction can be assumed to be spurious.

This is not to say that this type of situation cannot happen since under ice-sheet conditions many, if not all, corries were submerged by ice and ice could flow in directions contrary to those that existed under more limited ice cover. One such clear example exists at NN 059465 where a small corrie faces to the ESE(Figure 11.1, corrie 112). The bedrock just to the E of the corrie comprises frost-riven Meall Odhar granite that lies outside the inferred limit of the Loch Lomond Advance. Good numbers of crescentic fractures preserved on flat rock surfaces and concave to the NW(Table 3.1) demonstrate ice-flow in that direction toward the corrie. A few inferred reversed crescentic fractures that are concave to the SE exist at the same locality, but since a comprehensive range of evidence(i.e.friction cracks,erratics and roches moutonnées) over a wider area supports an ice-flow direction toward the W to NW these are regarded as true reversed crescentic fractures.

CHAPTER 4

TRIMLINES : THEORETICAL CONSIDERATIONS, METHODS AND PROBLEMS

4.1 Introduction

Various forms of glacial evidence, other than end and lateral moraines, have long been used to delimit the ice-margins of former glaciers in many glaciated areas in the world(Flint,1971; Embleton and King,1975; Bowen,1978). Evidence that has been used extensively includes the limit of thick drift, hummocky drift,fluvio-glacial landforms and erratics, as described in chapter 2. In contrast little attention has been paid to contrasts in bedrock surfaces that can be used to delimit the margins of former glaciers. This is reflected in the fact that recent major reviews of glacial geomorphology paid scant attention to such contrasts in bedrock surfaces. Flint(1971) only devoted half a page to glacier trimlines while Sugden and John (1976) did not refer to trimlines at all; Embleton and King(1975) referred only briefly to trimlines, but used the term in a different sense from that used by Flint. Thus in addition to a paucity of information on trimlines problems of terminology exist.

Embleton and King(1975, p.433) refer to the sharp contrast to be seen in the vegetation cover of a valley side,caused by the recent thinning of a glacier, as a trimline. For example they stated that "The trimline separates an area of bare moraine below from vegetated moraine above". Flint(1971), however, used the term to denote where changes in the appearance of the bedrock occurred, below and above the former glacier margin. Thus it is implied that during an advance a glacier will 'trim' the bedrock by abrading and

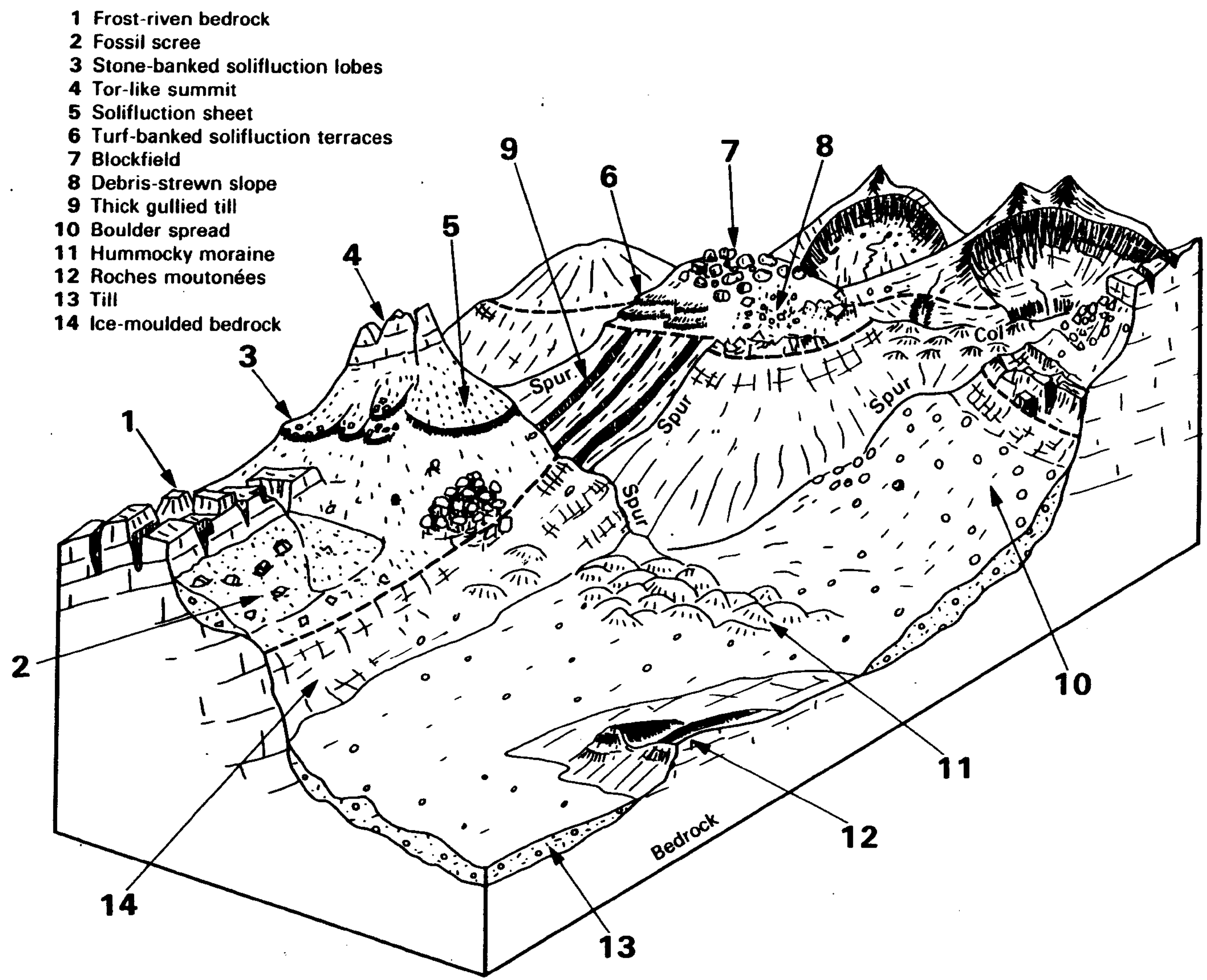


Figure 4.1 Idealised features used for identifying trimlines and other types of glacial limits. Postglacial features omitted

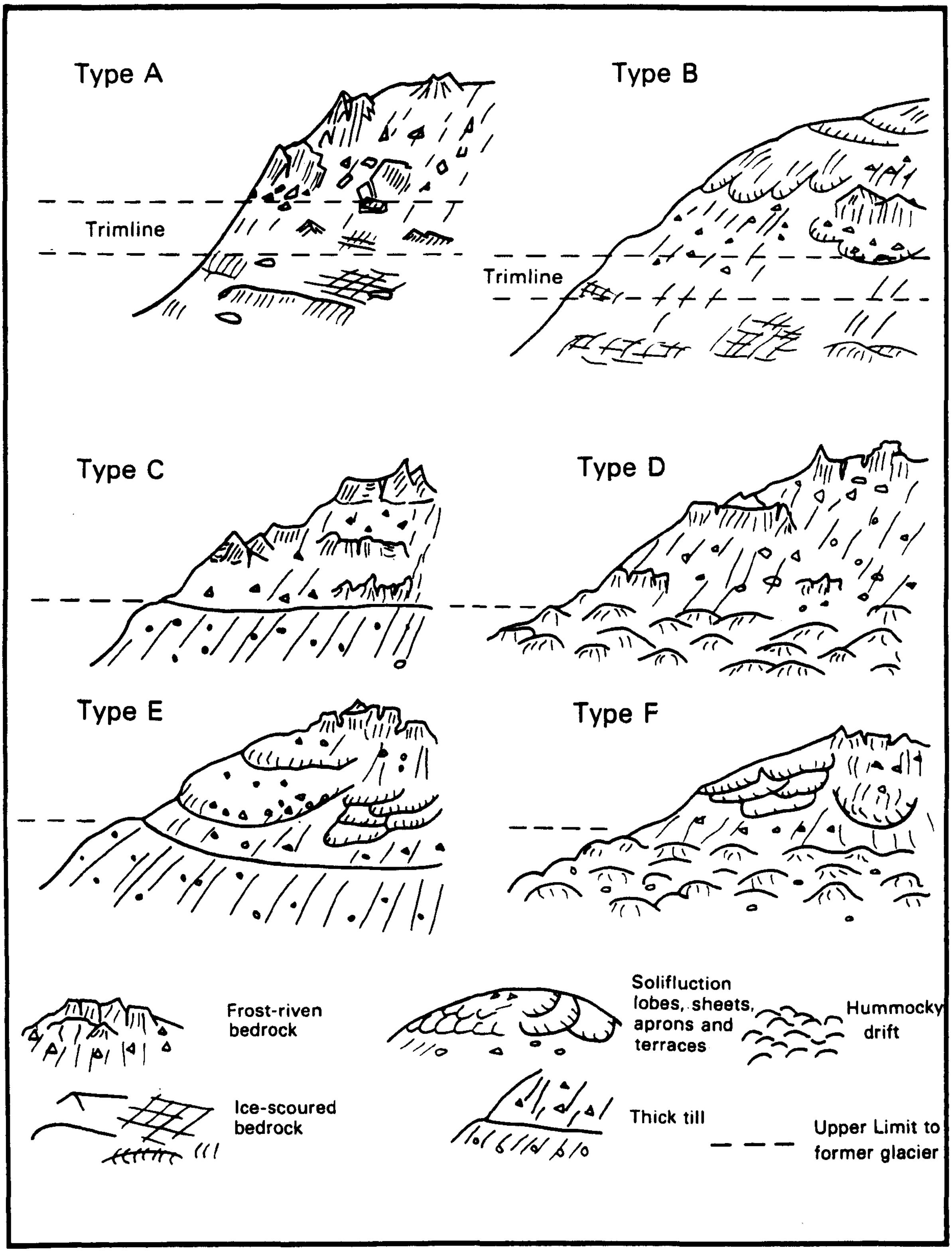


Figure 4.2 Idealised types of ice-limits

and smoothing rock surfaces, and by removing loose, angular blocks up to the level of its surface: above its surface the bedrock will remain unaffected by glacial processes, but will be affected by frost action. It is in this latter sense that the term will be used in this thesis.

One area where trimlines have been described and used to delimit former glacier margins is Labrador(Ives,1957,1958; Løken,1962). Ives(1957, p.72) stated, "... there is a marked contrast between surfaces above and below the 2,000 to 2,200 foot level. Below this level evidence of widespread glaciation is abundant, mainly in the form of erratics and perched blocks, glacial striations, ablation moraine, roches moutonnées and fluvioglacial deposits. Above this level evidence of glaciation is almost entirely lacking". He further described the frost-shattered, angular debris and blockfields that covered many of the mountain slopes above that altitude.

In general terms, therefore, trimlines can be identified on the collective basis of a range of periglacial and glacial features above and below the trimline respectively. The most important features are illustrated diagrammatically in Figure 4.1. The two types of trimlines(<u>sensu stricto</u>) described in this chapter together with the other types of ice-limits described in chapter 2 are portrayed in Figure 4.2. For detailed descriptions of these ice-limits see Appendix A.

4.2 <u>Trimline evidence and glacier variables: theoretical considerations</u>

In the literature, and as the term itself suggests, it is often implied that trimlines mark sharp boundaries between contrasting surfaces. This is rarely the case in the study area. Instead a transition zone between contrasting surfaces of bedrock, rather than a distinct line, is more usually observed. Moreover, it is to be expected from theoretical considerations(see below) that such zones will vary both in clarity and width. For convenience the term trimline is retained in this thesis, although it is used in its broadest sense to include 'trimlines' that are transition zones up to ca 60m in width.

Variations in the width and clarity of the transition zones will result from the influence of the following factors:

i) The presence and abundance of debris in the ice will be an important factor since clean ice will not abrade rock surfaces. Since the margins of the Advance glaciers(see chapter 7) contained abundant debris, the problems posed by clean ice are assumed to be unimportant. Much more important in terms of the clarity and width of the transition zone of ice-scoured rock will be the degree of concentration of the debris within the marginal zones of the glacier. High particle content will increase rates of abrasion so long as the velocity of the ice is not decreased substantially by the viscosity of the debris-rich zone in contact with the bedrock. If this happens then lodgement of debris can take place along the ice/rock interface and rates of abrasion will decrease rapidly(Sugden and John,1976). With ice velocity held constant low quantities of debris in the basal ice would result in weakly-abraded bedrock surfaces and the production of a wide transition zone between strongly-abraded bedrock and strongly frost-riven bedrock. Conversely, abundant debris is more likely to produce a narrower transition zone as a result of higher rates of abrasion.

ii) In the accumulation zone above the firn line the surface of the glacier is normally concave and the flow vector is directed in toward the glacier rather than outward and no lateral moraine is likely to form. Instead the margin of the glacier above the firn line will be characterised by slopes of snow and névé that are more likely to produce a diffuse trimline. Since glacier ice rather than snow produces smooth, abraded bedrock surfaces it is to be expected that the transition from snow to ice along the glacier margin will be reflected by a corresponding transition from non-abraded to abraded bedrock. The width of the zone in turn will relate to spatial and temporal variations in the depth of the snow and névé.

iii) In the ablation zone, where the glacier surface is normally convex, the flow vector is directed toward the margin of the glacier and debris may accumulate. Variations in amounts of ablation from year to year will result in the margin of the glacier occupying different positions on the valley side. Such variations in the glacier margin are likely to produce bedrock transition zones that will differ in clarity and width.

iv) The thickness of the ice is an important variable since the ice pressures exerted on any rock surface,ignoring pore water pressures, will increase with ice thickness(Boulton,1974). Boulton has suggested that with increasing ice thickness abrasion will be enhanced, until a point is reached where increasing friction will retard the movement of the rock particle, thus reducing the rate of abrasion. It follows that along the margin of a glacier abrasion will be at a minimum near the surface, but will increase with depth. Such variations in abrasion will be reflected by the bedrock surfaces in the transition zone becoming less well-abraded in an up-slope direction on the valley side.

v) Rates of abrasion of the bedrock can also be a function of glacier velocity because the effect of higher velocities of basal sliding is to increase the number of rock particles dragged over the bedrock in a given time. Glacier velocities vary both transversely and longitudinally as described below:

a) Studies of transverse velocity profiles of glaciers(Meier, 1960; Raymond,1971) show the greatest velocities in the centre and a four- or five-fold decrease within 50 metres of the glacier margin. This demonstrates that basal sliding velocities are least in the ice-rock interface zone along the glacier margin, but increase with depth toward the median line of the glacier.

b) It is well known that in a trough of uniform depth and width maximum velocities occur at the firn line and decrease both toward the snout and the source area of the glacier. Thus, with all other factors being equal, it would be anticipated that the clearest trim-lines should be located in the vicinity of the firn line at the time when the mass balance of the glacier reaches a state of equilibrium. Since glaciers do not flow in channels of uniform slope, width or depth this simple pattern of ice-flow can be over-ridden by other factors. Variations in the slope of the valley floor can create extending and compressing flow and hence variations in the velocity of the glacier. Similar variations in ice velocities can be induced by other topographic factors. Widening of a valley can create divergent flow of ice so reducing the longitudinal velocity of the glacier. Conversely, convergent flow of ice at narrow constrictions in a valley can lead to greater longitudinal velocities. Therefore, by implication rates of abrasion can decrease or increase depending on whether divergent or convergent flow of ice is taking place.

Since basal sliding velocities decrease toward the lateral margin of a glacier it follows that the intensity of ice-abraded bedrock will also decrease in the same direction and help in the creation of a transition zone with varying widths. Additionally by implication the widest transition zones are likely to be located where rates of abrasion are reduced by compressive or divergent flows of ice. Conversely, narrower transition zones are likely in locations characterised by extending or convergent flows of ice.

vi) The degree of glacial abrasion of the bedrock, and hence the width and clarity of the trimline, will reflect the length of time during which the glacier remained at equilibrium. If a glacier remained in equilibrium for a long period of time a relatively well-defined upper limit to well-abraded rock surfaces would be expected(although studies of present-day glaciers suggest that many glaciers rarely achieve a steady-state situation for any appreciable length of time). Wide transition zones will occur if the surface level of the glacier fluctuates considerably whereas a narrow transition zone will occur if the glacier surface does not fluctuate.

Taking into account the preceding points, and ignoring complications produced by different types of bedrock(see next section), it is to be expected that transition zones(trimlines) will vary considerably in clarity and width. In theoretical terms the clearest trimlines should occur close to the firn line of very active glaciers, that remain in equilibrium for some time and that carry a great deal of debris along their margins. This will be particularly true if the firn line of the glacier coincides with a narrow constriction in the valley. Conversely, the most diffuse trimlines should occur toward the snout and source areas, especially if these areas widen out. To what extent these assumptions match up with the fieldwork evidence will be discussed in section 4.8.

4.3 Trimlines and geological variables

The clarity or 'sharpness' of the trimline will also depend on the response of different types of rock to glacial and periglacial processes. Visual observations of numerous rock surfaces in the study area show that rock composition, texture and structure are major factors in the qualitative assessment of the degree of ice-moulding or frost-riving that has taken place(p.47).

Figure 2.2 demonstrates the considerable differences that exist between different rock types in their resistance to frost-riving. As a crude approximation it is suspected that their resistance to chemical weathering processes is the reverse to that shown in Figure 2.2(Gilluly et al,1968; Pitty,1971; Blatt et al,1972), but this is more difficult to confirm from visual observations and from the limited evidence obtained in the field.

The following criteria, based on visual observations, were used in distinguishing trimlines inferred from contrasts in bedrock surfaces. These were:

i) degree of angularity or rounding of edges of rock outcrops;

ii) number and depth of open joints;

iii) number, size and depth of weathering pits and grooves; and

iv) amount of granular disintegration(only applicable to some rock types).

A sharp increase in the angularity of bedrock edges, depths and numbers of open joints and weathering pits and in the amount of products of granular disintegration was taken as representing the upper limit to the former glacier surface. This evidence was used in conjunction with other forms of evidence shown in Figure 4.2.and described in detail in chapter 2.

Chemical weathering is considered to be of some relevance to the correct identification of trimlines, since the trimline technique partly relies on the differences in the angularity of the edges of rock outcrops, above and below the inferred trimline. For example many of the massive granites(these are strictly speaking more often diorites or granodiorites) show rounded joint-block edges <u>both</u> above and below the inferred trimline. Whereas the rounding of the edges of rock outcrops above the trimline are mainly inferred to be the result of chemical weathering(generally indicated by considerable quatities of the products of granular disintegration), the rounded edges of rock outcrops below the trimline are inferred to be primarily the product of glacial abrasion. Thus this particular criterion has to be used with great caution and only in conjunction with the other forms of evidence, such as amounts of frost-riven debris and depths of open joints, when mapping in granite bedrock areas.

In contrast rock types rich in quartz, such as rhyolites, quartzites and some of the acid intrusive rocks, generally have highly angular, joint-block edges above the inferred trimlines. In some places the edges of rock outcrops are still angular, even below the inferred trimline, where glacial abrasion processes were inadequate to round off the edges, or where blocks had been plucked away from the outcrops.

In summary the surfaces of many coarse-grained 'granites' are generally the most difficult to interpret in terms of contrasting bedrock evidence, above and below the inferred trimline, because although they may be easily weathered by chemical processes they are generally very resistant to freeze-thaw processes. These factors minimise the contrasts that are often very apparent on other rock types such as the quartzites.

4.4 Trimlines related to the Loch Lomond Advance: previous studies in Scotland

It is well over 100 years since it was first proposed that the recession of the last ice-sheet was succeeded by an advance of valley glaciers(Chambers,1853,1855); a fact supported by the work of the Officers of the Geological Survey later in the century(e.g. Bailey et al, 1960). They recognised (p.12) that "the summits of the high ridges were certainly exposed to frost action during late glacial times when the valley bottoms were still occupied by glaciers". Trimlines, however, were not referred to nor were any mapped, although numbers of ice-limits based on moraines were suggested. Only within the last 15 years have trimlines(but not referred to by that term) been used to delimit some of the margins of the Loch Lomond Advance glaciers. Examples are given below.

Sissons(1977a, p.227) delimited the margin of part of the former Moriston glacier on the basis of trimline criteria: " repeated traversing of the critical zone showed that on the higher ground the extensive rock outcrops are in places disrupted by severe frost action into large angular blocks, while local accumulations of such material occur at the base of some slopes. At lower altitudes the glaciated surfaces are relatively fresh, frost action being mainly confined to granular disintegration and small-scale wedging. The limit of these two types of terrain was located in a zone, rising in altitude westwards, that occupies a height interval of some 10 - 15m".

The main area where trimlines have been used is in the NW Highlands where they were used to partly delimit six of the larger glaciers(Sissons,1977c). Nevertheless, this represents only a small proportion of the total number of 70 glaciers that were delimited. In the Gaick plateau area the only trimlines to be mapped were in the vicinity of the nunataks Beinn Bearg and Beinn Bhreac (Sissons,1974a). On Rhum Ballantyne and Wain-Hobson(1980) defined part of the margin of one glacier on the basis of contrasting rock surfaces. As far as is known no trimlines have been mapped elsewhere in Scotland, outside the study area.

The present study, therefore, represents a major departure from previous fieldwork studies of the former Loch Lomond Advance glaciers since trimlines form a very important element in the reconstruction of the main ice mass described later in this thesis (chapters 5 and 6).

4.5 Mapping trimlines in the field

Although an extensive search has been made in the literature,no published accounts have been found that deal in detail with the methodology and problems of mapping trimlines systematically over a wide area (e.g. Ives,1957,1958; Løken,1962: Flint,1971; Porter,1975; Sollid and Sorbel,1979; Porter and Orombelli,1982; Meierding,1982).

Thus when the present author began a detailed mapping programme in 1977 very little information was available on the methodology and problems of mapping trimlines. From fieldwork experience obtained over the last seven years it is suggested that the following guidelines need to be adhered to when mapping the trimlines of a former large ice mass or valley glacier. Some of the guidelines have already been outlined in Thorp(1981b). A complete list of such guidelines is given below:

i) As many trimlines as possible should be mapped in a given area. A minimum number of 20 - 30 is suggested.

ii) Spurs should be especially selected to ascend a mountain since:

a) the bedrock surface is generally less obscured by scree,till or hillwash deposits;

b) the bedrock has generally suffered less dissection by postglacial streams in comparison with many valley-side slopes;

c) the spurs can often provide a continuum of glacially scoured bedforms from near the valley floor to the inferred trimline as a result of the scouring action of the ice across the spur; and

d) the spurs are sufficiently numerous to provide an adequate number of inferred trimline values for reconstruction purposes and for cross-checking for fieldwork errors.

iii) Both sides of a valley should be mapped to act as a cross-check on the trimline values derived from the spurs.

iv) The evidence on both low-level and high-level cols needs to be obtained, both to supplement the trimline values and to act as an independant check on such values.

v) Mountain summits below the inferred trimline values need to be examined for glacial or periglacial evidence. Such evidence can be used to support or question the validity of the trimline values.

vi) Ideally bedrock surfaces should be measured for micro-relief features, such as the depth of open joints(Ballantyne,1982) and the degree of rounding/angularity of the bedrock edges. In realistic terms this is only possible, given the limitations of time, within a small area and hence was not carried out systematically as part of this study.

vii) Ice-marginal deposits such as end and lateral moraines and the upper limit of thick till need to be mapped, to act as a check on the accuracy of the trimlines. For example, trimlines 223, 229 and 230(Figure 5.4 and Appendix A) were found to be underestimates of the glacier surface by between 20 and 40m, in comparison with the altitude of lateral moraines,eventually mapped farther to the E.

viii) When mapping along a spur or valley side care must be taken to distinguish between plucked and abraded bedrock surfaces, both below and above the inferred trimline. Observations in an up-glacier direction often reveal rough,plucked rock surfaces, whilst ice-smoothed rock surfaces are most noticeable in a down-valley direction. The former could result in the recognition of a minmal ice-limit whilst the latter could give a spurious ice-limit, as a result of the sporadic preservation of ice-moulded bedrock above the inferred trimline that was produced by earlier glacial activity.

xi) Glacial markings need to be recorded since they can provide supplementary evidence in situations where the bedrock is the same above and below the inferred trimline(Thorp,1981), although this type of evidence becomes less significant at altitudes below <u>ca</u> 500m O.D.(see p.205).

4.6 Calculating the altitude of trimlines and remarks on their accuracy

Since trimlines actually comprise a zone that varies in width the method for calculating the altitude of the trimline is as follows. Firstly, the highest limit of extensive outcrops of ice-smoothed bedrock is determined. Secondly, the lowest altitude of extensively frost-riven bedrock is similarly determined. A mean altitude can then be calculated from the minimum and maximum values described above(Appendix A). The reason for such a calculation is that in the field it is rarely possible to be confident that either of the limits described above actually represents the true limit of the former glacier surface. A mean value, therefore, is more likely to approximate to the true altitude of the surface of the former glacier than either the maximum or minimum limits.

At some localities it was not possible to map a trimline because of smooth grass-covered slopes or steep or vertical rock faces occurring at the critical altitude. Thus in these cases it is only possible to give a maximum altitude of strong glacial evidence and a minimum altitude of clear periglacial evidence. Nevertheless, these values are included in Appendix A as maximum/minimum altitudes since they do at least provide a check, even if only very approximate, on the ice-limits calculated from the trimlines.

Initially it was concluded that most trimlines were accurate to within $\pm$ 30m(Thorp,1978,1981b). Further field mapping has since indicated that errors of a greater magnitude are possible. This particularly applies to (a) the area consisting of Starav granite, and (b) the areas where the glaciers descended below altitudes of ca 400 - 500m O.D., as in parts of glens Coe,Creran and Etive and at the western end of Loch Leven.

Due to the apparent resistance of Starav granite to frost-riving(p. 47) a number of trimlines on this rock type are based on minimal changes in bedrock surfaces. For example, the trimlines on spurs 116,163 and 164(Figure 4.4 and Appendix A) are based on (a) the bedrock edges becoming more angular, although there are problems associated with the use of this particular criterion(section 4.3); (b) bedrock surfaces becoming more weathered with deeper weathering pits and grooves(cf Watts,1981,1983 for weathering under Arctic conditions); (c) a greater amount of granular disintegration (often indicated by a more extensive cover of mosses and lichens); and (d) deeper and wider joints. Yet numbers of frost-riven blocks are often small and apparent ice-moulded rock surfaces are common above the inferred trimline. Hence reservations are still held about the accuracy of the data in these localities.

Along spurs 113,114 and 125 it was not possible to distinguish even the minimal differences in bedrock surfaces outlined above and, therefore, no trimline value is given in Appendix A. For example, spur 125 comprises excellent ice-smoothed Starav granite to altitudes exceeding 800m O.D. This is considered to be spurious evidence since trimlines in the immediate vicinity on other rock types, such as quartzite,schist and Cruachan granite, are based on consistent evidence of former glacier surface altitudes within the range of 500 - 600m O.D.

At altitudes of less than 400 - 500m O.D. the evidence based on contrasting bedrock surfaces becomes much more difficult to interpret. This is due to two factors additional to those outlined in section 4.3. Firstly, superficial deposits begin to occupy a greater proportion of the spur and valley slopes below the inferred trimlines. Secondly, frost-riving of the bedrock is generally less severe than at higher altitudes and bedrock contrasts are less marked as a result.

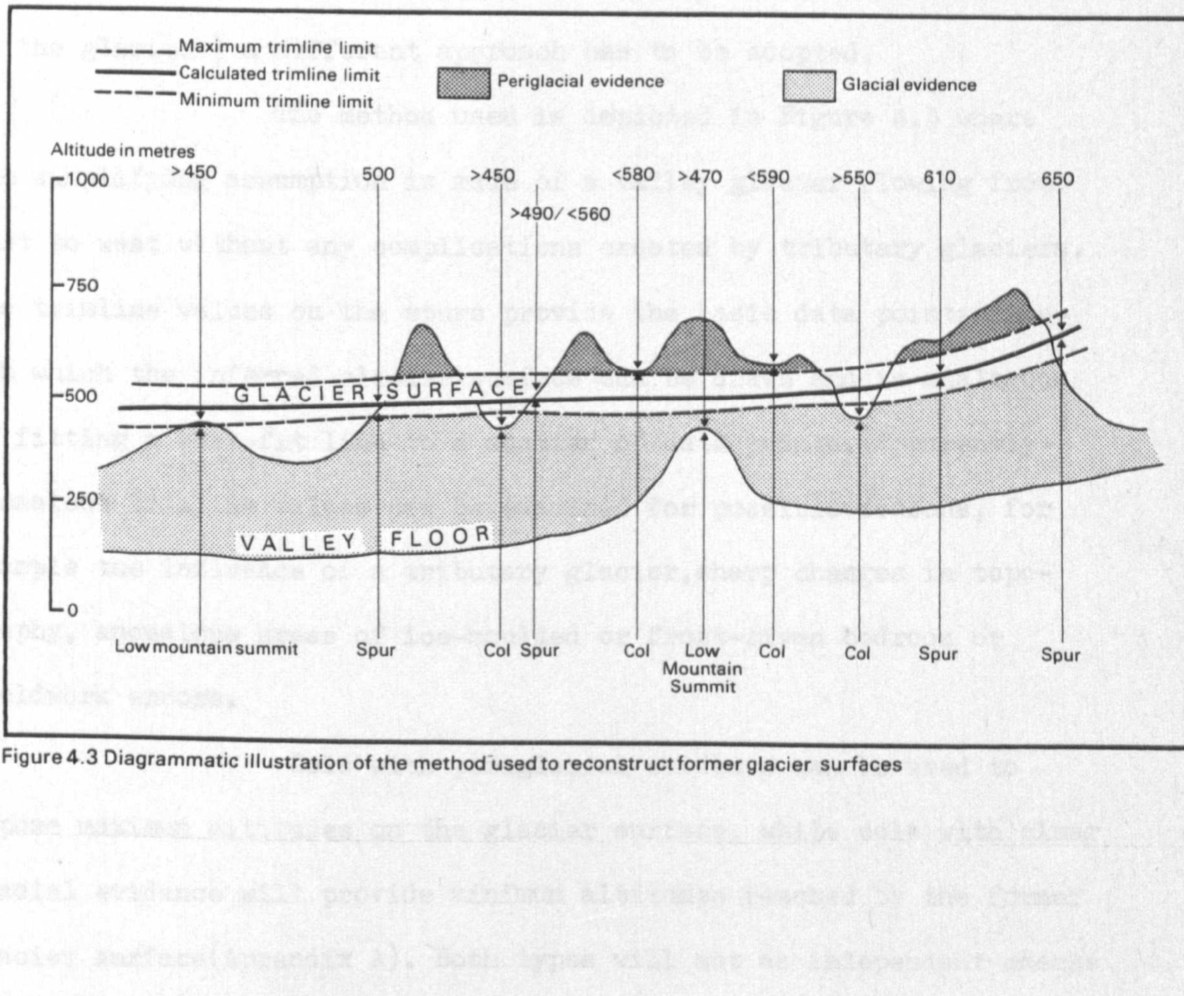


Figure 4.3 Diagrammatic illustration of the method used to reconstruct former glacier surfaces

4.7 Reconstructing former glacier surfaces using trimlines

The methods for reconstructing former glacier surfaces using end, lateral and hummocky moraines, the upper limit of thick till and the lower limit of periglacial evidence have been outlined by Sissons(1974c). Such methods are applicable to the lower ablating segments of the outlet glaciers shown in Figure 6.1, but in the Western Mountain zone where end and lateral moraines are largely absent(as would be expected in the main accumulation areas of the glaciers) a different approach has to be adopted.

The method used is depicted in Figure 4.3 where the simplifying assumption is made of a valley glacier flowing from east to west without any complications created by tributary glaciers. The trimline values on the spurs provide the basic data points through which the inferred glacier surface can be drawn and is analogous to fitting a best-fit line to a scatter of data points. Apparently-anomalous trimline values can be examined for possible reasons, for example the influence of a tributary glacier,sharp changes in topography, anomalous areas of ice-moulded or frost-riven bedrock or fieldwork errors.

Cols with periglacial evidence can be used to impose maximum altitudes on the glacier surface, while cols with clear glacial evidence will provide minimum altitudes reached by the former glacier surface(Appendix A). Both types will act as independent checks on the trimline values(Figure 4.3). Hill and mountain summits with clear glacial evidence can help to corroborate the minimum altitude reached by the glacier. Problems of interpreting the evidence,however, can arise where the mountain summit is perhaps within 20 - 60m of the inferred surface of the glacier(p. 127).

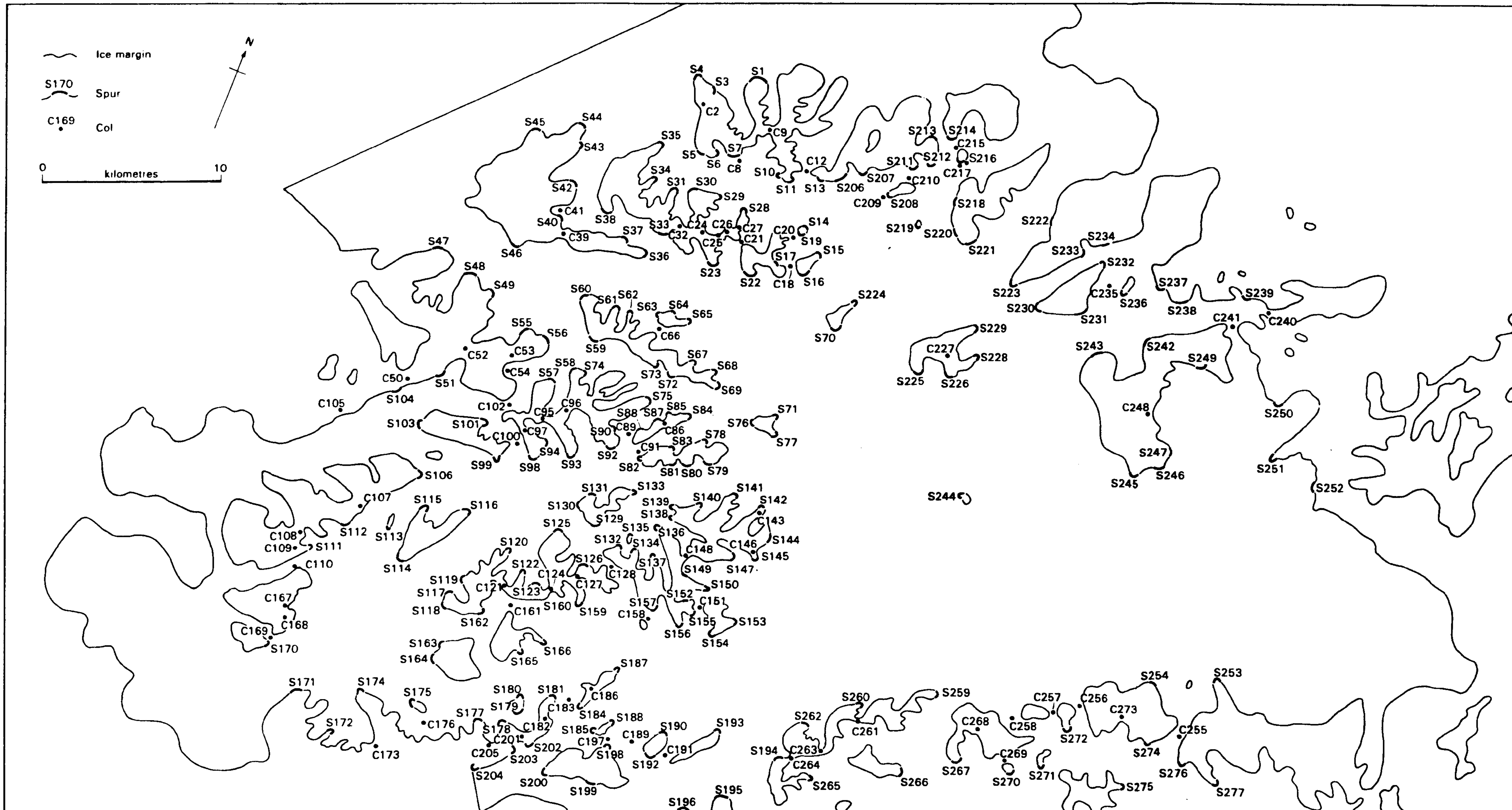


Figure 4.4 Location of cols and spurs used to reconstruct the former glaciers listed in Appendix A

Source areas of ice, such as tributary valleys and corries, can be related to the reconstructed valley glaciers and reconstruction is based on the reasonable assumption that glacier surfaces sloped upward toward these source areas.

Finally, the topography and the spatial distribution of forms of evidence such as striae,friction cracks,roches moutonnées,meltwater channels,hummocky moraines,thick till and boulder spreads, can be compared with the form of the glacier surface calculated from the trimline values and minor adjustments made where necessary.

Trimlines and other types of ice-marginal evidence(Figure 4.4) on over 200 spurs and the evidence on 73 cols (Appendix A) were used to reconstruct much of the ice mass shown in Figure 6.1 using the methods described above.

4.8 Conclusions

The term trimline is actually a misnomer since virtually all of the 'trimlines' mapped by the writer comprise a zone. The zone varies considerably in width and clarity depending as it does on a number of variables. The most important are inferred to be variations in the lithological and structural characteristics of rock type and the varying severity of periglacial activity,especially with altitude and exposure to frost. Variations in the amount of debris entrained by the glacier and variations in glacier velocity are considered to be of less importance.

Since trimlines contain an inbuilt, but varying magnitude of error they cannot match the accuracy of lateral moraines and unquestionably errors of interpretation of the field evidence are more likely to occur, Errors are most likely below altitudes of _ca_ 500m O.D., especially if the bedrock comprises a coarse-grained granite or massive schist with sparse jointing. Since some degree of error is inevitable it is essential that a large number of trimlines is mapped over relatively short distances, particularly if other forms of evidence such as lateral moraines or a distinct upper limit to thick till are poorly represented.

How long the former glaciers occupied their maximum position is unknown at the present time nor is it known whether surges occurred in any of the glaciers. Both of these factors could have influenced the width and clarity of the trimlines.

CHAPTER 5

THE LIMITS OF THE FORMER GLACIERS

5.1 Introduction

This chapter presents the detailed evidence mapped in the field which collectively enables ice-limits and ice-flow directions to be reconstructed for the former glaciers.

The evidence for reconstructing the maximum extent of the former glaciers is presented in Figures 5.2 to 5.5. The study area has been subdivided into four areas, mainly to enable the inclusion of detailed periglacial and glacial evidence that would otherwise be difficult to depict clearly at a smaller scale.

The corries are numbered c1 to c271 on Figures 5.2 to 5.5 and will be referred to by number in the text. The location of cols and spurs that have been used to determine ice-limits are also depicted in Figures 5.2 to 5.5 and these will be prefixed by the letters C and S where not referred to directly as cols and spurs. Additional forms of evidence and critical areas will be referenced by means of the numbers 1 to 103 as shown on Figures 5.2 to 5.5.

Since the bulk of the evidence is presented in map form much of the discussion will be centred on selected areas, with critical or controversial evidence, rather than on repetitive detailed descriptions of all the mapped field evidence.

Detailed descriptions of much of the field evidence and the ice-limits in Glen Nevis, the Lairigmor valley, Glen Coe and the Loch Leven area have already been given in Thorp

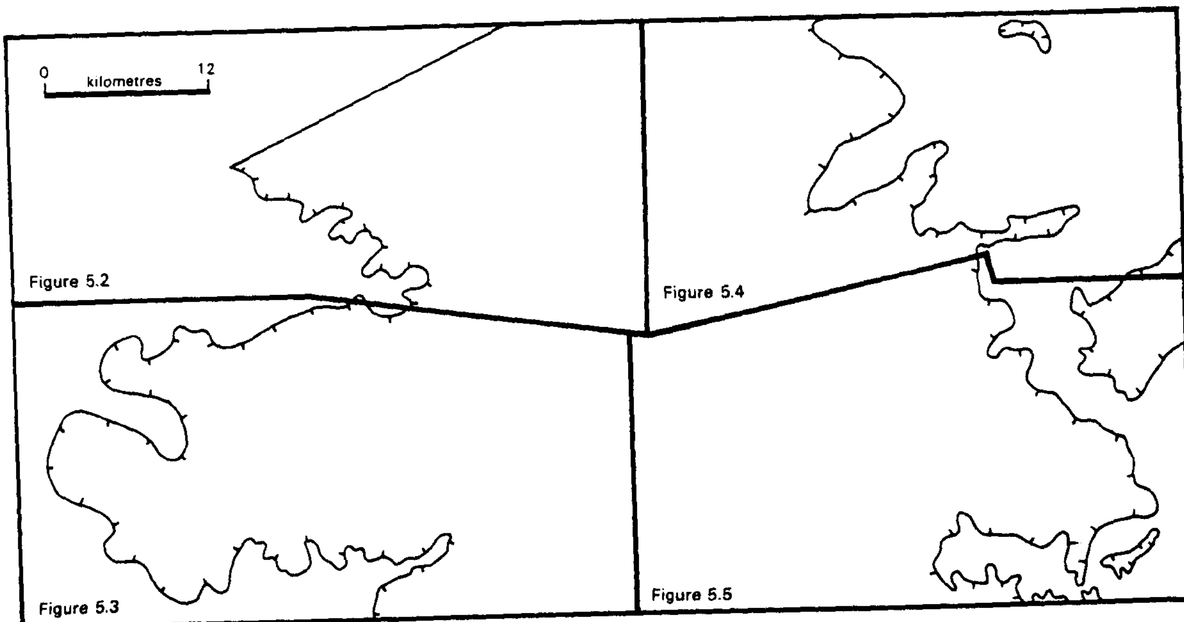


Figure 5.1 Areas of the main ice mass covered by detailed maps 5.2 to 5.5

Figure 5.2 Detailed field evidence and Loch lomond **Advance** limits for the N.W. quadrant of the study area

(1978,1981b). Descriptions in this thesis will incorporate this work and for some areas reference will be made to additional evidence mapped more recently. Corrections to some of the ice-limits have been made and will be discussed in section 5.2.

5.2 Nevis glacier

Field evidence shows that the Nevis glacier received ice from 22 corries and 8 tributary valleys with the largest volume of ice flowing westwards toward Loch Linnhe, with smaller quantities of ice flowing eastward toward Loch Treig. An ice-shed at the head of Glen Nevis divided these two streams of ice.

Hummocky moraine is not generally abundant except in lower Glen Nevis, although tills of varying thicknesses occur intermittently on the floor and slopes of the glen below the ice-limits.

Type A trimlines on S14,S15,S17 and S19 in the vicinity of 10 allows the surface of the former glacier to be placed at ca 800 - 850m O.D. That cols 18 and 20 at altitudes of ca 740m O.D. were overtopped by ice is suggested by smooth,ice-scoured bedrock and in some locations by the presence of numerous friction cracks and striae. Abundant,well-rounded erratic boulders of red porphyrite provide a striking contrast to the white quartzite bedrock on which they rest. On col 20 they are largely derived from corrie 49 that has been eroded along the line of a very large porphyrite dyke whilst on C18 they appear to be plucked out of dykes in the floor and walls of the col itself. The sub-radial spatial pattern of the striae and friction cracks in the vicinity of 10 demonstrates that this group of mountains was important for feeding ice into Glen Nevis and the Loch Treig and Loch Leven areas.

A former glacier surface declining from _ca_ 800m on S19 down to _ca_ 600m O.D. on S5 is inferred from the contrasting glacial and periglacial evidence with the most rapid decline of 75 to 100m/km taking place between S5 and S7. The lower limit of severely frost-riven granite bedrock and thick scree at these two localities and elsewhere impose strong limits on the maximum altitude reached by the former glacier surface.

Thick,dissected till in the tributary valleys at 3,4,5,8 and 9,and striae,friction cracks and roches moutonnées collectively indicate occupance of the valleys by ice that flowed N to join the main Nevis glacier. Cols 24,25,26 and 32 at altitudes of 761 to 870m O.D. provide strong controls in the form of periglacial evidence on the possible maximum heights reached by the surfaces of the tributary glaciers. However, C27 at 857m O.D. provides no such clear evidence and,although it is shown on Figure 5.2 as being outside the ice-limit,the possibility that ice may have flowed across the col from corrie 42 cannot be discounted.

Two corries in the Mamore Forest Range appear to have lacked corrie glaciers that might have existed contemporaneously with the Nevis glacier. In corrie 25 thick deposits of angular rock fragments cover the corrie floor and sidewalls and no visible glacially-smoothed bedrock could be located. This evidence has been interpreted as indicating a lack of ice,although the lower part of the corrie does contain low mounds of till.

Corries 35 and 36 provide examples of totally contrasting evidence in adjacent corries. Both corries are incised into quartzite but whereas corrie 36 is characterised by extensive areas of ice-smoothed quartzite whose surfaces are scored by great numbers of friction cracks, corrie 35 contains large numbers of angular blocks of quartzite resting on severely,frost-riven quartzite bedrock. Furthermore no friction cracks were located following a thorough search except for a few on one small quartzite surface. Thus this contrasting evidence is interpreted as indicating occupance by ice in one corrie but not in the other.

Fine examples of highly glacially-smoothed bedrock and roches moutonnées occur at a number of localities in Glen Nevis, as for example W of 7, N of 8 and 9, and especially near to 6. The ridge E of 6 is strongly ice-moulded to its summit(698m O.D.) and characterised by large streamlined bedforms and roches moutonnées that indicate a strong flow of ice to the SW from area 7. The southern limit of Ben Nevis granite erratics in the area(see Figure 7.3), striae and friction cracks and the restriction of thick till to the E side of the ridge give added support to a SW movement ice from the Ben Nevis-Aonach Beag massif toward the Nevis trunk glacier.

The evidence on S30,S34 and S35 provides only crude controls on the maximum height reached by the Nevis glacier (see Appendix A). On S35 the strong contrast between severely frost-riven bedrock and soliflucted debris above 700m O.D. and strongly ice-moulded schist bedrock below that altitude originally led the writer to postulate a trimline value of 700m O.D.(Thorp,1978). Subsequent mapping in the area and particularly the discovery of strong periglacial evidence down to ca 600m along S5 and S31 has led to a correction of that view and the ice-limit is now placed at about 550m O.D. on S35. The ice-moulded schist above this limit is now regarded by the writer as representing anomalous evidence created by the ice-sheet.

5.3 Leven and Lairigmor glaciers

The main source areas for the Leven glacier were i) Rannoch Moor and the Blackwater valley; ii) eight corries mainly on the S side of the Mamore Forest Range; iii) eight corries on the N side of the Aonach Eagach Range; iv)four corries in locality 16(Figure 5.2).

Extensive areas of Rannoch Moor and the Blackwater valley are covered by hummocky moraine and till. Large hummocks and ridges are especially prominent at 24 where ice flowing NW from the mountains at the head of Glen Coe left thick quantities of till and numerous mounds on the up-slope sides of ridges and along the floors of intervening valleys. The convergent pattern of striae and roches moutonnées across the area is remarkable for the strong convergence of ice toward the head of Loch Leven it displays.

Spur trimlines(S16,S22,S23,S64,S65,S67,S68,S69 and S71) on the surrounding mountains provide evidence for a glacier surface declining from over 700m O.D. at the eastern end of the Mamore Forest and Aonach Eagach ranges to ca 610m O.D. on S64 near the head of Loch Leven. The only strongly anomalous ice-limit was mapped on S70 where ice-moulded schists and flags occur to only ca 590m O.D; above that altitude the bedrock is increasingly frost-shattered and accompanied by thick scree. Thus there exists a discrepancy of ca 60m between this trimline and the surface of the former Leven glacier calculated from the remaining trimline values E of Loch Leven. A possible explanation for this discrepancy will be discussed in section 5.14.

The pattern exhibited by striae, friction cracks and roches moutonnées near 23 shows that ice flowed from corries 56, 57 and 58 to become confluent with the Leven glacier. Some of the ice from corrie 56 spilled westwards over C66. This diffluence is supported by hummocky moraine on the floor of the col and by friction cracks on ice-moulded bedrock up to altitudes of ca 630m O.D. A rapid change to frost-riven bedrock, thick scree and soliflucted debris occurs above this altitude.

Between 22 and the head of Loch Leven extensive areas of ice-moulded quartzites and schists on the slope above the loch testify to the occupance of the trough by a glacier with a thickness of ca 590m. Trimlines on S36, S60, S61, S62, S63 and S64 reflect the often abrupt change to bedrock strongly affected by former periglacial conditions and record the steady decline of the glacier surface from ca 610m O.D. on S64 down to ca 420m O.D. on S60.

Glacial diffluence across the col at 12 (Figure 5.2) is supported by the pattern of striae and friction cracks to the S and W and by the distribution of ice-moulded quartzite on the slopes above the col. The descent of the glacier surface westwards along the Lairigmor valley is recorded by the evidence and derived trimline values on S36(530m), S37(520m), S38(480m), S40(465m) and S42 (300/400m), although the contrasts in bedrock surfaces diminish at the lower altitudes in the W (p.84). Nevertheless, occupance of the Lairigmor valley by a glacier is strongly supported by numerous hummocky moraines with many perched granite boulders and by thick gullied till as far as 13. Small hummocky moraines on the valley slope S of 13 and on the low col to the SW and not beyond suggest that a lobe of ice finally terminated here.

The continuing westward decline in the altitude of the surface of the former Leven glacier is recorded by ice-limit values of 310 - 350m O.D. on S46,S48 and S49.

Ice-moulded bedrock and only minimal scree in the four corries(61 - 64) at 16 and hummocky moraines and thick till in the valley below the corries indicate that ice flowed N to become confluent with the Leven glacier. In contrast to this evidence corrie 65 is infilled with large quantities of frost-riven quartzite and no ice-moulded bedrock is to be seen. This implies that the corrie lacked a glacier and an interpolated ice margin is drawn across the lower part of the corrie between S48 and S49.

Farther W there is no clear evidence for an ice-limit until Kentallen(15) is reached and the evidence here will be discussed in section 5.5.

5.4 Coe glacier

The Coe outlet glacier was nourished by ice from five corries(60,99,102,103 and 104) at the head of Glen Coe and five corries(91,92,93,94 and 96) on the S side of the glen(Figure 5.2). Only one small corrie glacier(c69) supplied ice to the Coe glacier from the N side of the glen. Discussion of the evidence and ice-limits at the head of Glen Coe will be deferred until sections 5.7 and 5.10.

Superb ice-moulded rhyolites and andesites characterise the rock bar at 21 and on parts of the valley slopes farther E toward the head of the glen. Similar surfaces of ice-moulded volcanic and granite bedrock are to be seen in the three tributary valleys SW of 21.

The upper limit of the Coe glacier is defined by trimline values of 620 - 670m O.D.(S84,S85,S78,S141,S76) at the head of Glen Coe(Figure 5.5). A rapid descent of the glacier surface from ca 650m to ca 550m O.D. in the vicinity of the rock bar followed by a more gentle decline between spurs 73 and 59 is recorded by the spur trimline values(Appendix A).

The two corries W of 20(91 and 92) display contradictory evidence. The corrie on the northeast side of Bidean nam Bian(92) is incised into resistant volcanic bedrock that in places is highly polished and mamillated. The other corrie contains large quantities of frost-riven fault-intrusion granite debris and apparently lacks any clear ice-moulding. Since both corries are similar in shape,depth and orientation then the assumption is made that they both contained glaciers during the Loch Lomond Advance and that the anomalous frost-shattering relates to frost-riving of the weakly-resistant fault-intrusion granite during the postglacial period..

Thick till and hummocky moraines characterise the lower part of the glen(N of 20). For example at 20 till is exposed in the banks of a stream to minimum thicknesses of 7m. The stream has cut down through the till to form a fine series of steeply sloping terraces. Thick,gullied till is also plastered on the valley slope S of 22, whilst at 19 the tributary valley is choked with as much as 40m of till. W of 19 great numbers of granite,volcanic and schist boulders litter the floor and sides of col 54 suggesting a possible ice-limit in the vicinity of the col: W of the col till is scanty and well-developed morainic mounds are absent.

At 17 a belt of hummocks studded with granite and volcanic boulders extends along the floor of Gleann an Fhiodh. An irregular ridge of thick till on the N side of the valley may represent an end moraine. An ice-limit is therefore drawn where the hummocks and thick till fade away, where it is inferred that the tongue of ice failed to make any further progress up the reverse slope.

No clear evidence for the presence of ice exists in the valley at 18 and ice-limits are tentatively drawn across the exit of the valley and on the S side of the col(C53) to the S.

5.5 Linnhe glacier

The Linnhe glacier received ice from three main sources: i) the mountains on the W side of Loch Linnhe outside the study area; ii) the mountains flanking Glen Nevis; and iii) Glen Coe and the Leven trough(Figure 5.2).

A study of the distribution of erratics in the Spean valley (NW of 1) by Peacock(1970b) confirmed earlier observations(Jamieson,1863; Wilson,1900),that ice from the mountains W of the Great Glen flowed eastward along Loch Eil and Glen Loy and crossed the Great Glen to impound the Glen Roy series of lakes. Mapping of Moinian erratics by the present writer confirms that ice from W of the Great Glen also crossed the Great Glen farther S and pushed 2km into lower Glen Nevis(Figure 7.3 and p.186). Ice flowing northward down Glen Nevis was forced to flow westward over the valley side SW of 2. Ice-moulded outcrops of bedrock streamlined to the SW along the summit of the valley side confirms this ice-flow direction. Farther to the SW the ice-moulded bedforms are complemented by numerous hummocky moraines mantling the floors of the small valleys leading down to Loch Linnhe.

In comparison with the large mass of ice that flowed into the Great Glen from the W,only relatively small quantities of ice were supplied from the corries and tributary valleys on the N side of the Ben Nevis Range(Sissons,1979b). In this area the ice-limits near 1 are based on the present writer's mapping; the limits of the three tributary glaciers shown farther E are those according to Sissons(1979b).

The steep corrie walls and crags carved out of volcanic bedrock on the N side of Ben Nevis nourished a glacier that flowed into the Great Glen. Clear evidence of this former glacier exists in the form of highly, ice-polished rock steps composed of granite or volcanic rock on the valley floor and below the corries. At some localities(e.g. NN167722, NN173714, NN166726 and NN167724) numbers of crescentic gouges and crescentic fractures on andesite or granite bedrock demonstrate ice flowing firstly northward out of the corries and then turning NW down the steep-sided trough facing into the Great Glen.

The broad col(C2) W of 1 that contains Lochan Meall an t-Suidhe has no clear evidence of occupance by ice; its smooth, peat-covered floor lacks hummocky moraine or any unequivocally fresh-looking, ice-smoothed bedrock. The trimline value of 580m O.D. for S3 is based on an abrupt change from smooth,ice-moulded granite bedrock to angular,frost-riven surfaces,abundant sub-angular blocks of granite and gelifluncted debris. Although the broad,rounded granite ridge of Meall an t-Suidhe(W of 1) has been moulded and shaped by ice many of the rock knobs above an altitude of ca 550-600m O.D. show signs of periglacial frost action in the form of deep, open joints and loose, frost-riven blocks of granite.(The frost-affected, rock knobs orientated in a WSW-ENE direction across the ridge are therefore interpreted as forming under ice-sheet conditions and not under the more limited conditions of ice cover described in this chapter).

To the E of 1 the mountain slope in the vicinity of NN 165745 is mantled by large numbers of granite and andesite boulders. Solifluction boulder lobes with risers of 1-3m occur on the upper slopes and in places contain great numbers of andesite boulders derived from the upper slopes of Ben Nevis. The lowest limit of the lobes rises from ca 580m to above 650m O.D. farther E along S1. This presents a problem in terms of ice-limits that will be discussed in chapter 7 (p. 189). A low ridge of till, containing mainly granite boulders with some schist and andesite boulders occurs on S1 at an altitude of ca 640m O.D. Below the ridge quartzite and schist boulders rest on the granite bedrock, especially in the vicinity of NN 169748. Farther E at NN 185757 thick gullied till mantles the slope below an altitude of ca 620-630m O.D. Sissons (1979b, Figures 1 and 4E) has interpreted the preceding evidence as demarcating the upper limit of the Spean glacier declining eastward from ca 640m O.D. at an average gradient of 15m/km. However, mapping by the present writer in the immediate area and outside the study area(unpublished) suggests that modifications to this interpretation are necessary.

Firstly, evidence provided by erratics can be equivocal when used to reconstruct former ice-flow directions and ice-limits(see previous discussion p. 40). Secondly, the interpretation of such forms of evidence often assumes that such erratics have not been redirected during a later phase of glaciation. The area

N of Ben Nevis is no exception as Peacock(1970b) has remarked: "Though the general picture of eastward movement of ice into and across the Great Glen is clear the details of distribution of the erratics are less easy to explain, particularly in the southern part of the area described"(area N of 1). Peacock noted(p.187) that although Moinian material occurred up to altitudes of ca 600m O.D. on the slope N of 1, erratics from the Ben Nevis complex occurred in the moraines along the Lundy valley(4km N of 1). Peacock (1970b, Figure 1, p.186) interpreted this evidence as indicating that local ice from the Ben Nevis Range was able to penetrate into the Lundy valley at some stage during the Loch Lomond Stadial, although he did not specify how this ice-flow direction occurred at the same time as a large mass of ice was moving eastward into Glen Spean.

The present writer interprets the distribution of the erratics differently from Sissons and Peacock. It is suggested that the Moinian material occurring up to altitudes of ca 600m O.D. on the northern flank of the Ben Nevis massif relates to ice-sheet conditions when ice flowing E across the Great Glen was forced to diverge to the NE and SW around the massif. That the massif was not overwhelmed by external ice is suggested by the distribution of volcanic erratics outside the inferred ice-limits of the Loch Lomond Advance(as for example the andesite erratics from Ben Nevis occurring high on the ridge to the E of 1). The present distribution of Moinian erratics on S1 relates to a more recent movement of ice from the Ben Nevis massif incorporating the erratics and subsequently transporting them to their present location. An explanation for the Nevis erratics along the Lundy valley(Peacock,1970b) could be that in an early stage of the last ice advance ice from the Ben Nevis Range and possibly

Glen Nevis flowed into the Great Glen before ice from the W had begun to cross the Great Glen(Although they were at a disadvantage, in terms of size of source areas of snow, the corries near locality 1 did have the advantage of close proximity to the Lundy valley). As ice from the W became more powerful the Nevis glacier was forced to diverge to the SW(p.99), whilst the local ice from the Ben Nevis massif was forced to diverge mainly to the E along the flanks of the massif and could no longer reach the Lundy valley.

Using the above interpretation it is suggested that the local ice surface declined in altitude from ca 650m O.D., at the exit of the tributary valleys, down to ca 550-600m O.D. where the glaciers from the Ben Nevis massif became confluent with the Spean trunk glacier. On this interpretation the lateral moraine and the upper limit of thick till described by Sissons(1979b, Figures 1 and 4E) relate to the upper limits of locally-derived ice and not to the main Spean glacier. Further support for this interpretation is derived from the upper limit of thick till and hummocky moraines, and from trimline values on the mountains around Loch Eil(based on unpublished mapping by the writer). Consistent evidence of an ice surface not exceeding 650m O.D. at the head of Loch Eil, that declined gently eastwards to ca 600m O.D. at the eastern end of the loch demonstrates that the surface of the Spean glacier cannot have been at an altitude of 640m O.D. on S1 as suggested by Sissons.

Farther down Loch Linnhe evidence for the former glacier is sporadic. Between 13 and 14 the limit is interpolated, although it is partly based on the likely form taken by the glacier surface and partly on a crude estimate using sparse forms of evidence. Maximal limits are imposed on the altitude of the glacier surface in the form of frost-riven bedrock and geliflucted debris above ca 400-450m O.D. in the vicinity of NN062665. The distribution of ice-smoothed bedrock, striae, friction cracks, morainic mounds and thick till in the vicinity of 14 suggest approximate ice-limits. The ice-limit drawn across the valley E of 14 is tentative as no clear evidence for a limit exists here, except for a few small mounds on the south side of the valley.

An imposing dry, meltwater channel ca 1km in length and over 50m deep occurs S of 14. Walker(1932) suggested that the channel represented the original course of the Abhainn Righ stream before it was diverted to the NW as a result of a landslip blocking the valley. However, the sheer size and the orientation of the channel suggests a more likely origin from meltwater, probably during ice-sheet times. The channel is strongly asymmetrical in cross section with the W side comprising a smooth, sheer wall of quartzite, whereas its more gently-sloping eastern side is composed of substantial numbers of boulder erratics and till. An extensive area of gravel occurs at the seaward end of the meltwater channel at Onich at an altitude of approximately 10m O.D. Farther to the W is a narrow valley excavated along the faulted contact between quartzite and slate, that is partly infilled with outwash deposits overlain by till(Peacock,1971a,1977). Peacock(1977) has suggested that the deposits might be interpreted as relating to the maximum extent of the Linnhe glacier with the gravel deposits at Onich representing a reworked moraine.

Similarly, the large mass of outwash gravel at Corran has been suggested as being formed at the maximum limit of the Linnhe glacier(McCann,1966).

The above interpretations are rejected by the present writer. Instead it is suggested that all the features described above lay well within the ice-limits of the Linnhe glacier. It is likely that the meltwater channel at Onich and the narrow valley to the W were partly infilled with glacial and fluvioglacial deposits during the advance of the Linnhe glacier. During a retreat stage meltwater utilised the two channels and formed an outwash fan at Onich(see p.222). Since the eastern end of the feature lacks a steep gradient that would suggest an ice-contact slope and since it does not slope down toward the W then possible formation at the snout of the Leven glacier is rejected.

Evidence for the termination of the Linnhe glacier is believed to be found 7km to the SW near Kentallen(15). Extensive gravel deposits fringe the coast intermittently for 3km from NN004579 to NN975569. They are depicted as raised beach deposits on Geological Sheet 53, but this mode of formation is rejected on several counts:

i) Their exposed location on promontories rather than in bays is anomalous in comparison with many other raised beach deposits in western Scotland.

ii) There are no streams in the vicinity to supply debris to the coast.

iii) The deposits exposed in the foreshore include numerous large boulders that are unlikely to have been moved along the coast by longshore drift.

iv) Morphologically the deposits form four distinct landforms each with a similar morphology. Three of the units have a steep slope along their north-eastern edge and a surface that declines in altitude from ca 10m O.D. in the NE to sea-level in the SW.

The four gravel areas are interpreted as comprising four outwash fans formed laterally at the snout of the Linnhe glacier, with a steep north-eastern edge representing an ice-contact slope for three of the fans. The maximal limit of the Linnhe glacier is taken as the ice-contact slope of the westernmost outwash fan. The other three fans are regarded as forming during retreat stages of the glacier.

5.6 Duror glacier

Field evidence in Glen Duror suggests that a glacier, existing independently of the main ice mass, flowed down the glen toward Loch Linnhe(Figure 5.2).

Highly, ice-moulded granite bedrock on the floor and below the lip of finely-developed corrie 71 is taken as signifying occupance by a corrie glacier supplying ice to the Duror glacier. Corrie 72 is less pronounced, but has a steep rockwall facing E that extends through a vertical distance of 280m. Numerous crescentic fractures occur on quartzite bedrock below the rockwall, that indicate an ice-flow direction toward the N. In contrast large numbers of crescentic fractures, but orientated to the W, occur to the S at higher altitudes at ca 500-700m O.D.(Figure 5.3, NE of 32) on severely frost-riven quartzite outcrops. This pattern strongly suggests that local ice, nourished below the steep rockwall, was controlled by the topography and flowed N into Glen Duror. The friction cracks at higher altitudes demonstrate ice flowing to the W under ice-sheet conditions, but independently of the topography. Furthermore, the valley floor below the corrie is mantled by extensive deposits of gullied till that lend support to the existence of local ice.

The evidence in corrie 70 suggests that this corrie failed to nourish a local glacier. The bedrock appears too angular and frost-riven and considerable quantities of scree mantle the corrie walls and floor.

Till-mantled slopes and sporadic areas of hummocky moraine occur down Glen Duror as far as Auchindarroch farm(NM001553). Here the farm is sited on a large ridge of till to the N of the river Duror. Large boulders abound on the surface of the ridge, that extends for ½km before it becomes rock-cored. The ridge is taken as representing an end moraine that delimits the maximum extent of the Duror glacier. Commencing a short distance within the limit are terraces that extend down-valley for 2km. The composition, width and gradient of the suite of terraces suggests that they represent a dissected, outwash train formed by meltwater from the Duror glacier.

The limits of the Duror glacier, however, as shown on Figure 5.2 must only be regarded as approximate since no clear evidence relating to the upper limit of the glacier, such as lateral moraines or the upper limit to thick till, were located in the field. In addition because of the inferred low altitude of the glacier periglacial evidence is sparse or equivocal and renders it difficult to use in imposing maximum altitudes on the former glacier surface.*

5.7 Creran glacier

The major source areas of the Creran glacier lay at the head of Glen Creran itself and in the mountains surrounding upper Glen Etive(Figure 5.3).

Excellent roches moutonnées and smooth, ice-moulded

* **See addendum at the end of this chapter**

Figure 5.3 Detailed field evidence and Loch Lomond Advance limits for the S.W. quadrant. Evidence and ice-limits of the terminal zones of the Creran and Etive glaciers are based partly on J.M. Gray (1972, 1975). For explanation of symbols see Figure 5.2

schists in the vicinity of corries 83,87,88 and 89 indicate a flow of ice out of the corries that diverged NW into upper Glen Creran and southwards into upper Glen Etive. Similar bedrock evidence for C102 suggests a supplementary flow of ice from corrie 80 into Glen Creran. An upper ice-limit of ca500-550m O.D. is discerned from the trimline evidence on S99 and S101, although the ice in the corries would have been at a higher altitude.

In upper Glen Etive till and hummocky moraine occurs on the valley floor and lower parts of the valley sides and extends intermittently for a distance of 13km between Rannoch Moor and the head of Loch Etive.

In the valleys at 37 and 38 thick,superficial deposits obscure the bedrock and appear to represent an amalgam of till, till reworked by streams and scree washed down from the steep, upper slopes of the flanking mountains; easily recognisable hummocky moraine is absent.

About 25 upper ice-limit values(mainly type A trimlines, Appendix A) enable a former glacier surface to be reconstructed that sloped down from ca 600-700m O.D. in the vicinity of 37,38 and 39 to ca 500-550m O.D. at the head of Loch Etive. Here the ice diverged around Beinn Trilleachan(S of 35) with some of the ice continuing SW along the line of Loch Etive, whilst the remainder flowed strongly toward the glacial breach at 33. Evidence for the latter ice-flow direction is clearly demonstrated by the spatial pattern of friction cracks and roches moutonnées, by extensive areas of smooth, ice-scoured granite bedrock to the N and S of 35 and by ridges of bedrock and till,streamlined by the ice within the steep-sided breach.

Great numbers of friction cracks S of 35 and large areas of ice-scoured Starav granite show that ice flowed N out of the two shallow corries of 113 and 114 to add to the flow of ice through the breach. The rapid descent of the former glacier surface through the breach is indicated by the evidence for ice-limits in Glen Creran.

Large ridges of till and mounds to 10m in height and numerous smaller mounds extend down-valley for 8km from near 34 to W of the glacial breach at 33. Thereafter the valley floor is occupied by river terraces and floodplains, and by outwash and raised beach deposits as far as Loch Creran.

Various forms of evidence combine to suggest upper limits to the former Creran glacier. E of 32 severely frost-riven quartzite occurs down to altitudes of ca 400m O.D. on S104 and ca 375m O.D. on S51.Below these altitudes the valley slope is plastered with till and the few exposures of bedrock are smooth and unaffected by frost action. Col 105 at an altitude of 391m O.D. and the valley to the W of the col lack any clear evidence for ice diverging W over the col from Glen Creran. At 30 large morainic ridges orientated NNE to SSW occur on the valley side at an altitude of ca 300-330m O.D. and were probably formed close to the upper limit of the glacier. A large morainic ridge partially blocks the tributary valley at 29 and runs downslope from ca 200m O.D. to ca 140m O.D. at which point it has been breached by the tributary stream. However, this ridge is interpreted as a medial moraine formed at the junction of the Creran glacier with a tributary glacier flowing out of the valley at 29 and only provides a minimum altitude for the glacier surface(see previous discussion, p.32).

Collectively the evidence described above suggests a former glacier surface of ca 400-450m O.D. in the vicinity of 34, but declining to only ca 275-300m O.D. at 29.

NW of 34 a lobe of ice projected northward into a tributary valley creating evidence in the form of distinct hummocky moraine containing many boulders of granite derived from the local bedrock(Figure 5.2). The northern limit of the lobe of ice is drawn where the hummocky moraine terminates.

SW of 33 the Creran glacier received smaller quantities of additional ice from the tributary valleys at 27,28,29 and 30. The narrow and deeply-incised corrie (c111) to the ESE of 30 nourished a small glacier. The evidence consists primarily of ice-scoured bedrock and numbers of friction cracks that point to a NE-directed flow of ice out of the corrie. Excellent steep-sided hummocky moraines occur at the eastern end of the valley at 28 and are supplemented by other morainic mounds and quantities of till lower down the valley. This evidence is taken as indicating occupance by ice contemporaneously with the ice in Glen Creran.

Farther W at 27 deposits of thick till and numerous mounds extend to the W for over a kilometre from an altitude of ca 100m O,D. down to ca 50m O.D.(Gray,1972). Deep enclosed hollows by the road at NM966404 suggest that large blocks of ice in the embayment wasted down in situ. W of the mounds a number of small meltwater channels converge at the entrance to a large meandering meltwater channel that has been cut parallel to the valley side. A partial source for the meltwater may have been the flow of meltwater from the Etive glacier at 52.

The northern limit of the Creran glacier is represented in the Strath of Appin by large morainic mounds at 26(Gray, 1972) that are interpreted as forming an end moraine. The moraine has been breached by two large meltwater channels, one 5m below the intake of the other, and possibly formed during the retreat stages of the glacier from its maximum position. This evidence is supplemented by thick till exceeding 15m infilling two tributary valleys immediately to the E.

The maximum extent of the Creran glacier is believed to occur in the vicinity of 25(Peacock,1971a; Gray,1972,1975a). An arcuate moraine runs parallel to the coast S of South Shian. However, both Peacock and Gray refer to till mounds surrounded by outwash deposits to the W of the moraine and suggest that the ice advanced farther W than the end moraine. Gray cites supporting evidence in the form of a broad ridge 300-400m wide that reaches a maximum height of 18m O.D. and extends from Baravullin(NM902402) to Culcharron(NM915394). This may represent an additional end moraine and is taken as such by the writer in delimiting the maximal extent of the Creran glacier as shown on Figure 5.3.

5.8 Etive glacier

Numerous sources supplied ice to the Etive glacier. One major source area in the mountains at the head of Glen Etive has been discussed previously(p.109). Another important source area was the Ben Starav Range, extending from 40 to 59, and which from the field evidence contained 30 corrie glaciers that supplied ice to the main Etive glacier. Other sources were the Ben Cruachan(55 to 62) and Creran-Etive(33 to 50) Ranges.

In the valley at 43 a massive morainic ridge projects westwards from the spur between corries 106 and 120. Till to a minimum thickness of 12m has been exposed along its margins as a result of stream erosion. The ridge is interpreted as a medial moraine formed from debris dumped by ice converging from the two corries. This evidence is supplemented by extensive deposits of till on the valley floor and sides and W of 43, where the valley narrows, by striking areas of polished and ice-moulded granite bedrock. This evidence is in direct contrast to the frost-riven bedrock and soliflucted debris on the upper mountain slopes above the inferred ice-limit(see Appendix A, spurs 136,138,139 and 140). Similar contrasting evidence is to be seen in the vicinities of 44,45,46 and 47. S of 46 the Starav granite bedrock has been swept clear of debris over a considerable area and is remarkable for showing virtually no evidence for frost-riving up to altitudes of 600m O.D. or more.

A massive accumulation of till to minimum thicknesses of 10m and several large morainic ridges N of 59 partially infill a tributary valley. They indicate the convergence of ice, flowing from two shallow corries on the east side of Ben Starav, with the larger Etive glacier. Numbers of friction cracks on bedrock in the two corries and in the vicinity indicate ice flowing south-eastward out of the corries, turning S along the shallow valley N of 60 and finally descending westward toward Loch Etive. Just N of 60 the valley floor and valley sides are mantled by till, hummocky moraines and ridges of till that extend up the valley sides to altitudes of <u>ca</u> 550-600m O.D.; their abrupt termination at this altitude suggests an ice-limit for the glacier flowing S toward Glen Kinglass.

A low plateau averaging 400-450m O.D. occurs at 59. The local bedrock comprising massive Starav granite is remarkably smooth and swept clear of debris and this suggests that ice overtopped the plateau. Supporting evidence is provided by a few hummocky moraines and quantities of till in a shallow depression on the S side of the plateau. No easily-discernable trimlines,however, could be located on the slopes and spurs immediately to the E of the plateau and S163 and S164 provide only approximate values of ca 500m O.D. for the upper ice-limit in this area.

Farther S the valley at 56 contains numerous well-defined hummocky moraines, moraine ridges and moderately thick till. This evidence is complemented by fresh-looking, ice-moulded bedrock in most of the six corries on the N side of Ben Cruachan. These provided the ice that accumulated in the valley at 56. Col 173 and S171, S172 and S174 enables the former surface of the ice to be reconstructed from ca 550m O.D. at the head of the valley down to ca 475m O.D., where the tributary glacier became confluent with the Etive glacier.

Only two corries(108 and 130) are believed to have supplied ice to the Etive glacier on its western side. Unlike a number of corries in the Creran-Etive Range corrie 108 is well-developed with a steep, craggy backwall. Here the evidence consists of ice-moulded bedrock and good numbers of friction cracks on and below the corrie floor.

Great numbers of large boulders of porphyritic Starav granite, averaging 1-3m in length and many hummocky moraines extend from the vicinity of 48 to almost as far as 35. Some of the mounds and ridges exceed 10m in height and are particularly prominent along the floor of the valley to the NE of 48.

Several lines of evidence combine to suggest that the maximum altitude of the surface of the Etive glacier between 33 and 49 lay between values of ca 460-550m O.D. The evidence on spurs 111,112 and 115 and the upper limit of low hummocky moraines suggest upper ice-limits of ca 500m O.D. in the area. The periglacial evidence on C108 imposes an approximate maximum limit of ca 570m O.D. on the altitude reached by the former glacier surface in the area N of 49. Similarly, the evidence farther S on C167 and C168 in the form of moderately frost-riven bedrock and geliflucted deposits implies that the ice was unable to overtop these two cols. This places a maximum upper limit to the glacier surface of ca 450-500m O.D. in this area(Appendix A). However, the presence of a few steep-sided mounds, packed with boulders derived from the local bedrock, on the eastern side of col 109 at ca 520m O.D. indicates that ice had just begun to advance across the col. Therefore the surface of the ice in the vicinity of the col is likely to have been at ca 530-540m O.D. The strongest evidence for the minimum altitude of the former glacier surface is demonstrated by numerous well-defined hummocky moraines on the floor of col 110 and on the slopes below the col. Since the altitude of the col is 460m O.D. this suggests that the surface of the ice must have reached at least ca 480-490m O.D. in order to spill over the col. Thick deposits of till, hummocky moraines and great numbers of granite boulders in the vicinity of 52 corroborate the view that ice spilled over C110 from Glen Etive and flowed to the SW.

Thick till, numerous ridges of till and large numbers of granite boulders at location 50 demonstrate that ice also overtopped col 169. With an altitude of only 305m O.D. this particular col is considerably lower than the other cols along the Creran-Etive Range.

The ice-limit at 55 is taken as the upper limit of hummocky moraines littered with granite boulders at about 380-400m O.D. Above the limit slopes are smooth and lines of boulders suggest transport by gelifluction processes from the steep crags at higher altitudes. That those processes no longer operate today is suggested by the lack of ploughing blocks at that altitude. Below the limit the ground consists of numerous rock outcrops, that have been streamlined in a NNE-SSW direction by the ice. Where exposed the surface of the outcrops is strongly smoothed and mamilated by former glacial processes, although many of the rock knobs are covered by varying thicknesses of till.

From 55 to 53 clear evidence for an ice-limit is lacking and the limit shown is an interpolated one based on the evidence at locations 51,53 and 55, and on the topography and on the likely form taken by the Etive glacier(see section 6.4).Several attempts by the writer to find evidence of an ice-limit failed. For example, the area W of 54 consists of an irregular terrain of steep-sided rocky valleys and craggy ridges composed of the Lorn Volcanic Series. Neither hummocky moraine nor thick till were located in this area. The assumptions are that either the ice was very clean and had lodged much of its debris before reaching this area or that the piedmont lobe of ice flowing out of the mountains was smaller and more strongly controlled by the form of Loch Etive than is shown in Figure 5.3. Conversely, if Gray's view(1972) is correct the piedmont lobe of ice may have been much larger and extended as far as Glen Feochan (5km SE of Oban). However, in view of the field evidence presented in this thesis this possibility is considered an unlikely one by the writer. Clearly further work will be required in the future to resolve this particular problem.

That ice from the Etive glacier flowed into the Pass of Brander(54) is suggested by a number of moraine ridges and mounds, and by the series of outwash terraces mapped by Gray(1972). Clearer evidence for an ice-limit, however, is to be found N of Loch Etive at location 51. Here thousands of large boulders,especially Cruachan and Starav granite boulders and a considerable quantity of till were carried into the tributary valley. Downcutting by the river Esragan has since dissected the till into bluffs 10-15m high. Moraine mounds and a small rock-walled meltwater channel in the vicinity of NM988370 indicate that ice penetrated beyond the col at the head of the deep meltwater channel sloping down to the N toward Loch Creran(51 to 27).

The terminus of the Etive glacier is drawn beyond the large outwash spread at the mouth of Loch Etive(McCann,1966). This is based on the assumptions that the large mound at Saulmore farm (NM894334) represents an end moraine(Gray,1972) and that the large kettle holes in the outwash spread provide a minimal position reached by the ice-front.

The thinning and/or retreat of the Etive glacier to the E is clearly illustrated by the detailed work carried out by Gray(1972) on the kame terraces bordering the loch.

5.9 Kinglass glacier

The Kinglass glen is a narrow, steep-sided valley flanked on both sides by mountains exceeding 900m in height. The mountains on the W side present a stark landscape with considerable areas comprising bare slopes of ice-scoured Starav granite.

Trimlines on spurs 157,159,160,166 and 187 at the head of Glen Kinglass(64) demonstrate a glacier surface of _ca_ 600m O.D. to _ca_ 670m O.D. declining in altitude toward the SW down the glen and southwards toward Glen Strae. Areas of hummocky moraine and till on the valley floor and sides contrast starkly with the bare granite on the upper slopes, scoured clear of debris by the ice.

Numbers of friction cracks on aplite veins in the Starav granite in corries 145 and 156 and on the valley sides at 61 and 64 clearly show ice flowing firstly to the S or SE only to turn sharply to the SW on becoming confluent with the Kinglass glacier.

An anomalous area of bare Starav granite, that extends for about 2km down-valley, occurs at 63. It is anomalous because it occurs across the valley floor whereas in all the other areas of Starav granite the valley floor is normally covered by till, hummocky moraine and fluvial deposits. Its location directly below a low-level col(NN190413) just N of 63 at an altitude of 245m O.D. is of some significance. Ballantyne(1979) has suggested that the col operated as an englacial escape route for the waters of glacial Loch Tulla when the level of the lake fell from 315m O.D. to 248m O.D. Since on this interpretation a jökulhlaup is likely to have occurred as the lake drained quickly down to the lower level(Sissons,1977a, 1981), the zone of bare rock below the col becomes explicable. It represents a belt of water-washed rock formed by the catastrophic release of lake water,as a result of a new escape route being opened up at a lower level.

SW of 63 a diffluent flow of ice from the Kinglass glacier overtopped col 183 and combined with the ice in Glen Strae. Farther W at 62 the distribution of hummocky moraine and thick till inside the inferred ice-limit and frost-riven outcrops of bedrock and gelifluted debris outside the limit enable the limits of an ice-tongue to be discerned that failed to overtop col 201.

Between 62 and 56 a series of N-facing corries and rockwalls nourished ice that fed into the Kinglass glacier. Trimlines are generally far more difficult to discern than on the mountains at the head of the glen(Appendix A, spurs 174 to 178),because of the decreasing altitude of the former glacier combined with the high resistance of the Starav granite to frost-riving.

In the vicinity of 57 numerous roches moutonnées and meltwater channels trend to the W oblique to a series of small valleys and low ridges that are orientated in a WNW to ESE direction. Clearly here the flow of ice was controlled not by the topography, but by the surface gradient of the ice sloping down to the W toward the narrow exit from the mountains between 50 and 55. A typical pattern of till and hummocky moraine on the valley floors and bare ice-scoured ridge summits predominates in this area.

5.10 Rannoch Moor ice-cap in the west

Only a brief description is provided in this section as most of the evidence is shown on Figures 5.4 and 5.5 and discussed in sections 5.14, 5.15 and 5.17.

Strongly ice-moulded bedrock in corries 60,102,103, 104,105 and 127 show that ice occupied the corries and flowed eastward into Rannoch Moor. Supporting evidence is provided by numerous erratics of volcanic material, derived from several of the corries, that can be found spread eastward across the Moor(Hinxman, *et al*, 1923: Bailey, *et al*, 1960; mapping by the writer).

Trimline values along the mountain rim on the W side of the Moor range from *ca* 635m to *ca* 700m O.D.(spurs 77,141, 142,145,150 and 153) and demonstrate that the height of the ice-cap in this area reached a maximum height of *ca* 700m O.D. Evidence of frost-riven bedrock and gelifluoted debris on cols 143,146 and 148 impose further maximal limits on the upper ice-limit of the ice-cap since the altitudes of the cols are all close to 700m O.D.(Appendix A).

A number of low morainic mounds occur on the floor of a shallow depression at location 41 between altitudes of 700m and 760m O.D. and E of a mountain (Meall a' Bhuiridh) that reaches to 1108m in height. The mounds are full of small fragments of volcanic rock derived from the slopes of Meall a' Bhuiridh. The majority of the mounds have been modified by solifluction processes; numerous small solifluction terraces, that appear to be active at the present time, occur on the slopes of the mounds.

The spatial arrangement of the mounds suggests that a small glacier was nourished below the moderately steep slope on the E side of Meall a' Bhuiridh. This glacier became confluent with the Rannoch Moor ice-cap and probably helped to provide the volcanic erratics that are to found further E on Rannoch Moor(see above). The presence of solifluction terraces within the inferred Loch Lomond Advance limit is believed to be related to the nature of the volcanic debris, which consisting of many small rock fragments, is well-drained and is particularly susceptible to solifluction processes.

5.11 Strae and Orchy glaciers

The ice that formed the Strae and Orchy glaciers was primarily derived from external source areas. The main source areas were Rannoch Moor, the corries NW of Loch Tulla and the corries on the mountains at the head of Glen Lyon(Figures 5.3 and 5.5).

Large numbers of hummocky moraines and considerable quantities of till abound over much of the lower ground, especially in parts of the Tulla basin. Various ice-direction indicators show that ice from the three main source areas accumulated in the Tulla basin and flowed down Glen Orchy. Evidence in the form of large morainic mounds and ridges on cols 189,191 and 197,at altitudes of 370m to 520m O.D.,indicate that ice from the Tulla basin eventually overtopped the cols and flowed into glens Strae and Orchy. In areas such as 65,66 and 68 till to thicknesses of 25m accumulated on the floors of the valleys.

Although striae and friction cracks are relativ-ely sparse in the area, at two locations they provide clear evidence of small tributary glaciers becoming confluent with larger valley glaciers, as for example NW of 70 and S of 67.

From 69 to 70 a distinct upper limit to thick till and morainic mounds provides an ice-limit declining to the SW from about 430m O.D. down to 380m O.D. The slopes above the limit are characterised by thick accumulations of frost-riven debris, some of which has been gelifluoted downslope to form lobes and terraces.

Trimline values derived from spurs 184 and 204 provide strong evidence for the Strae glacier surface declining in altitude from ca 530m O.D. to ca 350m O.D. over a distance of 7km.

5.12 Small independent glaciers in Glen Salachan

The only glacial evidence for possible glaciers other than in Glen Duror, existing independently of the main ice mass on its western side, is located near the head of Glen Salachan.

At 31 till to minimum thicknesses of 10m extends for over a kilometre along the floor of a valley(Glen Dubh) tributary to Glen Salachan. Some well-defined hummocky moraines with large boulders on their surfaces occur at two localities on the valley sides. Immediately to the W of this evidence lies a steep rockwall below a broad, flat-topped mountain reaching to 654m O.D. The rockwall is taken as the source area for a small glacier whose limits are defined, in part, by the distribution of thick till and hummocky moraine. However, some doubt must be cast on these limits since thick till ($>$10m) occurs in a deep gully(NM999517) some $\frac{1}{2}$km farther down the glen beyond the postulated glacier limit(Figure 5.2).

Till with thicknesses exceeding 10m in places chokes the floor of a small valley at 32. Streams have dissected the till to form steep bluffs that are still undergoing active erosion. The till lies E of a low rockwall that forms the eastern edge of another broad, flat-topped mountain that has a summit altitude of only 560m O.D. In view of the similarities with the evidence at 31 a small glacier nourished below the rockwall is suggested and the northernmost limit of the thick till is interpreted tentatively as indicating the terminal limit of the glacier.

Nevertheless, the writer still has some reservations about the existence of the two proposed glaciers in Glen Salachan since it is possible that the ice-sheet could have deposited the thick till in the hollows between the mountain ridges. Abundant friction cracks to the E of 32 show that ice-sheet flow was perpendicular to the two tributary valleys. Against this proposal, however, are a number of observed facts:

i) Thick till is absent elsewhere in Glen Salachan except at NM999517.

ii) It seems more than fortuitous that the thick till in each case lies directly to the E of a steep rockwall.

iii) The writer has not located any such other extensive areas of till <u>outside</u> the proposed ice-limits in the western half of the study area. But it must be noted that both Sissons(1974a) and Cornish(1981,1982) have described thick deposits of gullied till outside the limits of the Loch Lomond Advance in the eastern Grampians and western Southern Uplands respectively, so that the possibility of the till in Glen Salachan being deposited by the ice-sheet cannot be excluded.

To resolve this problem work will be necessary in the future involving provenance work on the rock material in the till to determine whether it was derived from the bedrock to the E or alternatively from the rockwalls to the W.

5.13 Nevis-Laire glacier

Much of the evidence relating to the western part of the Nevis glacier has been described in section 5.2. In this section the evidence relating only to the eastward flow of ice from Glen Nevis will be described(Figure 5.4).

Thick,dissected till mantles the sides and floor of the valley at 73 at the head of Glen Nevis. Inferred ice-limits on spurs 207,208,211 and 212 with respective values of 820,760,860 and

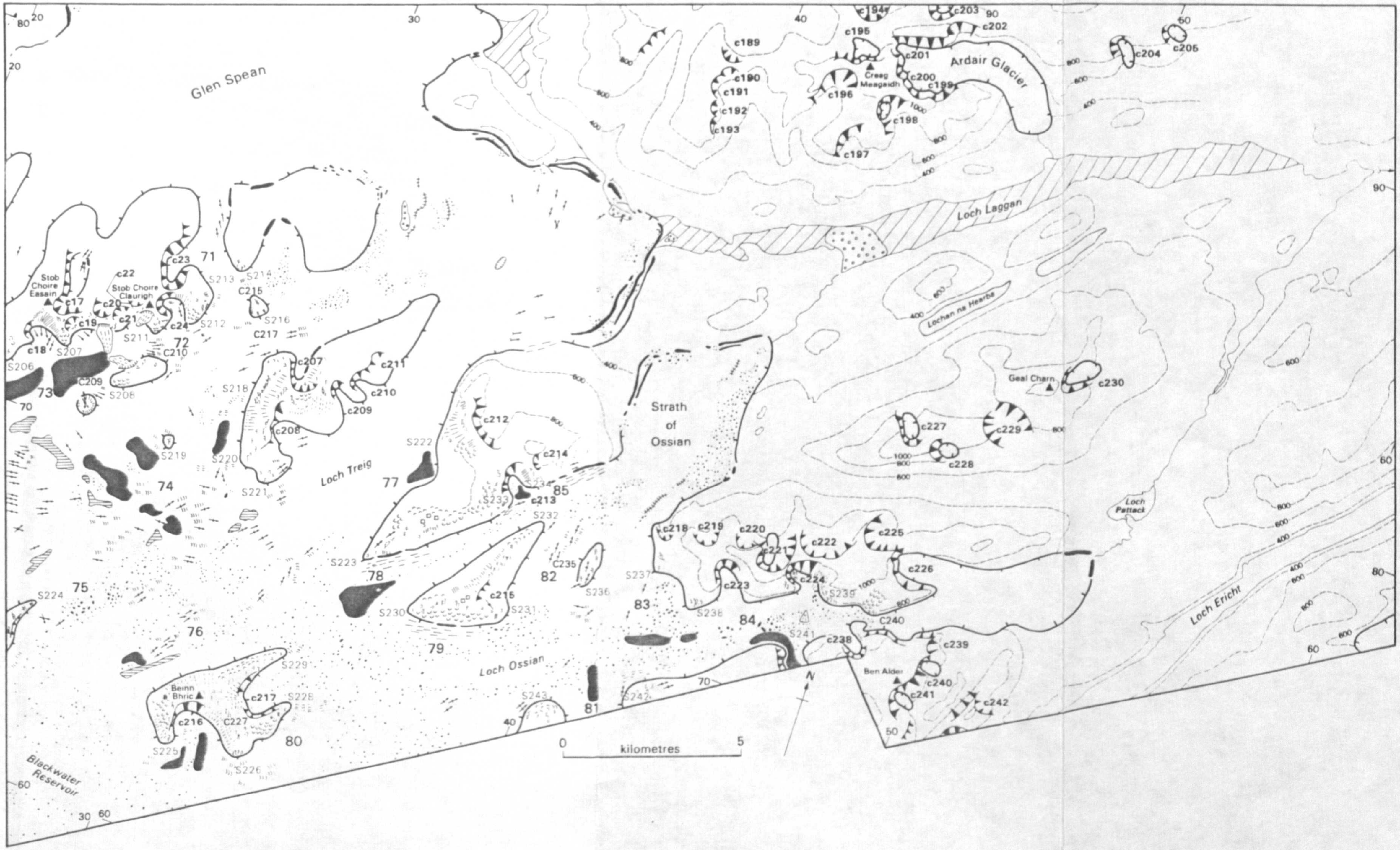


Figure 5.4 Detailed field evidence and Loch Lomond Advance limits for the N.E. quadrant of the study area. Based on Sissons (1979, unpublished) and the writer's own mapping. For explanation of symbols see Figure 5.2. Contours omitted on nunataks

740m O.D. signify an upper limit to the glacier surface within a range of 750-850m O.D.

Four SE-facing corries occur at the head of Glen Nevis on its N side. Corrie 24 is the best developed with steep, craggy rockwalls and it contains evidence of occupance by ice in the form of strongly ice-smoothed rock knobs and roches moutonnées on its floor. Thick,gelifluctеd quartzite debris and solifluction lobes mantle the SW-facing slope of the corrie and provide a maximum upper limit to the corrie glacier of ca 800m O.D. on that side of the corrie. Corries 19 and 21 are much smaller and both are choked with large quantities of frost-riven quartzite debris. This implies that ice failed to form in the two corries and an ice-limit is accordingly drawn across the exits of the two corries. Although the eastern side-wall of corrie 18 is mantled by thick,quartzite scree a small glacier may have occupied the western side of the corrie where numbers of friction cracks score the surfaces of ice-scoured, quartzite bedrock.

An easterly movement of ice from the head of Glen Nevis is recorded by numerous friction cracks E of 72 and S of 73, on smooth,ice-polished,quartzite surfaces. Ice spilling eastwards over cols 209.210.215 and 217 toward the Laggan valley is similarly indicated by roches moutonnées and by the plucked edges of ice-smoothed rock knobs.

Strong controls on the maximum upper limit of the Nevis-Laire glacier are imposed by the distribution of severely frost-riven bedrock and gelifluctеd debris, as for example above the trimline along spurs 212,214,216 and 218. The trimline values indicate an ice surface declining from ca 750m O.D. at 72 to about ca 650m O.D. at 71, which is slightly at variance with Sissons'(1979b) interpretation.

The main discrepancy is in the vicinity of S212 where Sissons has shown an ice-limit at an altitude of over 800m O.D. Very severely frost-riven quartzite bedrock, however, extends down to ca 750m O.D. on S212 and Sissons' limit is taken as an over-estimation based on extrapolation.

Farther E the evidence for the limits of the Nevis-Laire glacier are taken from Sissons(1979b, Figures 1 and 3) and, overall, there is very good agreement between those limits and the limits mapped by the present writer farther W.

5.14 Treig glacier

Comprehensive field evidence shows that the Treig glacier was nourished by ice from several major sources. These comprise: i) the mountains at the head of Glen Nevis; ii) the mountains between the Blackwater valley and Loch Treig; and iii) the northern area of Rannoch Moor.

The relief of the area roughly encircling locations 74, 75 and 76(Figure 5.4) consists of numerous low mountains and ridges, generally below 650m O.D. in altitude, that have been dissected and isolated by the development of small valleys, most of which slope down toward Loch Treig.

Hummocky moraines mantle much of the valley floors. Particularly good examples, to over 10m in height, extend down-valley to the NE for over 2km at 75 and are accompanied by great numbers of psammitic boulders. Roches moutonnées abound in the area and range from highly, ice-polished examples fringing the southern shores of Loch Treig to those along the summits of ridges at altitudes of ca 600m, as near 74. These, together with numbers of friction cracks, demonstrate a strong convergence of ice toward the deep,glacial breach occupied by Loch Treig. At 76 large numbers of friction cracks on ice-scoured, psammitic bedrock demonstrate that ice flowed N <u>across</u> the watershed between 75 and 80 from the Blackwater valley area.

Ice-smoothed bedrock on the summits of a number of low mountains provide minimum values for the ice surface of <u>ca</u> 530m O.D. to <u>ca</u> 600m O.D. in the area between Glen Nevis and Loch Treig. Trimline values and other ice-limits in the area range from 620m O.D. to 690m O.D.(spurs 219,220,221,224 and 229) and indicate a maximum ice surface declining gently in altitude toward the head of Loch Treig. Nevertheless, anomalous evidence does exist in the area and requires some discussion.

Frost-riven bedrock and appreciable quantities of frost-riven debris occur above heights of <u>ca</u> 600m O.D. along the summits of two mountain ridges W and E of 75(NN270660 and NN285648) and, therefore, lie below the altitude of the glacier surface calculated above. A possible solution to resolve this anomaly is that whilst the valleys between the ridges were occupied by thick,active ice the summits of the two ridges were covered by only a thin carapace of inactive ice that failed to remove the debris or to smooth the frost-riven bedrock.

This anomalous evidence together with the low trimline value of 640m O.D. for S229 led the present writer to believe that the altitude of the glacier surface approximated 600-640m O.D in the area S of Loch Treig. The discovery of a clear lateral moraine at an altitude of 660-670m O.D. on the valley side E of S223(N of 78) led to a revision of this estimate. The moraine extends for about a kilometre and forms a low,broad ridge that is separated from the valley side by a low depression,averaging only a metre in depth.Below the moraine the valley side is covered by many morainic mounds and ridges.

Hummocky moraines and extensive areas of till occur in corries 216 and 217. These corries nourished ice, some of which flowed N toward Loch Treig. This ice-flow direction is supported by the distribution of Rannoch Moor granite erratics. A large boulder field of such erratics occurs N of 80 and extends down-valley as far as Loch Treig. A distinct limit to the boulders runs NE to SW near the head of the loch(Sissons,1979b) and signifies convergence of ice from Rannoch Moor with ice derived from the areas described above.

Granite erratics from Rannoch Moor are also to be found in the valley E of 76 and this supports evidence cited earlier (p.127) that ice flowed N toward Loch Treig from the Blackwater valley area.

Beyond the thick till at 77 the ice-limits are those of Sissons(1979b) with which the present writer's evidence and ice-limits strongly correlate.

5.15 Ossian glacier

The maximal limits of the Ossian glacier are defined by two prominent end moraines(Sissons,1979b) and by numerous hummocky moraines within the limit.

At 85 a series of morainic ridges trend across the valley side toward the NE up to a maximum altitude of ca 600-620m O.D. Above this altitude there is much bare Rannoch Moor granite bedrock that in places is strongly frost-riven. This ice-limit can be traced to the SW to corrie 214 where a morainic ridge extends across the corrie exit and implies that the corrie lacked a glacier at this time. Conversely, corrie 213 contains hummocky moraine and till and occupance by ice contemporaneously with ice in the main valley is indicated.

An excellent series of northward-directed roches moutonnées and much bedrock scoured clear of debris at 82 at an altitude of ca 600m O.D. contrasts with the extensive areas of hummocky moraine at lower altitudes and with much frost-riven bedrock and thick geliflucted deposits at altitudes greater than ca 650m O.D.

Evidence for upper ice-limits in the Loch Ossian area occurs as the upper limit to morainic ridges at ca 670m O.D. trending WSW to ENE across the valley side at 79. Above the uppermost ridge the slope comprises thick quantities of geliflucted material derived from the rocky crags above. In many places the debris is arranged into large solifluction lobes and terraces. This evidence complements the hummocky moraine evidence mantling the floor of a shallow valley E of 83 up to maximum altitudes of ca 660m O.D.

N of 82 and 83 the evidence shown is derived from Sissons(1979b).

Although the valley of Uisge Labhair(location 84) sloping down to the SW contains hummocky moraine, morainic ridges and thick till(>15m in places) along much of its length, upper ice-limits are difficult to discern. Spurs such as 238,239 and 242 provide possible limits ranging from 640m O.D. to 740m O.D. Further support for ice-limits in this altitudinal range are derived from the glacial evidence on cols 240 and 241. These provide minimum altitudes reached by the glacier surface of 722m O.D. and 653m O.D. respectively. Till to thicknesses of _ca_ 5m and hummocky moraine cover the floor and sides of C241 whilst on C240 hummocky moraine and large numbers of boulders occur below frost-riven bedrock and gelifluction forms on the upper slopes.

Corries 224 and the rockwall SE of 84 provide strong evidence for nourishing glaciers. Corrie 224 contains quantities of till and on its eastern side an ice-margin can be distinguished separating morainic mounds from thick geliflucted deposits. Thick,dissected till mantles the lower part of the steep rockwall SE of 84 and this is taken as indicating nourishment of a glacier below the steep rockwall.

A major source area for the Ossian glacier was the upper Blackwater valley and corries 216 and 217. Trimlines on spurs 225,226 and 228 signify that the surface of the Rannoch Moor ice-cap reached a maximum altitude of _ca_ 700-730m O.D. in this part of the Moor.

Immediately N of 81 thick till infills the valley floor to depths greater than 8m,whereas to the S till is relatively sparse and the peat-covered slopes are generally smooth,except for one or two small morainic mounds. A possible ice-limit across the tributary valley is implied by this evidence, but corrie 232 that lies at the head of the tributary valley is the largest and deepest corrie in Rannoch Forest Range. Thus the present writer finds it difficult to believe that this corrie did not nourish a glacier and, therefore,an ice-limit has been drawn to include corrie 232, although this must be regarded as only tentative until stronger evidence is discovered.

5.16 Small independent glaciers in the Laggan area

The limits of the nine small corrie glaciers and of the larger Ardair glacier shown on Figure 5.4 are based on published and unpublished information by Sissons(1979b), and the evidence upon which they are based is not discussed here.

5.17 Rannoch Moor ice-cap and the Rannoch glacier

The Moor of Rannoch is covered by large numbers of hummocky moraines, although they are not ubiquitous everywhere since they are absent on the ridges and low mountains that rise above the peat-covered level of the Moor.

Good numbers of roches moutonnées occur in the vicinity of 86 and these provide consistent evidence of an easterly flow of ice across the Moor from the mountains along the western edge of the basin. This corroborates evidence derived from the distribution of volcanic erratics of an easterly flow of ice(see p.119).

Evidence has been cited previously to demonstrate that ice flowed westward(p.95), northeastward(p.126) and southward (p.121) from the Rannoch Moor ice-cap. The largest mass of ice, however, flowed eastward down the Rannoch valley where its width was greater than 10km for most of its length. Striae mapped by the Officers of the Geological Survey, friction cracks, roches moutonnées, meltwater channels and eskers all combine to provide convincing evidence of this massive flow of ice to the E.(For example see the evidence between locations 91 and 96).

Evidence for ice-limits is mainly derived from the upper slopes of the mountains fringing the S side of the Moor and from the sides of the Rannoch valley. Spurs 253,254,259,260 and 262

Figure 5.5 Detailed field evidence and Loch Lomond Advance limits for the S.E. quadrant. Ice limits and evidence for the terminal zones of the Rannoch and Lyon glaciers and independent glaciers partly after Thompson (1972) and Sissons (1979). For explanation of symbols see Figure 5.2

provide values ranging from 615m O.D. to 660m O.D. for upper ice-limits. At 96 the orientations of striae and roches moutonnées show that the ice flowed to the NE from the head of Glen Lyon through a deep, glacial breach to become confluent with ice flowing eastward across Rannoch Moor.

Farther E at 95 the eastward-flowing ice overtopped the summit of a ridge at an altitude of ca 600m O.D. and partly infilled a tributary valley with thick till that has since been deeply dissected by streams. The upper limit of thick till is taken as the ice-limit in the tributary valley, although the evidence is not altogether clear since moderately thick till and a few mounds occur in the upper part of the valley. Nevertheless, since the valley has no steep rockwalls and is very shallow it is difficult to envisage ice forming in the upper part of the valley and hence the limit that has been drawn.

Similar areas of deeply dissected till, with river bluffs greater than 15m in height, are to be found near 93 and 94. The thick till terminates abruptly at 94 where a low morainic ridge occurs. The E - W ridge is interpreted as an end moraine that crosses the valley floor climbs the spur to the E to become an irregular-shaped ridge composed of heaps of locally-derived psammitic boulders. In addition boulders are numerous within the limit but are very sparse outside the limit. This evidence implies that the corrie to the SW (c267) either lacked a glacier or only nourished a very small glacier during the Loch Lomond Stadial. The former assumption appears to be supported by the lack of ice-moulded bedrock, till or hummocky moraine within the corrie. Instead the sides and floor of the corrie are covered by thick deposits of soliflucted debris that are undergoing dissection by streams at the present time. Hence no glacial limits are shown within the corrie(Figure 5.5).

Evidence of an additional ice-limit exists E of 93. Here a series of low, arcuate, morainic ridges curve across the floor of a col at an altitude of ca 620m O.D. and impound a small lochan. On the S side the main ridge runs parallel to the hillside creating a small depression only 1-2m deep that has deflected the streams flowing down the hillslope. These features are interpreted as two end moraines that merge into lateral moraines farther W. This ice-limit at ca 620m O.D. correlates well with Thompson's(1972) evidence of an ice-limit at ca 600m O.D. a kilometre to the N near to where a massive lateral moraine crosses a tributary valley(NN549-518).

Farther E the ice-limits and much of the morphological evidence shown on Figure 5.5 are derived from Thompson's(1972) thesis.

On the N side of the Rannoch valley at 88 a steep-sided morainic ridge, with many Rannoch Moor granite boulders on its surface, runs straight down the valley side from an altitude of 530m O.D. down to 470m O.D. near the valley floor. S of the ridge hummocky moraines are numerous and moderately thick till occurs on the valley floor. N of the ridge is a valley(Allt Eigheach) 4km long with rockwalls along its western side. No clearly-defined hummocky moraines or thick deposits of till occur along the entire length of the valley; smooth, peat-covered slopes tend to predominate.

The ridge is interpreted as an end moraine formed along the very steeply-sloping margin of the Rannoch Moor ice-cap whilst the E-facing rockwalls of the Allt Eigeach failed to nourish any local ice and the valley remained free of ice. Possible snow beds, however, may have existed below the rockwalls in places, since mounds of psammitic boulders occur at NN426650 and NN427656.

Farther E trimlines on S245,S246 and S247 indicate a former ice-surface at _ca_ 630m O.D. Above that altitude frost-riven psammitic bedrock and numerous solifluction lobes and terraces occur intermittently on the upper slopes of the mountain ridge between 88 and 89.

In the vicinity of 92 a series of clear end and lateral moraines record the ice-margin of the former Rannoch glacier. At NN523662 an end moraine(but rock-cored at first) crosses the floor of a shallow valley at an altitude of _ca_ 530m O.D., climbs steeply to the NE to an altitude of 580m and then turns sharply to the E to become a lateral moraine. The lateral moraine forms a ridge generally only 1-2m high on the distal side, but reaching 4m in height on the proximal side. Many Rannoch Moor granite boulders are scattered across its surface. Eventually the moraine turns sharply to the SE and descends to the floor of the Allt Ghlas valley at 92 down to 530m O.D. to form an end moraine. On the distal side of the moraine the broad floor of the valley contains thinner deposits of till and the river is not incised. This contrasts with the thick till(>7m), numerous boulders and incised streams to be found within the limit.

The end moraine continues up the E side of the Allt Ghlas valley as a low ridge and finally peters out on the spur at NN539659. The ice-limit is next discerned on S252 at _ca_ 590m O.D. where numerous boulders and morainic ridges below the inferred limit contrast with smooth, debris-strewn slopes above the limit. Farther to the SE thick,dissected till to thicknesses of 7m infills a small valley and numerous morainic ridges curve around the hill of Meall Garbh(NN548639).

The ice-limits and evidence E of Meall Garbh are according to Thompson(1972) and demonstrate very close agreement with the present writer's own ice-limits to the W.

5.18 Ericht glacier

Some of the ice from the Rannoch Moor ice-cap flowed NE into the deep breach occupied by Loch Ericht. This is shown at 91 by striae, meltwater channels and some of the eskers that are directed toward the NE.

Patches of glacially-smoothed bedrock, numerous hummocky moraines and very extensive boulder fields, comprising Rannoch Moor granite, cover much of the area E of corries 234,235 and 236(W of 89). This suggests that local ice developed below the steep E-facing backwalls of the corries, combined with ice from Rannoch Moor to flow eastward into the Ericht trough. Local ice, generated below the steep rockwall W of 90(Figure 5.4) flowed south-eastward along the Alder Burn valley at 90. Along the eastern edge of the valley hummocky moraines and thick deposits of dissected till are replaced at altitudes higher than ca 600-650m O.D. by extensive solifluction sheets and lobes and by thick accumulations of frost-riven debris on the upper slopes of Ben Alder.

Distinct morainic ridges and low mounds mantle the slope NE of 89 to altitudes of ca 650-700m O.D. to indicate a former ice-margin declining in altitude eastward. Trimlines on spurs 247,250 and 251 range in value from 590 to 630m O.D. and give support to the eastward decline in the former glacier surface.

The ice-limit along Loch Ericht, however, is an estimate since the present writer has not mapped farther E than S250. Sissons has suggested(personal communication), based on the distribution of hummocky moraine along the loch sides, that the ice did not extend very far into the breach and, therefore, the ice-limit shown on Figure 5.5 is largely based on this premise.

5.19 <u>Lyon glacier</u>

Only the evidence and ice-limits relating to the N side of Glen Lyon will be dealt with in this thesis(Figure 5.5). Thompson(1972) showed the distribution of hummocky moraine and proposed limits for the Lyon glacier on a one inch O.S. map(sheet N^{o} 48) in his thesis. An additional area not mapped by Thompson but mapped by the present writer and shown on Figure 5.5 is Glen Daimh(location 100). Independent mapping of the distribution of hummocky moraine in Glen Lyon by the writer shows a very close agreement with the pattern depicted by Thompson, although a different ice-surface altitude and gradient is proposed.

The cluster of corries at the head of Glen Lyon (248-251,254-255 and 257-259) and a number of steep rockwalls(<u>e.g.</u> at NN359411 and NN375415) nourished ice that flowed into Glen Lyon. For example, W of 97 huge numbers of boulders are strewn on the slope below corries 248 and 249 implying a derivation from the corrie walls above. At 97 the valley is choked with deeply dissected till exceeding 8m in thickness and implies a similar derivation.

E of 97 hummocky moraine, sometimes forming mounds and ridges greater than 10m in height, extend almost continuously down upper Glen Lyon.

The upper limit of hummocky moraines and a number of upper ice-limit values(spurs 266,267,270,271,272 and 275) provide very consistent evidence for the surface of the ice mass reaching a maximum altitude of between 700-740m O.D in upper Glen Lyon and Glen Daimh. Four high-level cols(255-258) N of Glen Daimh with altitudes ranging from 650-707m O.D. display evidence in the form of low hummocky moraines that signify that ice spilled northward toward the Moor of Rannoch. In contrast to these cols that provide minimum values for the former height of the glacier surface cols 264,268 and 273, that all approximate to 745m O.D., lack such glacial evidence and impose maximum values on the altitude reached by the surface of the ice. These cols, together with col 261 at an altitude of 813m O.D., are characterised instead by frost-riven outcrops of rock and slopes smoothed by gelifluction processes.

The mapping of the distribution of periglacial evidence, especially along spurs, gives added support to the assumption that the ice-surface reached a maximum altitude of between ca 720-740m O.D. Ice-limit values mainly lie within a narrow range of 695-730m O.D.(Appendix A).

Thompson(1972,p.139) suggested that the ice-surface in upper Glen Lyon exceeded 760m O.D., but this value was based on extrapolation. The evidence presented by the writer and described above refutes this view and nowhere in Glen Lyon is the surface of the main trunk glacier likely to have been greater than ca 750m O.D. except in the corries and in some of the tributary valleys.

No striae or friction cracks(except for a few poor examples on C257) were located in Glen Daimh or in the part of Glen Lyon shown in Figure 5.5. Instead, the orientations of roches moutonnées have been used to determine former ice-flow movements, but only in areas where they were either very abundant or well-defined, since they can be less reliable that friction cracks(cf Rastas and Seppala,1981). The resultant pattern shows that the highest level of ice existed in upper Glen Lyon with ice flowing S toward Glen Orchy, N toward Rannoch Moor and E down Glen Lyon.

Ice-moulded bedrock and clearly-defined eastward-directed roches moutonnées along the spur at 101 demonstrate the eastward flow of ice farther down Glen Lyon. At 101 a series of shallow meltwater channels, cut into till, cross the spur at an oblique angle(Thompson,1972).

N of 101 numerous hummocky moraines and morainic ridges are spread across a tributary valley and E of a low rockwall. Their abrupt termination at 102 implies an ice-limit and is interpreted as such in Figure 5.5. Near 103 two steep-sided valleys are choked with till to at least 15m in thickness and in places contain morainic mounds. Ice-limits are drawn where the thick till and mounds terminate up-valley in each case. These ice-limits, mapped independently by the present writer, are virtually identical to those mapped by Thompson(1972). Hence no further fieldwork was undertaken farther E and the ice-limits shown on Figure 5.5 E of 103 are those according to Thompson.

Addendum

Since this thesis was completed additional field mapping in the Glen Duror area suggests a limit to the Duror glacier

alternative to the one proposed in section 5.6. An outwash surface occurs at Achara House (NM 988544) that slopes steeply down toward the W at the exit of Glen Duror. The steep eastern edge of the landform is taken as an ice-contact slope. If this interpretation is correct,and if the feature was not formed during decay of the ice-sheet, it suggests that the Duror glacier may have extended for <u>ca</u> 1½km beyond the limit described in section 5.6. On this basis it is possible that the Linnhe and Duror glaciers became confluent rather than remained independent as shown on Figure 5.2. If this assumption is correct the 'end moraine' at Auchindarroch farm (NN 001553) can be interpreted as a medial moraine formed at the junction of the two glaciers(see Figure 2.1 and p. 32). Thick till and a large morainic mound E of the motel at NM 994549 may also support the view that the limit of the Duror glacier was farther W than that shown in Figure 5.2. However, no firm conclusions can be suggested since levelling of the terraces in Glen Duror needs to be undertaken to ascertain the possible relationships between the former glacier limits and contemporary sea level and this has yet to be done.

PART 11

THE RECONSTRUCTED ICE MASS AND THE SPATIAL IMPLICATIONS OF THE EVIDENCE

CHAPTER 6

RECONSTRUCTING THE SURFACE FORM OF THE MAIN ICE MASS IN THE WESTERN GRAMPIANS AND COMPARISONS WITH SPITSBERGEN GLACIERS

6.1 Introduction

In section 6.2 the field evidence and ice-limits described earlier in this thesis are used to reconstruct the form and extent of the main ice mass and of some of the independent glaciers to a contour interval of 50m(Figure 6.1). The relationships between the reconstructed ice mass in the study area and published data relating to the remainder of the ice mass in western Scotland are briefly described and shown on Figure 6.3.

The techniques employed to reconstruct former glaciers using different types of field evidence have been outlined in many other studies(e.g. Ives,1957,1958; Løken,1962; Thompson,1972; Gray,1972,1982a; Gray and Brooks,1972; Sissons and Grant,1972; Sissons, 1972,1974a,1977a,1977b,1977c,1979a,1979b; Porter,1975,1977,1979; Pierce, 1979; Ballantyne and Wain-Hobson,1980; Cornish,1981; Thorp,1981; Meierding,1982; Porter and Orombelli,1982). However, few studies have attempted to relate the parameters of the reconstructed glaciers to the parameters of glaciers existing today. To establish some means of comparison attention is turned in section 6.3 to glaciers existing today, in locations analogous to those that existed in western Scotland during the Loch Lomond Stadial. Such modern analogues clearly require glaciers reaching tidewater along fiords from sources in mountains not exceeding 1400m O.D. One such area that fulfils many of the requirements is southern Spitsbergen. Six glacier parameters (namely length,area,volume,maximum thickness,mean ice-surface gradient and mean basal-ice gradient) of selected Spitsbergen and Loch Lomond Advance glaciers are compared using statistical techniques. Graphical methods are used to compare area/altitude distributions and the surface forms of the two separate populations of glaciers. The results and implications of such analyses are discussed in sections 6.4 and 6.5.

6.2 Reconstructing the former glaciers in the study area

The outlines of the former glaciers, when they were at their maximum extent, were reconstructed mainly by using the methods described in detail by Sissons(1974a). The delimitation of a number of glacier snouts at their maximal position has been described earlier in this thesis, based on the work of Gray(1972,1975a),Thompson(1972), and Sissons(1979b) and in a number of areas(e.g.lochs Rannoch and Treig, and the Strath of Ossian) accurate limits can be drawn back from the snout positions using lateral moraines. When reconstructing the form of the glacier snouts, the heights of the calving ice-cliffs of the glaciers terminating in tidewater(e.g. Etive and Linnhe glaciers) were assumed to be ca 30-60m after Flint(1971). Farther back from the terminal zones of the former glaciers, and especially in the former accumulation zones, trimlines and other forms of ice-marginal evidence (Appendix A) can be used to delimit glacier margins, but usually to a lower degree of accuracy compared with the limits based on moraines (see section 4.7).

Within corries it is often difficult to decide where the upper limit to the ice lay on the backwall since this is mostly roughened by glacial plucking and lacks clear evidence of an ice-limit. Sissons reconstructed ice-limits in corries by assuming that the ice surface was ca 30m below the top of the backwall(quoted in Gray,1982,p.127). However, as Thorp(1981b)and Gray have pointed out this may not always be a valid assumption since some glaciers only occupied part of their host corrie. Thus ice-margins within the corries were drawn taking into account the morphology of the corrie, the constraints imposed by the trimline values and the morphological evidence within the corrie. Reconstructing the corrie glacier was a fairly simple exercise where ice-moulded bedrock indicated occupance of the complete width of the corrie by ice, as in the deep corries along the Aonach Eagach Range(corries 52-58,Figure 5.2). In other cases only small glaciers, in relation to the size of the corrie, were nourished below the steep crags on their W side. This is often suggested by only small areas of ice-moulded bedrock on the corrie floor compared with extensive areas of frost-riven bedrock and accumulations of frost-riven debris, particularly on the south- and west-facing corrie sides. In addition a number of corries lack clear glacial or periglacial evidence or contain what is considered to be equivocal evidence and assumptions have been made about ice-limits in these corries(see p. 130). In many other areas in the Scottish Highlands where the inferred Loch Lomond Advance glaciers were relatively small,discrete glaciers(e.g. in the NW Highlands,the Cairngorms and the SE Grampians) the location of end and lateral moraines generally makes it readily apparent which corries nourished glaciers and which did not during the Loch Lomond Stadial. In much of the study area this type of evidence cannot be used to indicate which corries contained ice during the stadial as the glaciers comprised large transection glaciers that flowed for considerable distances from their main source areas(e.g.20-40km).

In a number of areas abundant striae and friction cracks enabled the surface form of the glacier to be drawn quite accurately,as these were taken as being perpendicular to the ice-flow direction. Such considerations apply particularly to upper Glen Nevis, the Loch Leven area and parts of glens Etive and Kinglass. Elsewhere the ice-surface contours were drawn taking into account source areas of ice, directions of ice-flow as indicated by roches moutonnées,stoss-and-lee forms, meltwater channels and trimline values(Chapter 4, p.86)

A surface concave down-glacier is assumed in the upper part of the glacier while a surface convex toward the snout is assumed on the lower, ablating segment of the glacier. Modern analogues are provided by present day glaciers in Spitsbergen, as for example Austre Torellbreen(Norsk Polarinstitutt sheet B12). This simple pattern, however, may not always have applied due to topographical factors where convergence and divergence took place and this was further taken into account when drawing ice-surface contours. For example, marked convergence of ice in Glen Etive toward the narrow gap, NW of Ben Cruachan between 50 and 55(Figure 5.3), is demonstrated by the spatial pattern of striae,friction cracks and roches moutonnées up-valley from the gap. The ice-surface contours are inferred to have been markedly concave down-valley in the vicinity of the gap(cf the flow of liquid in a venturi flow meter) with the ice-surface steepening sharply.

The location of interpolated ice-margins was facilitated in many areas by the contouring exercise since the form of the ground frequently imposed constraints on their location and helped to pin-point possible trimline errors.

An ice-shed was drawn in upper Glen Nevis on the

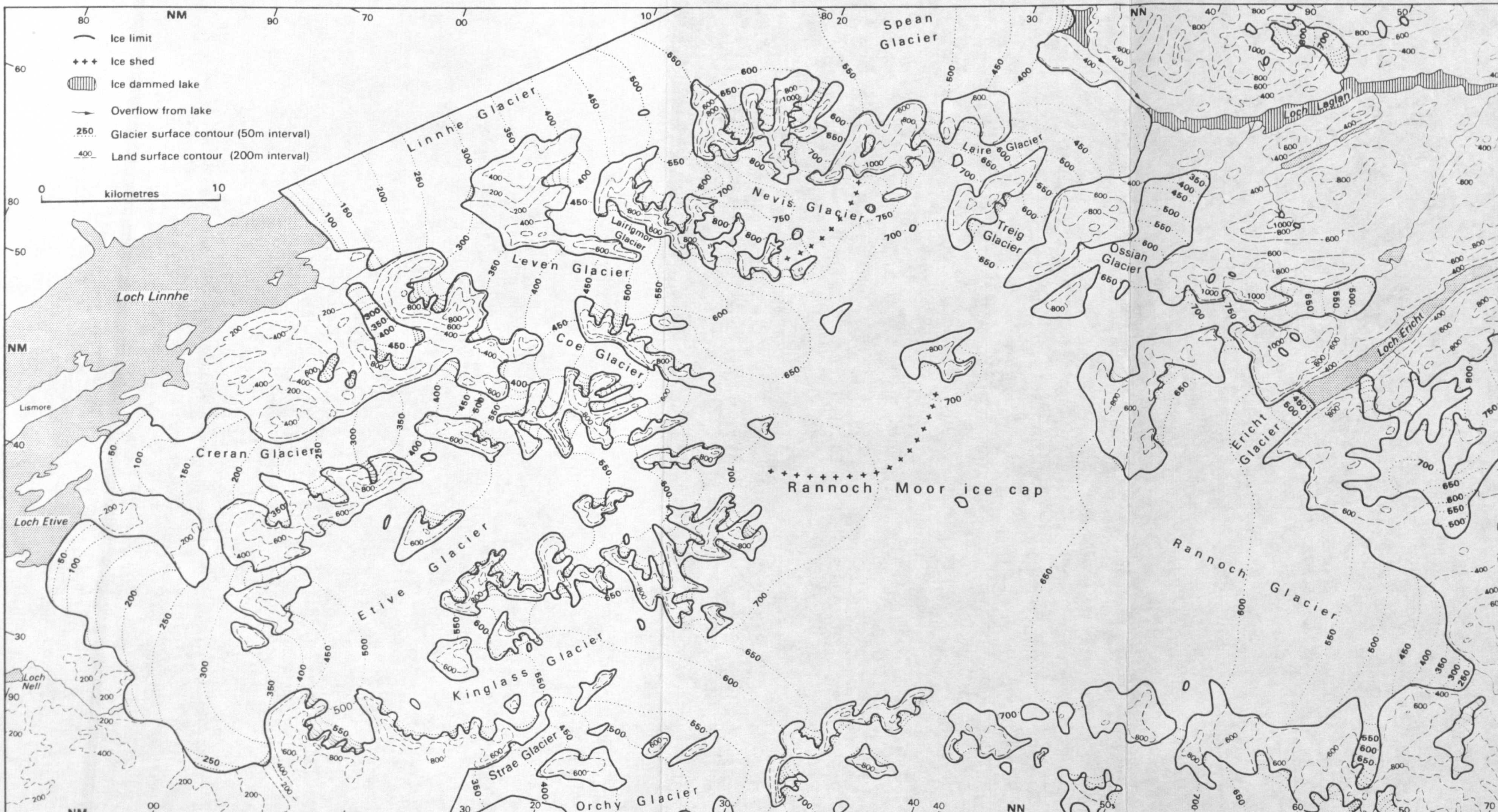


Figure 6.1 Extent and form of the stadial ice mass in the western Grampians.

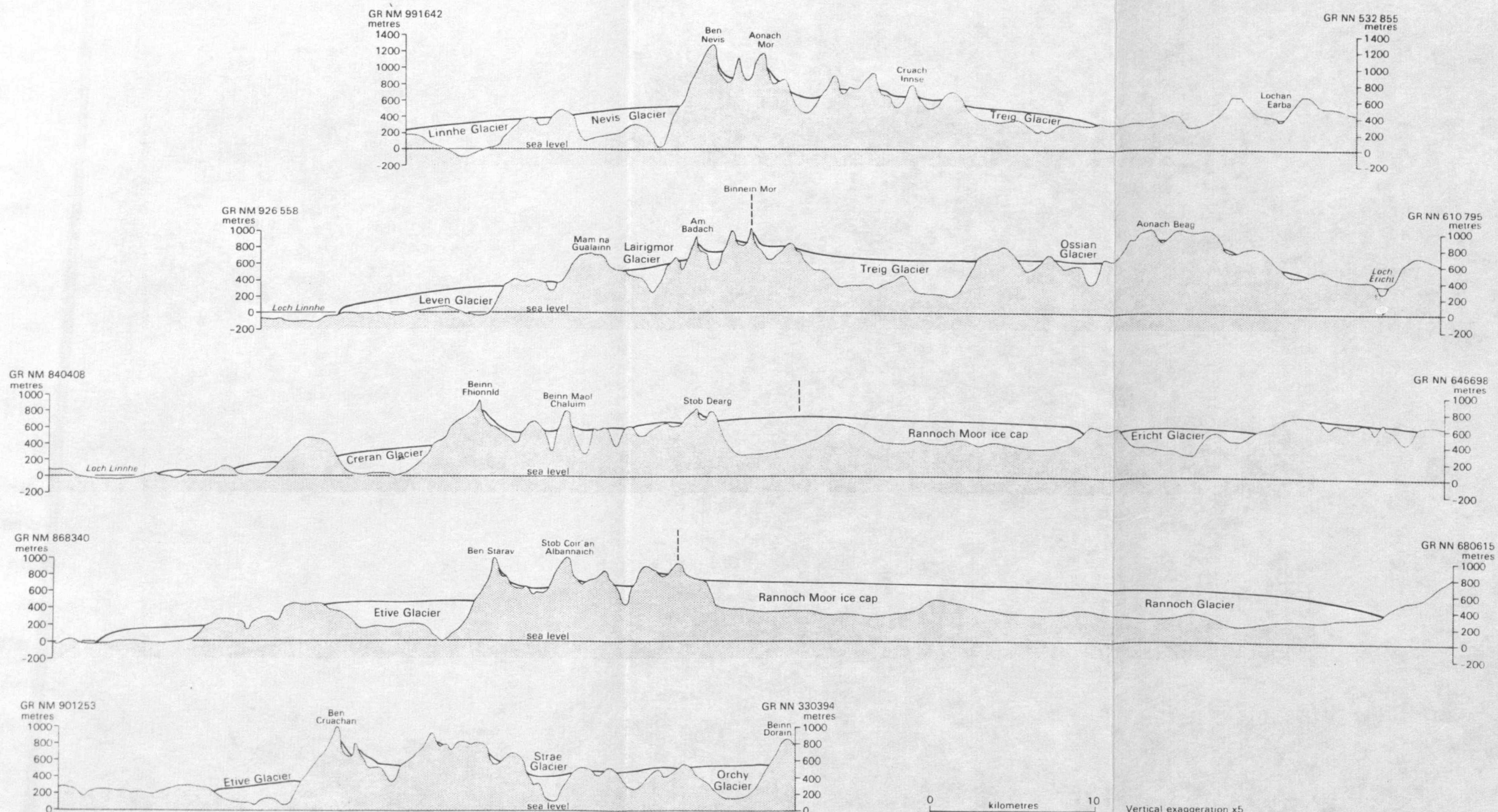


Figure 6.2 Sections across the ice mass in the Western Grampians drawn at intervals of approximately 9 kms in a W.S.W. to E.N.E. direction. Main ice shed shown by pecked line.

basis of diverging striae and friction cracks. The ice-shed shown in the NW of Rannoch Moor is probably less accurate because of a lack of a sufficient number of ice-direction indicators in this area. However, striae and friction cracks in the Leven-Coe area and in the area between Loch Treig and the Blackwater valley, and Rannoch Moor granite erratics to the N and eastward-directed roches moutonnées to the SE all impose strong controls on the location of such an ice-shed.

The final form of the large, reconstructed ice mass indicates that it attained a maximum width of 80km from NE to SW (Figures 6.1 and 6.2). Maximum ice-levels above 700m O.D. were attained in the Glen Nevis and Glen Lyon areas, and in the western part of Rannoch Moor - these areas fed ice to both the western and eastern sides of the ice mass. The views of Thompson(1972) and Sissons(1980) that the Rannoch Moor ice-cap probably reached an altitude of ca 850-900m O.D. are not upheld.

Rannoch Moor was filled with ice to levels mainly between 650 and 700m O.D. and formed an ice-cap ca 400km^2 in area. Ice from this source flowed sub-radially outwards by means of outlet glaciers down glens Orchy, Etive, Creran and Coe, and along the deep troughs occupied by lochs Leven, Treig and Rannoch.

The morphology of the former ice mass was strongly asymmetrical with the highest mass of ice on the NE side, whereas the outlet glaciers flowing to the SW were at much lower altitudes and flowed for considerable distances along fiords. In places the outlet glaciers descended steeply to the W, as in central Glen Nevis and in upper Glen Coe, where the ice flowed over steep rock bars. At its

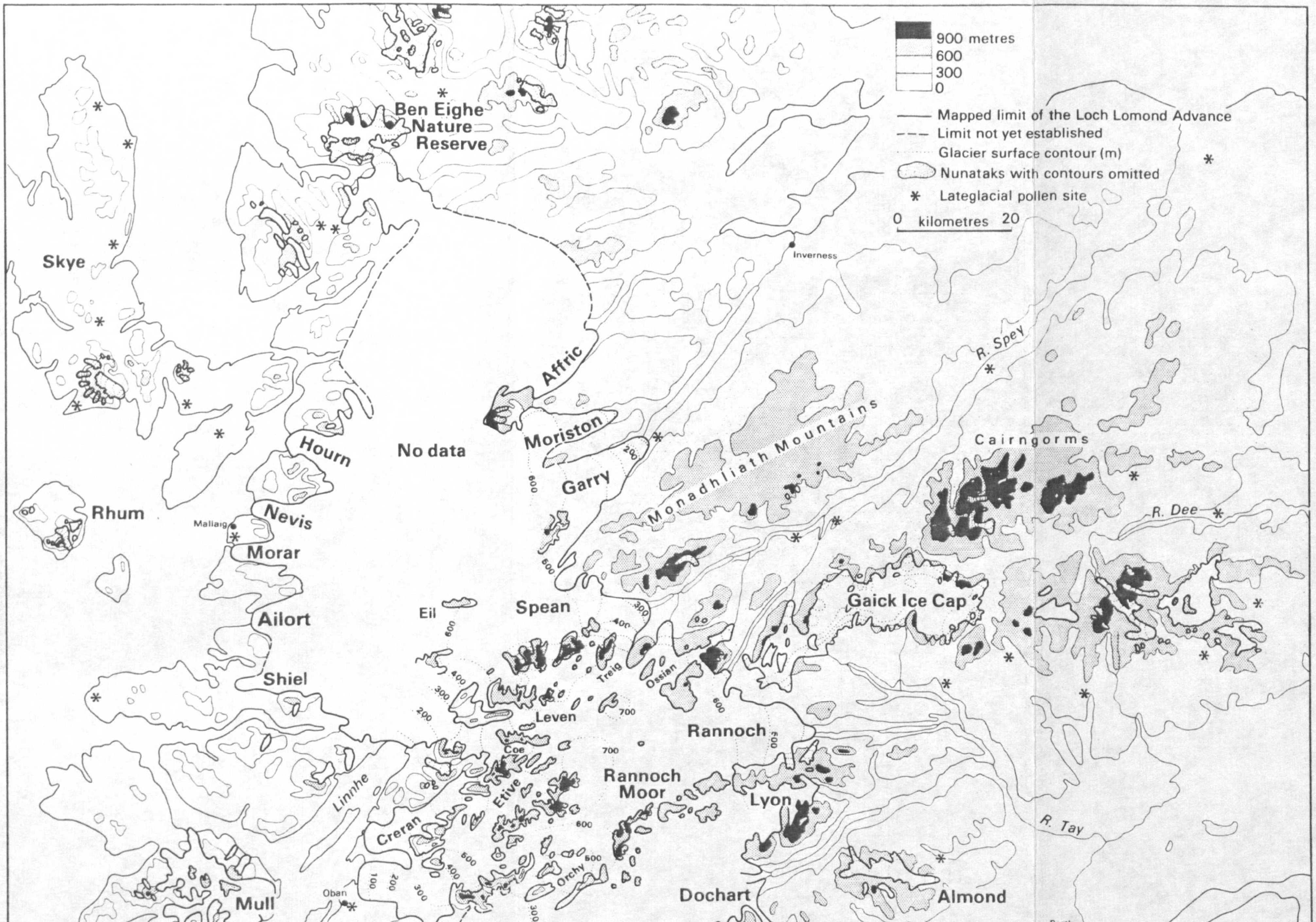


Figure 6.3 Limit of the Loch Lomond Advance in part of the Scottish Highlands. For sources see text p.150

maximum extent the ice mass contained over sixty nunataks, most of which existed in the Western Mountain zone.

In order to place the reconstructed part of the main ice mass in the western Grampians in a wider regional setting Figure 6.3 shows the ice-limits and reconstructed Loch Lomond Advance glaciers in much of the Scottish Highlands and Inner Islands,derived from various sources(Thompson,1972; Gray,1972; Gray and Brooks,1972; Sissons and Grant,1972; Sissons,1967,1972,1974a,1977a,1977b,1977c, 1979a,1979b; Robinson,1977; Ballantyne and Wain-Hobson,1980; Thorp, 1981b;Boulton et al,1981). Surface contours of some of the glaciers are omitted where data are insufficient for such reconstructions.

The distribution of the former glaciers demonstrates the importance of the large mass of ice that built up in the Western Highlands and the western Grampians and especially emphasises the study area as the heart area of the ice mass. The palaeoclimatic implications of such a distribution are discussed in detail in chapters 12 and 13.

6.3 Present day glaciers in Spitsbergen

Table 6.1 summarises some of the essential characteristics relating to Spitsbergen today and to the Scottish Highlands during the Loch Lomond Stadial.

The area of Spitsbergen that corresponds most closely with the topography of the western Grampians and which contains transection glaciers similar to those that are inferred to have existed during the Loch Lomond Stadial is southern Spitsbergen. The topography in southern Spitsbergen is partly Alpine in type with numerous corries and pyramidal peaks linked together by arêtes, but as in west-

	Spitsbergen (Present day)	Scottish Highlands (Loch Lomond Stadial)
Latitude	76 - 80°N	56 - 58.5°N
Width (km)	210	260
Length (km)	375	292
Area (km^2)	78,750	76,000
Highest peak (m)	1,656	1,344
Topography	Alpine in S and along W and E coasts. Plateau in N-central area.	Alpine in W. Plateau in E and NE.
Glacierised area (km^2)	24,539=	ca 10,000
Equilibrium firn lines (m)	180(SE) to 700+(N)*	ca 250(SW) to 1,000+(NE)+

= Derived from Macheret and Zhuravlev(1982)

* Based on firn line calculations for 153 glaciers in southern Spitsbergen(S of latitude 79°N)

\+ Derived from Sissons(1980)

TABLE 6.1 Selected data for Spitsbergen and the Scottish Highlands ern Scotland the mountains are deeply dissected by through troughs. Many troughs show progressive widening toward the sea and truncation of spurs by ice(Linton,1967). Mean values of the heights of mountain peaks vary from 775m O.D. in the SE, 790m O.D. in the S to over 900m O.D. in the SW. These values are similar to the mean height of peaks in the western Grampians(i.e. ca 900m O.D.). Farther N in central Spitsbergen the mean height of the mountains increases to over 1100m O.D.. The coast of Spitsbergen is highly indented by fiords with some such as Isfjorden extending inland for more than 80km(Figure 6.4).

Southern Spitsbergen experiences an Arctic Maritime climate with the northward flow of the North Atlantic Drift having an ameliorating effect on temperatures. Even so Isfjord station on the west coast has a mean annual temperature of -4.4^{o}C and is only ice-free for 3 to 5 months in the year(Rowan et al,1982). Mean annual precipitation is less than 300mm yr^{-1} with maximum amounts falling as snow in the winter months. Much of the winter precipitation is produced by cyclones emanating from the trough controlled by the Icelandic Low trending NE between Norway and Spitsbergen. Highest amounts of precipitation occur on the east coast with SSW to SSE snow-bearing winds being the most important.

The spatial distribution of the glaciers in the Svalbard archipelago is strongly asymmetric(Boulton,1979; Jonsson,1982; Macheret and Zhuravlev,1982) with the most extensive areas of ice occurring on the eastern side of the archipelago. Radio echo-sounding data(Macheret,1981) indicate that ice thicknesses steadily decrease from S to N in southern Spitsbergen and from E to W in northern Spitsbergen where conditions are extremely dry. Such a pattern has been explained by Macheret and Zhuravlev(1982) as reflecting the decrease

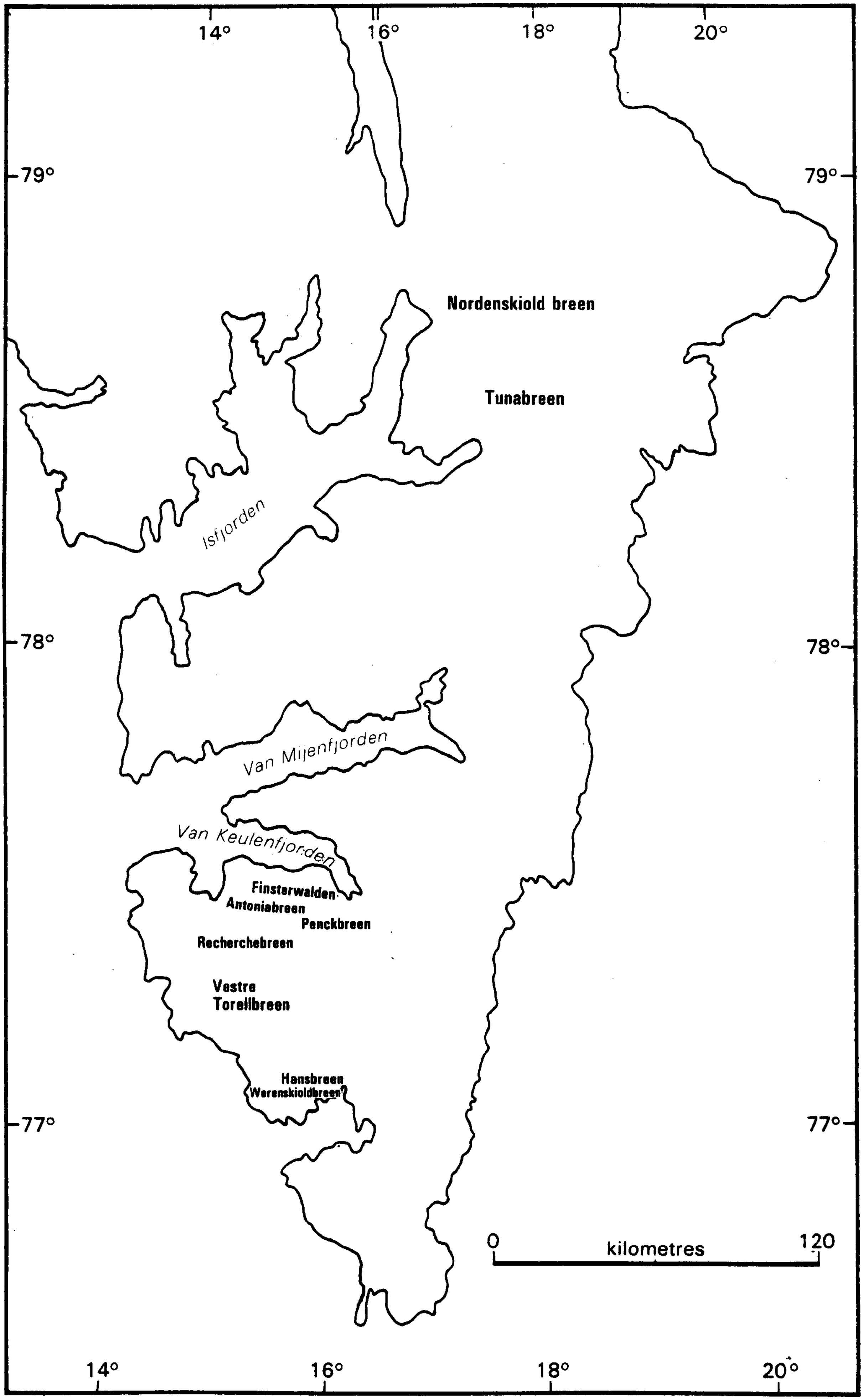


Figure 6.4 Location of the nine glaciers in Spitsbergen used in the comparisons with former glaciers in the study area.

in precipitation from the south and east coasts toward the NW part of Spitsbergen. Trend surface analysis of the firn lines of 153 glaciers in southern and north-central Spitsbergen by the writer (unpublished) indicates that precipitation decreases markedly inland from both the E and W coasts and this pattern is superimposed on the general trend described above.

6.4 Comparisons between Spitsbergen and Loch Lomond Advance glaciers

To compare the glaciers on Spitsbergen with the reconstructed glaciers in the western Grampians nine glaciers were selected from southern and central Spitsbergen(Figure 6.4) using three main criteria. Firstly, most of the glaciers needed to be selected from southern Spitsbergen where topographical conditions closely parallel those to be found in the western Grampians. Secondly, map coverage to scales of 1:100,000 was only available to the writer for southern and central Spitsbergen(S of latitude 78° 40'), except for the Royal Geographical Society map to a scale of 1:125,000 that covers part of northern Spitsbergen. Thirdly, data relating to glacier volumes and thicknesses had to be available. Such data could be obtained from Macheret(1981),and Macheret and Zhuravlev(1982) that to date provide the most comprehensive data relating to parameters of glaciers on Spitsbergen.

The nine glaciers and their main parameters are shown on Table 6.2. Lengths(L) and ice-surface gradients(S) were derived from the topographical maps using the methods shown in Figure 6.5. Lengths were measured from the glacier snout along the median line of the glacier either to a clearly demarcated ice-divide(*e.g.* the ice-divide between the Vestre Torellbreen and Paierlbreen on

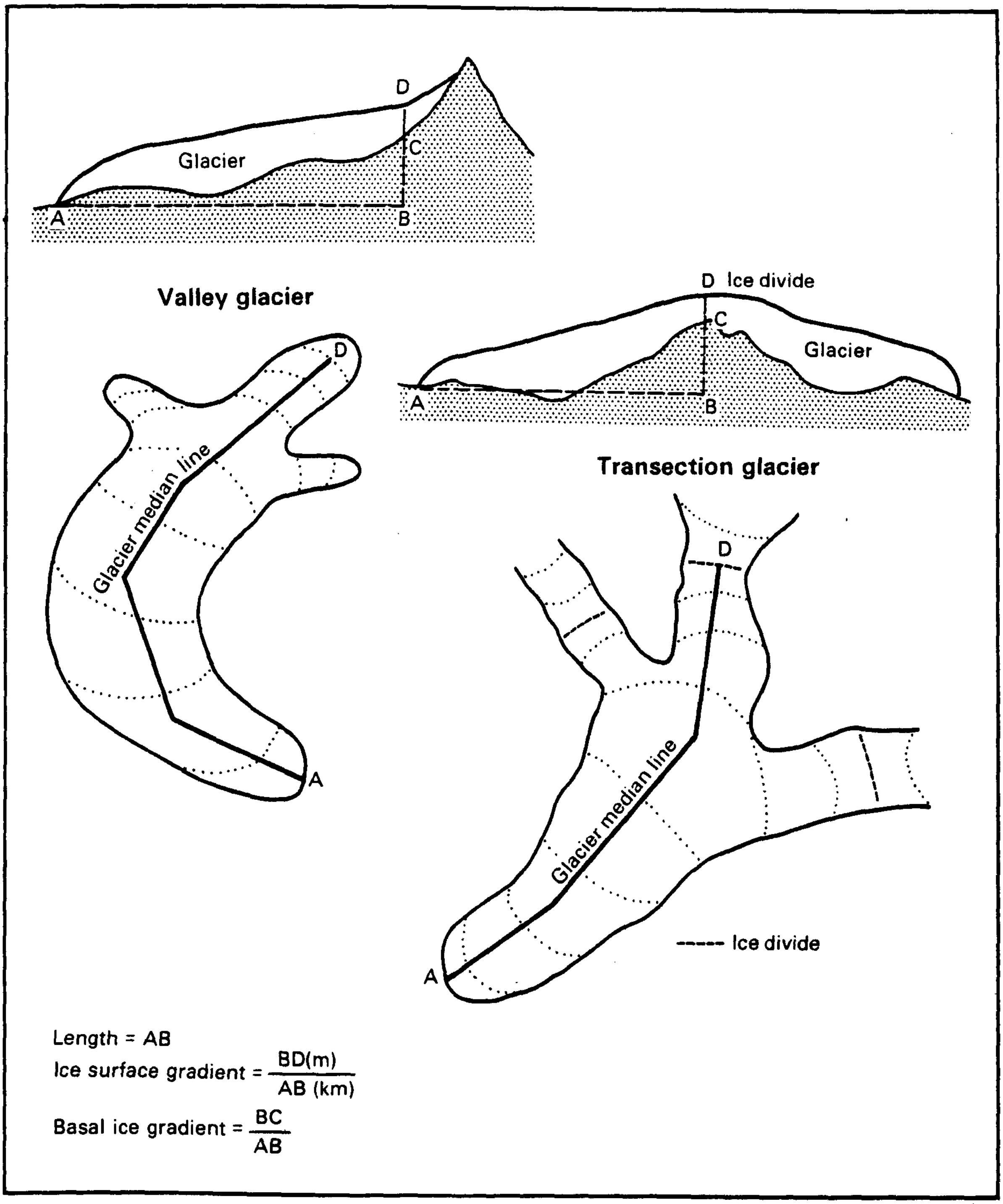


Figure 6.5 Methods used to measure selected parameters of Spitsbergen and former Scottish glaciers

Glacier	F (m)	L (km^2)	A* (km^2)	V* (km^3)	T* (m)	S (m per km)	B* (m per km)	V/A
Werenskioldbreen	331	8.0	34	3.0	320	62.5	8.75	0.0882
Penckbreen	415	21.0	83	8.85	415	29.05	15.39	0.1066
Finsterwalderbreen	417	11.8	38	2.8	175	55.09	32.48	0.0737
Antoniabreen	431	12.7	32	4.0	350	51.97	12.0	0.125
Hansbreen	303	17.0	72	11.91	330	25.53	23.4	0.1654
Recherchebreen	447	25.0	82	24.06	430	32.0	1.67	0.2934
Vestre Torellbreen	391	35.5	217	59.16	430	20.56	0	0.2726
Nordenskioldbreen	713	24.8	130	10.92	165	45.16	45.46	0.084
Tunabreen	677	33.0	77	3.56	125	30.91	24.75	0.0462

* Values derived from Macheret(1981) and Macheret and Zhuravlev(1982)

TABLE 6.2 Firn line(F), length(L), area(A), volume(V), maximum observed thickness(T), volume/area ratio(V/A), surface gradient(S) and the basal gradient(B) of certain glaciers in southern Spitsbergen.

Glacier	F (m)	L (km)	A (km^2)	V (km^3)	T (m)	S (m per km)	B (m per km)	V/A
Creran	432	40.6	167.55	29.1	360	17.73	5.67	0.1767
Etive	419	40.6	297.0	60.6	580	16.01	9.85	0.204
Leven	567	37.5	177.15	41.8	590	18.67	10.4	0.236
Nevis	586	33.7	83.2	16.0	510	22.85	13.65	0.1923
Coe	488	27.7	54.5	11.4	460	25.99	7.22	0.2092
Ossian	647	22.8	90.6	20.3	310	16.67	0.88	0.2241
Treig	617	22.0	158.7	30.3	500	22.73	6.36	0.1909
Rannoch	630	40.7	578.0	136.1	430	12.29	1.23	0.2355

TABLE 6.3 Firn line(F), length(L), area(A), volume(V), maximum thickness(T), volume/area ratio(A/V), surface gradient(S) and the basal gradient(B) of certain reconstructed glaciers in the western Grampians.

sheets B11 and B12) or to the corrie farthest from the snout in the main source area where the glacier surface begins to steepen sharply(e.g. Werenskioldbreen on sheet B12). Ice-surface and basal-ice gradients(B) were calculated from the difference in altitude between B and D, and B and C, divided by the length AB. Due to incomplete radio echo-sounding data for some of the glaciers, as a result of the scattering and absorption of electromagnetic energy in wet snow and firn in the accumulation areas(Macheret and Zhuravlev,1982), basal-ice gradients were only calculated for part of some glaciers (see Figure 6.8).

Areas(A),volumes(V) and maximum observed thicknesses(T) of the nine glaciers were derived directly from Macheret and Zhuravlev(1982, Table 1, p.307). Firn lines(F) for both the Spitsbergen and the former Scottish glaciers were calculated using the methods outlined in chapter 12. Length, ice-surface gradient and basal-ice gradient were calculated in the same way for the former Scottish glaciers as for the Spitsbergen glaciers. Area,volume and maximum thickness parameters for the Scottish glaciers were obtained using the methods described in chapter 12. The parameters for the former Scottish glaciers are shown in Table 6.3.

Nordenskioldbreen and Tunabreen in north-central Spitsbergen are thin,outlet glaciers descending fairly steeply to tidewater from the plateau ice-cap of Lomonosovfonna. Finstwalderbreen and Werenskioldbreen are discrete, valley glaciers that descend steeply from corries in their source areas(Table 6.2). The remaining five glaciers are primarily transection glaciers, characterised by many nunataks in their source areas. The considerable variations in size and shape of the nine Spitsbergen are reflected in the wide range of volume/area ratio(V/A) values from 0.0737 to 0.2934.

In an attempt to analyse the inter-relationships between the different glacier parameters some of the data in Tables 6.2 and 6.3 were subjected to simple regression methods. Tables 6.4 and 6.5 present the results of the regression analyses. Correlations significant to a level of at least 0.05 exist between the parameters of the Spitsbergen glaciers, except between ice-surface gradient and basal-ice gradient. This lack of correlation suggests that the surface slope of the Spitsbergen glaciers is largely determined by the shear stresses of ice(Nye,1952; Weertman,1961; Paterson,1981) rather than by the underlying slope of the ground. However, caution is necessary with this interpretation since these particular results are based on only partial data for the sub-relief of the glaciers(see above).

The positive correlations between length,area, volume and maximum measured thickness of the Spitsbergen glaciers (Table 6.5) show that close inter-relationships exist between these parameters as glaciers increase in size. Inverse correlations between parameters L,A,V and T and the surface and basal gradients of the glaciers indicate that such gradients generally decrease as glaciers increase in size.

The correlation coefficients for the Loch Lomond Advance glaciers indicate similar relationships between the glacier parameters, although at generally lower levels of significance than for the Spitsbergen glaciers. However, the positive correlations between thickness and basal-ice gradient and between length and basal-ice gradient(Table 6.5) are opposite to the results obtained

	T	L	A	V	S	B
T		0.813	0.479	0.443	-0.495	0.770
L			0.669	0.658	-0.553	0.224
A				0.982	-0.906	-0.774
V					-0.890	-0.743
S						0.505

TABLE 6.4 Correlation coefficient between maximum thickness(T), length(L), surface area(A), ice volume(V), ice-surface gradient(S) and basal gradient(B) of certain reconstructed glaciers in the western Grampians.

	T	L	A	V	S	B
T		0.803	0.684	0.772	-0.798	-0.853
L			0.851	0.854	-0.916	-0.745
A				0.930	-0.813	-0.764
V					-0.781	-0.698
S						0.357

TABLE 6.5 Correlation coefficient between maximum measured thickness(T), length(L), surface area(A), ice volume(V), ice-surface gradient(S) and basal gradient(B) of certain glaciers in Spitsbergen.

Correlation at the 0.01 level of significance

Correlation at the 0.05 level of significance

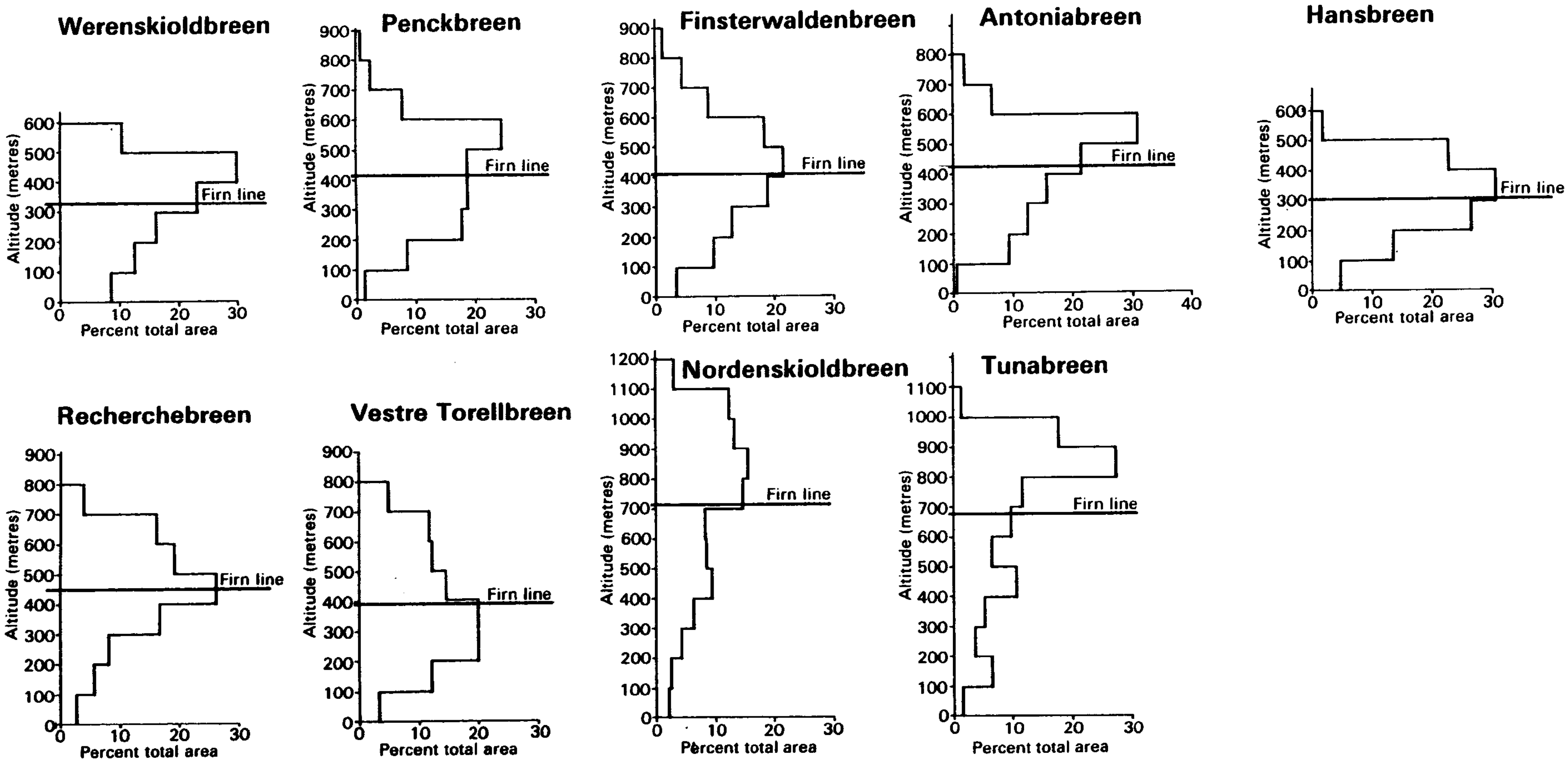


Figure 6.6 Graphs of surface area plotted against altitude of selected Spitsbergen glaciers

for the Spitsbergen glaciers. These apparent anomalies may be explained by topographical differences across the western Grampians. The outlet glaciers in the W(e.g. Creran,Etive,Leven,Nevis and Coe) were generally thicker and descended to lower altitudes than those outlet glaciers flowing E from Rannoch Moor(e.g. Ossian,Treig and Rannoch). The last three-named glaciers flowed across relatively gently-sloping ground in contrast to most of the glaciers in the W (Table 6.3). Thus the thickest and longest glaciers tended to have the steepest basal-ice gradients which is not in accord with the results of similar studies elsewhere(Buckley,1969).

Since the correlation coefficients between a number of the parameters of the Scottish glaciers were less significant,statistically,than for the Spitsbergen glaciers further comparisons were undertaken using data derived from the firn line calculations. These data were used to draw area/altitude graphs (Figures 6.6 and 6.7) to illuminate any major similarities or differences between the Spitsbergen and former Scottish glaciers.

The graphs for the Spitsbergen glaciers show approximately similar area/altitude distributions without any modal value exceeding 35% of the total area of the glacier. However, the graphs for the Scottish glaciers show two main groups with distinctly differing area/altitude distributions(Figure 6.7). Glaciers such as the Creran,Nevis and Coe flowing westward from sources in the Western Mountain zone show broadly similar area/altitude distributions to the nine Spitsbergen glaciers. In contrast the Leven,Ossian,Treig and Rannoch glaciers display very different area/altitude distributions with modal values ranging from 46 to 80%(Figure 6.7). Such characteristics reflect the importance of the Rannoch Moor ice-cap as the main

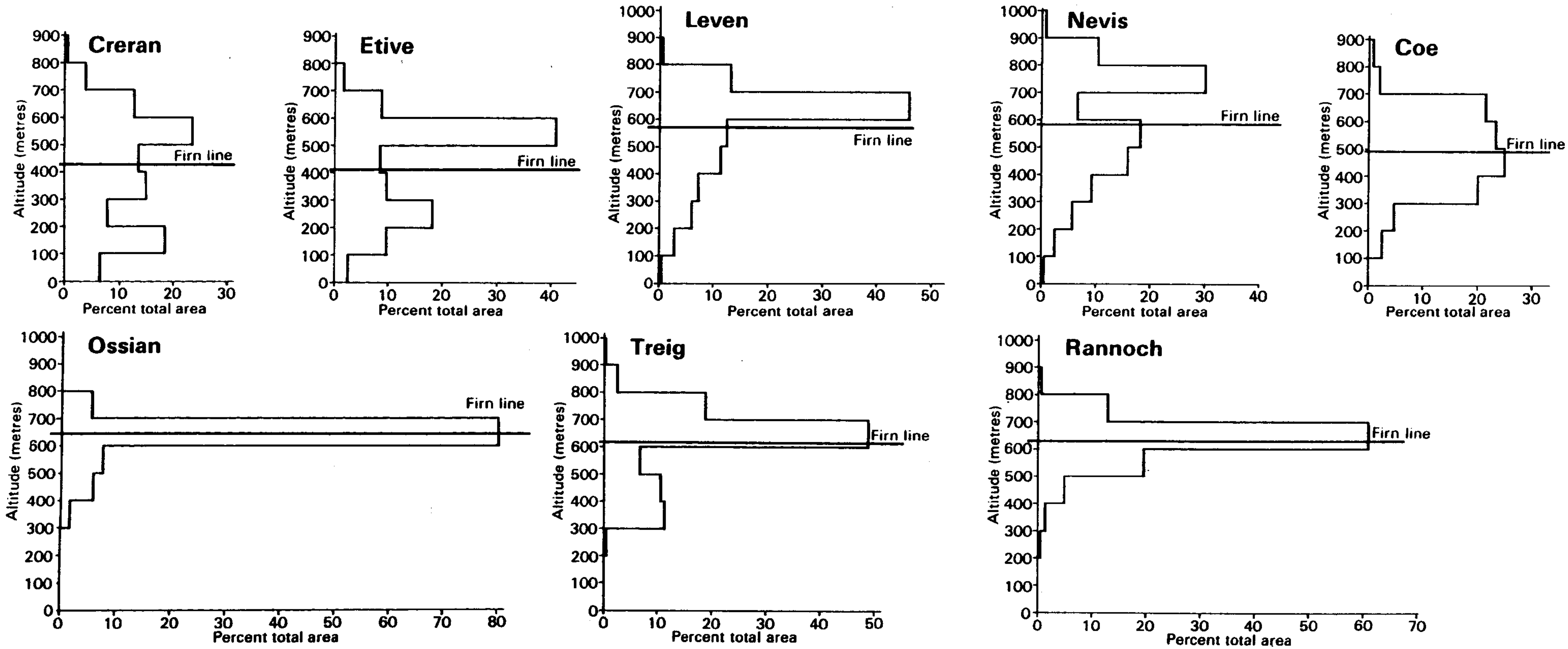


Figure 6.7 Graphs of surface area plotted against altitude of former Scottish glaciers

source area for the last four-named glaciers. The area/altitude distribution for the Etive glacier displays characteristics intermediate between the two groups. This may be explained by the basin-like form of Glen Etive NW of Ben Cruachan(Figure 1.2) that became filled with ice during the Loch Lomond Stadial.

The differences between the two main groups of glaciers in the western Grampians <u>may</u> partly explain why the correlation coefficients between some of the parameters of the former Scottish glaciers were not significant statistically. An added complication is that it is not possible to provide exact topographical analogues with the western Grampians, since no such area comparable with the Rannoch Moor basin exists in southern Spitsbergen.

Further comparisons between the Spitsbergen and former Scottish glaciers were undertaken by constructing surface and basal long profiles for each glacier along the median line using Norsk Polarinstitutt maps to a scale of 1:100,000 for the Spitsbergen glaciers and O.S. maps to a scale of 1:50,000 for the Loch Lomond Advance glaciers(Figures 6.8 and 6.9). The long profiles of the surfaces of the former Scottish glaciers, unlike the Spitsbergen glaciers, show a consistency of form that reflects the fact that the gradient and curvature of the glacier surface are largely determined by the plastic properties of ice once the glacier has reached a certain size and thickness. In addition the V/A ratios for the former Scottish glaciers show a close correspondence, with a mean value of 0.2086 and a standard deviation of only $\pm$ 0.018 compared with 0.143$\pm$ 0.083 for the Spitsbergen glaciers. These differences demonstrate that whereas the former Scottish glaciers were all transection or ice-cap outlet glaciers, derived from basically the same population,

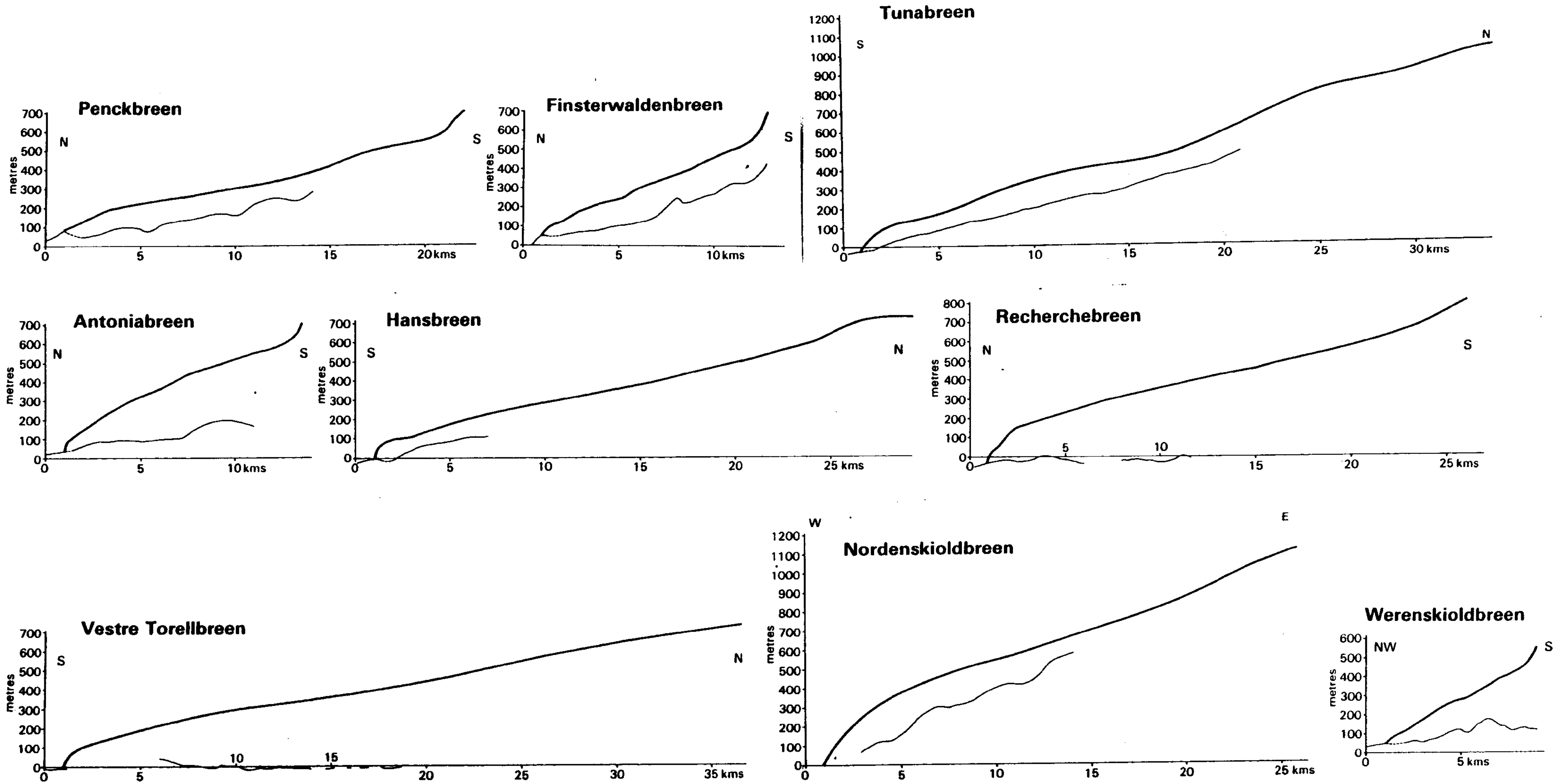


Figure 6.8 Sections of certain Spitsbergen glaciers drawn along the median line of each glacier. Sub-relief data derived from Macheret and Zhuravlev (1982).

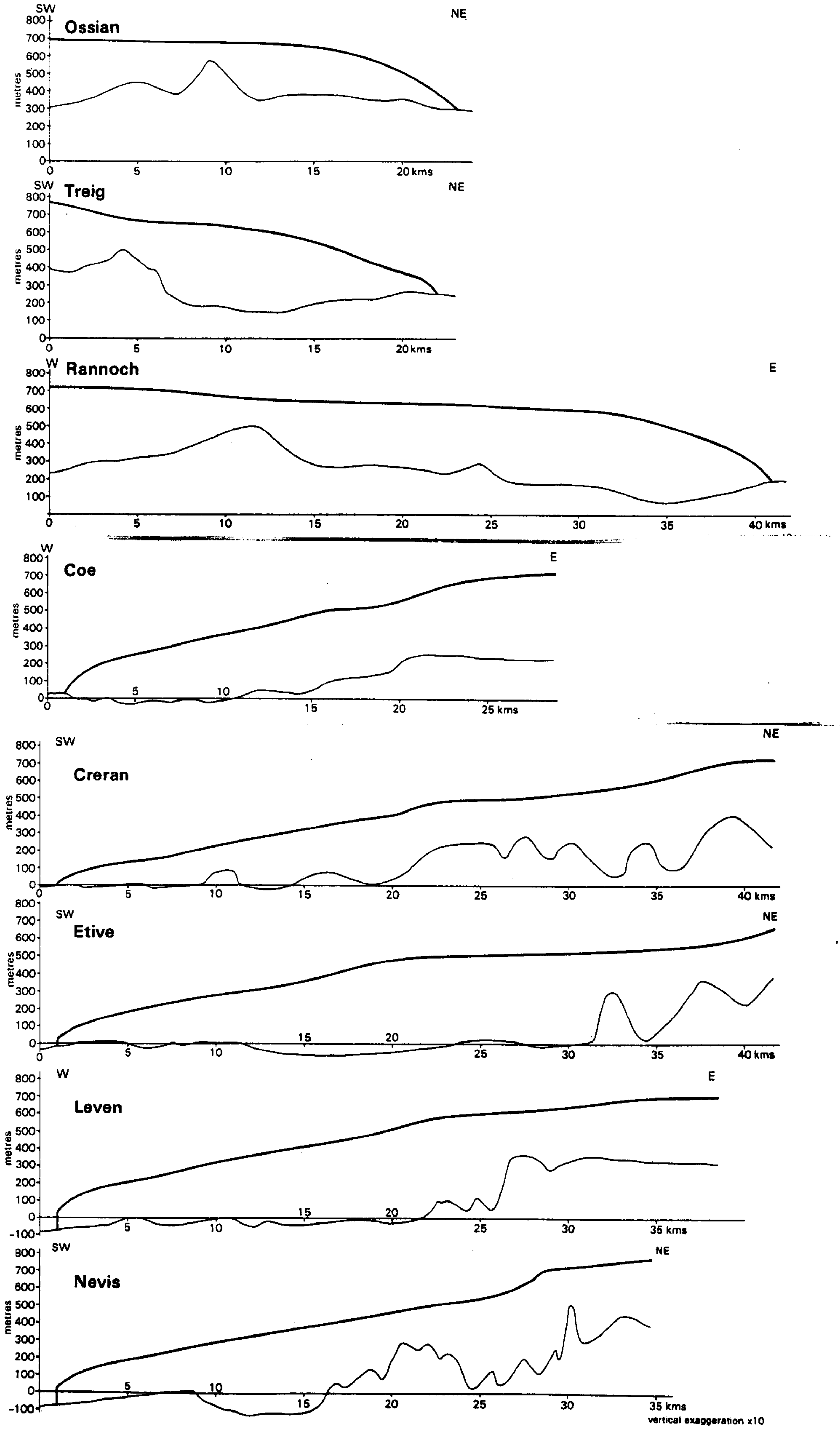


Figure 6.9 Sections of former Scottish glaciers drawn along the median line of each glacier.

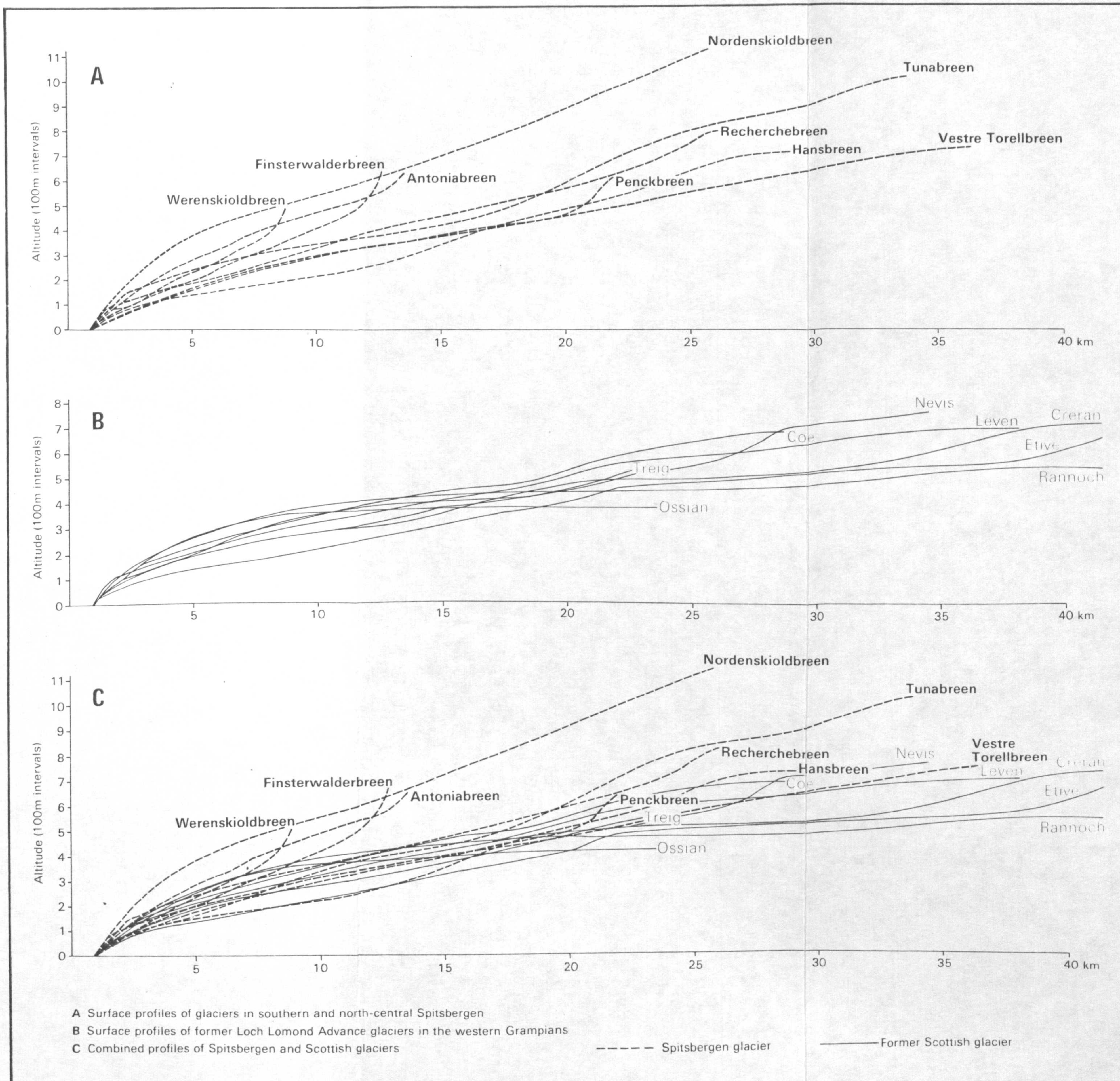


Figure 6.10 Superimposed profiles of Spitsbergen and former Scottish glaciers

the nine Spitsbergen glaciers comprise different types of glaciers from very differing populations.

Thus further attempts were made to compare the Spitsbergen and former Scottish glaciers. This was done by superimposing the long profiles of the glacier surfaces. Superimposed long profiles of the Spitsbergen and former Scottish glaciers are shown separately in Figures 6.10.a and 6.10.b and together in Figure 6.10.c. In order to compare the profiles it was necessary to reduce the altitude of each glacier snout to a common datum level, namely sea level. The influence of topography on the surface profiles of the plateau outlet glaciers(e.g. Nordenskioldbreen and Tunabreen) and the small, discrete valley glaciers(e.g. Werenskioldbreen,Finsterwalderbreen and Antoniabreen) is readily apparent. However, the surface profiles of the larger transection glaciers in southern Spitsbergen(e.g. Vestre Torellbreen,Recherchebreen and Hansbreen) show a close correspondence with the surface profiles of the former Scottish glaciers(Figure 6.10,c). This suggests that the inferred Loch Lomond Advance glaciers in the study area,reconstructed objectively from geological evidence, are compatible in form and extent with existing ice masses and that such data comparisons outlined in this chapter are of some validity when assessing glacier reconstructions based on field evidence.

6.5 Conclusions

The form of the ice mass reconstructed from many different lines of field evidence and depicted in Figure 6.1 is consistent with the hypothesis that such an ice mass relates to the Loch Lomond Stadial(Sissons,1979d). Support for this hypothesis is provided by the spatial distribution of radiocarbon dates within the study area(section 1.5) and from elsewhere in the Scottish Highlands(summarised in Sissons,1979d). No clear morphological evidence was located either on the nunataks or beyond the mapped ice-limits shown in Figure 6.1 to suggest that any other significant ice-limits relating to readvances or stillstands of the Devensian ice-sheet occurred in the study area(cf Robinson and Ballantyne, 1979 for the Wester Ross area of the Scottish Highlands).

Although present day glaciers on Spitsbergen and the former Loch Lomond Advance glaciers in western Scotland are not strictly comparable since the former have low activity rates and are mainly cold-based(Boulton,1972,1979) whereas the latter are inferred to have been highly active and warm-based, the results of the comparisons described in this chapter suggest that a sufficient degree of similarity of glacier form exists to validate such an exercise. However, a complicating factor that could have influenced such comparisons, that was ignored by the writer, is the change to the form of a glacier created by surging(Paterson,1981), particularly as it is known that a number of Spitsbergen glaciers have surged in the past (Liestøl,1969; Rowan et al,1982; Macheret and Zhuravlev,1982). Nor is it known whether the selected Spitsbergen glaciers are in a state of equilibrium, or are advancing or are retreating, while the possibility that the Loch Lomond Advance glaciers were subjected to rapid ice decay, before they had achieved a maximum equilibrium position related to 'average' Loch Lomond Stadial climatic conditions, could have occurred. Such factors could clearly influence the form of the glaciers, but were not considered in the above comparisons.

As more radio-echo-sounding data and other types of data become available from Spitsbergen it should be possible to use a greater number of transection glaciers comparable in size and in similar topographical situations to the former Scottish glaciers. This should enable closer correlations to be made between the Spitsbergen and former Scottish glaciers' parameters than was possible with the data available to the writer. If this assumption is correct it should be possible to use such glaciological data, not only to support glacier reconstructions based on glacial geological evidence as described in this thesis, but also to predict, as a first approximation, the likely extent and form of other former transection glaciers in areas of western Scotland that have yet to be mapped in detail.

CHAPTER 7

IMPLICATIONS OF THE SPATIAL DISTRIBUTION OF THE MORPHOLOGICAL EVIDENCE

7.1 Introduction

Examination of the spatial patterns of the differing forms of evidence described in chapters 2 to 5 and depicted in Figures 5.2 to 5.5 suggests that a number of the forms of evidence can be linked to variables such as bedrock lithologies and structures, topography, glacier dynamics and palaeoclimatic influences. This chapter attempts to demonstrate possible links between these factors and the mapped evidence. In order to show the distributions more clearly selected forms of evidence are isolated and shown on separate maps covering the whole of the study area. Glacial and fluvioglacial deposits, ice-scoured bedrock and the Main Rock Platform are shown on Figure 7.1, while distributions of erratics and distinct limits to erratic trains are depicted on Figure 7.3.

7.2 The distribution of end and lateral moraines

End and lateral moraines are best developed and most abundant on the E side of the main ice mass where about twenty examples have been mapped. Only about six such moraines have been mapped on the W side of the ice mass and these are generally poorly developed in comparison with those to the E. For example, in places the Treig glacier is delimited by triple and double end moraines that have classic textbook examples of steep proximal and distal slopes. In total the former Treig glacier is delimited by end and lateral moraines that extend for about 15km(Sissons,1979b). No comparable

Figure 7.1 Ice-scoured bedrock, glacial and fluvioglacial deposits and the Main Rock Platform related to the Loch Lomond Advance Based on published data: Gray (1975), Peacock (1971), Sissons (1979, 1980) and unpublished data: Gray (1972) Thompson (1972), and on the writers own mapping

clearly-defined features exist on the western side of the ice mass in the study area.

Several explanations that may or may not be interrelated can be offered to explain such contrasts in moraine development:

i) The three main glaciers in the W (i.e. Linnhe, Creran and Etive) all terminated in tidal water. If an allowance is made for a sea level 10-12m O.D. higher(relative to the land) than the present(Gray,1972), the maximum depth of the sea at the maximal limit of each glacier was respectively about 90m,15m and 30m. Significantly the only glacier to develop large end moraines was the Creran glacier. The former glacier appears to have stabilised in the vicinity of a number of islands and in shallow water at the seaward end of Loch Creran. This appears to have favoured the formation and preservation of the inferred double moraine. Such conditions did not apply to the Linnhe and Etive glaciers where rapid calving of ice would have taken place into Loch Linnhe and may have prevented the formation of an end moraine. Investigations of the sea bed in the vicinity of the inferred limits of the two glaciers might reveal evidence similar to that recorded by Boulton et al(1981) for the Nevis glacier on the W coast of Scotland.

ii) The glaciers flowing to the SW on the W side of the ice mass would have had a different glacial régime to those flowing to the E and NE. The SW-flowing glaciers would have been characterised by high accumulation rates, but this would have been counterbalanced by high ablation rates(Sissons,1979c). Thus abundant supplies of meltwater would have been available, even when the glaciers were advancing, to wash away a good proportion of the debris being transported by the glaciers. This may explain why the writer has not yet found a single, unequivocal lateral moraine in the western half of the study area. Good modern analogues of such contrasts in moraine development exist in the coastal ranges of Alaska,where on the NE side of the ranges end moraines are large,clear features,while on the SW Pacific side of the ranges they are poorly-developed or absent(Flint,1971).

The evidence described above is directly opposite to the distribution of end moraines mapped by Sissons(1977b,1977c) on Skye and in the NW Highlands. In these areas the largest end moraines are associated with glaciers that faced mainly between W and SE. This apparent contradiction is less significant when it is seen that such glaciers were very much smaller than those in the study area and would have generated much less meltwater. In addition all the smaller glaciers terminated on land and this would have favoured the growth and preservation of end moraines.

Since large volumes of meltwater are postulated for the SW side of the main ice mass, meltwater channels would be expected to be relatively abundant in the W. Yet this is true for only small areas in glens Creran and Etive(Figure 8.1). The reasons for this are not clear. One possible reason why meltwater channels are virtually absent in the Loch Linnhe and Loch Leven areas is that the sides of the two fiords are relatively straight and provided few topographic obstructions to the flow of englacial and subglacial meltwater. Meltwater channels are also absent in the terminal zone of the Etive glacier, but this could relate to much of the meltwater escaping N into Glen Creran(p.111) and S into the Pass of Brander.

iii) A possible factor to explain the paucity of clear moraines in the terminal zones of the main SW-flowing glaciers could be that the ice was clean and only carried minimal quantities of debris. Such an explanation has been used by Peacock(1970a,p.48) to account for the general lack of drift W of the main ice-shed in the West Highlands. However, this explanation is unlikely for the study area, except perhaps locally, since accumulations of till of more than 10m in depth occur at a number of localities in the W and these clearly imply that the ice was not clean.

iv) The Etive and Linnhe glaciers did not remain at their maximal limits for any length of time to form clear moraines(p.220).

7.3 The distribution of hummocky moraine

It is significant to note that hummocky moraines are especially abundant in many areas lying within the postulated glacial limits described in this thesis; limits that are mainly derived from trimlines in the glacier source areas and from end moraines or outwash landforms in the former ablation zones of the Loch Lomond Advance glaciers. Except for a few mounds in Glen Salachan in the vicinity of NM995515 well-defined hummocky moraines are conspicuously absent from areas lying outside the mapped glacial limits. Since the area inside the inferred limits in the study area is greater than 2,000km^2 this spatial pattern would seem to provide strong support for the hypothesis that well-defined hummocky moraines in the Scottish Highlands are largely a product of the Loch Lomond Advance(Sissons, 1974c).

Hummocky moraines are generally much more abundant E of the main ice-shed and in these areas are frequently found close to or at ice-limits at high altitudes, as in the Rannoch Moor,Glen Lyon, Glen Daimh and Strath of Ossian areas. In the W hummocky moraines tend to be more localised and the upper limits of such features are frequently more difficult to discern. Major exceptions include the low ground W of Loch Etive(NN055405) and the area SW of Loch Tulla.

7.4 The distribution of thick till

The distribution of thick till (>5m) within the ice-limits shown on Figure 7.1 reflects both the importance of glaciological and topographical factors. Thick sequences of tills appear to have accumulated in several differing situations as depicted schematically in Figure 7.2:

A) Where extra-large quantities of rock debris, derived either from corrie walls or valley sides, were transported and deposited by the glacier within a short distance of the source of the rock debris e.g. below corrie 248 in Glen Lyon and below corrie 29 in Glen Nevis.

B) Where the ice-flow direction was transverse to hollows and tributary valleys thick till was lodged on the floors of the valleys and on the up-glacier side of ridges, probably due to a reduction in glacier velocity as a result of compressive flow taking place. The lodgement process would have been aided by shearing within the basal layers of the ice(Boulton,1970,1971; Moran,1971). Good examples of this type of deposition occur at NN270690,in the area between Glen Nevis and Loch Treig,and at NN510613,a kilometre to the S of Loch Ericht.

C) Thick till often occurs at the confluence point of two glaciers where rock debris,transported supra-glacially and en-glacially along the lateral margins of the glaciers, combined to form either a medial moraine or a chaotic arrangement of mounds and ridges(Figure

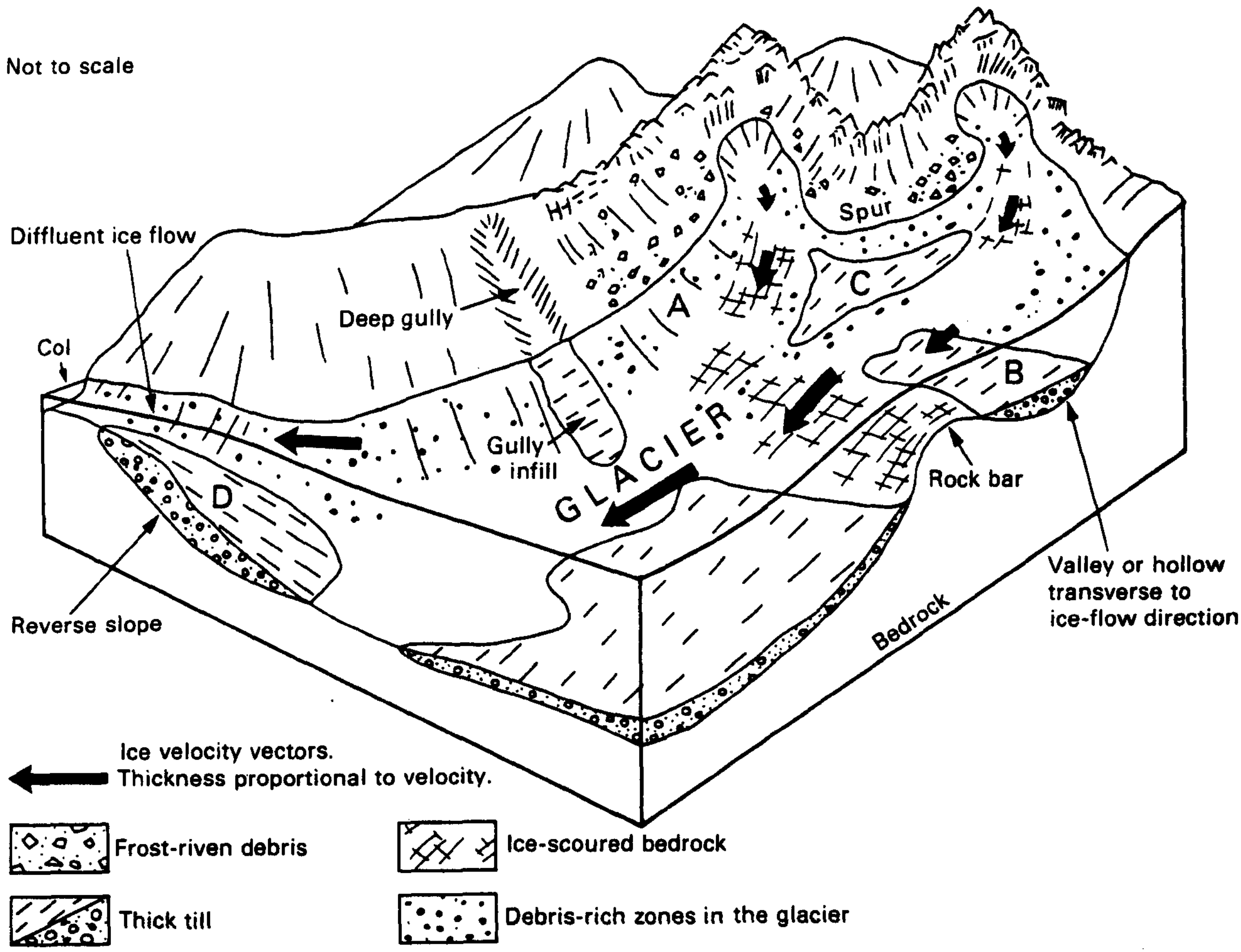


Figure 7.2 Depositional model for thick till within the limits of the main ice mass.
See text for explanation of letters A–D.

2.1). Such examples occur at NN105400 by Loch Etive, at NN121558 in Glen Coe and at NN149679 in Glen Nevis.

D) Thick till is frequently present where basal ice-flow took place up a reverse slope, such as where a lobe of ice pushed up a valley or where diffluent ice-flow took place across a col. This situation tended to produce some of the thickest till sequences to be seen in the study area(*cf* Sissons,1979b, for the Glen Roy area). Where ice-flow occurred across a col thick till is generally found on the up-glacier side of the col *e.g.*at NN470470 by Loch Daimh and at NN262383 SW of Loch Tulla. Less frequently, thick till can accumulate on the down-glacier side of the col as at NN445450 S of Loch Daimh. Especially noteworthy,however, are the substantial quantities of till to be seen in the Rannoch-Lyon area(Figure 7.1) where ice-tongues pushed into tributary valleys.

7.5 The distribution of exposed ice-scoured bedrock

The spatial distribution of strongly, ice-abraded bedrock exposed at the surface indicates a very noticeable concentration in the Western Mountain zone(Figure 7.1). Although fresh-looking, ice-abraded bedrock is likely to be present in many other localities in the study area,as demonstrated by recent construction work (see p. 43 this thesis; Gray and Lowe,1982; Gray,1982b),such bedrock is largely concealed by superficial deposits or by water bodies.Since the identification of trimlines partly depends on the mapping of areas of exposed, ice-abraded bedrock there is a strong correlation spatially between the distribution of trimlines and this type of bedrock surface. Therefore, it is proposed to deal with the implications of the distributions of exposed, ice-abraded bedrock surfaces and trimlines simultaneously. For example, Table 7.1 shows that trimlines (Appendix A) are especially dominant in the Western Mountain zone, where areas of exposed, ice-abraded bedrock are far more abundant.

TABLE 7.1 Analysis of types of trimlines and other ice-limits on spurs

	West of main ice-shed		East of main ice-shed		Totals	
	N°	%	N°	%	N°	%
A	107	68.5	25	52.0	132	64.7
B	4	2.6	4	8.3	8	3.9
C	3	1.9	8	16.7	11	5.4
D	4	2.6	2	4.2	6	2.9
E	8	5.1	2	4.2	10	4.9
F	2	1.3	2	4.2	4	2.0
G	16	10.3	4	8.4	20	9.8
H	4	2.6	0	-	4	2.0
I	2	1.3	1	2.0	3	1.5
J	6	3.8	0	-	6	2.9
	156	100.0	48	100.0	204	100.0

For explanation of letters see Appendix A.

The reasons for the existence of extensive areas of exposed bedrock surfaces, that have been strongly abraded by glacial processes, can be mainly ascribed to the following topographical and glaciological factors operating in the Western Mountain zone:

i) Relative relief is at a maximum with numerous peaks exceeding 1,000m O.D., whilst the floors of adjoining glacial troughs are below or just above sea level _e.g._ Loch Leven, Glen Coe, GLen Creran and Glen Etive. Hence a great deal of bare rock is exposed on the precipitous mountain slopes.

ii) Many of the valleys cut through the mountains are narrow and steep-sided or have sections that narrow to form steep, rock-walled constrictions. The higher glacier velocities inferred to have occurred at these topographical constrictions can be correlated with extensive areas of beautifully ice-polished bedrock. For example, excellent, ice-moulded rock surfaces occur in Glen Nevis(Figure 5.2, S of spurs 5 and 7), in upper Glen Coe(Figure 5.2, vicinity of 21), at the head of Glen Creran(Figure 5.3, near spur 101) and in upper Glen Etive(Figure 5.3, near spur 136).

iii) The glaciers flowing to the W were characterised by high accumulation and ablation rates(Sissons, 1979c) and hence by high velocities(_cf_ with glaciers at the present time in similar temperate areas such as western New Zealand, western Norway, western U.S.A and southern Iceland). The net result, both in ice-sheet times and during the more limited ice cover of the Loch Lomond Advance, has been to scour large areas of bedrock clear of thick debris.

The relatively few trimlines identified to the E of Rannoch Moor and on the lower ground to the W of the Western Mountain zone can be ascribed to generally less steep slopes, decreasing glacier velocities and to the fact that most of these areas would have been within the ablation zones of the glaciers, with deposition of the debris from the glaciers generally becoming much more important.

7.6 The distribution of fluvioglacial landforms within the ice-limits

The main fluvioglacial landforms that have been identified within the limits of the Loch Lomond Advance are primarily confined to the terminal zones of the main outlet glaciers. For example, most of the major outwash spreads and kame terraces of the Etive, Creran and Linnhe glaciers occur within the last 10-14km of these particular glaciers(Figure 7.1).

A similar distribution applies to eskers, except for a system of eskers on the S side of Rannoch Moor at NN415482. Thus clearly-defined fluvioglacial forms are absent from a very large part of the study area, formerly occupied by the main ice mass.

Thompson(1972) recorded a similar distribution of fluvioglacial forms within the Loch Lomond Advance limits in the area extending from the Rannoch valley southward to the Callander area. Although the source areas have yet to be mapped a similar distribution would seem to apply to the areas occupied by the former Moriston and Garry glaciers(Sissons,1977a,1979b).

The localised restriction of major fluvioglacial landforms to the terminal zones of the large Loch Lomond Advance glaciers did not apply to the last ice-sheet. Particularly extensive suites of outwash landforms, kame terraces and esker systems occur in many parts of the central and eastern Grampians and extend as far as the E coast of Scotland(Sissons,1976; Clapperton and Sugden,1977; Young,1978). This would seem to imply that the stagnation of the last ice-sheet took place under different glaciological conditions to those that applied to the glaciers of the Loch Lomond Stadial. Exactly how these conditions differed has yet to be determined (Sissons,1979d).

Furthermore, the fluvioglacial features within the Loch Lomond Advance limits tend to be small features in comparison with the often massive features associated with ice-sheet decay(Young,1978), although there are important exceptions such as the large outwash spreads at Corran Ferry and Connel Bridge. For example,all the eskers within the Loch Lomond Advance limits in the study area are only small features, rarely more than 4m high,compared with the large eskers mapped outside the limits(Sissons,1967).

Most of the meltwwater channels that have been mapped in the study area are small features,usually no more than a few metres in depth or greater than 300m in length. For example, channels with vertical, rock-walled sides, but only 2-3m deep and 3-5m wide,and cut across spurs or across the floors of cols,occur on the S side of Glen Etive(NN098398 and NN085358), on the N side of Glen Etive(NM988369), S of Loch Creran(NM984435) and E of Loch Creran(NN026464). These demonstrate highly localised areas of fluvioglacial erosion,as a result of the obstruction of the free flow of englacial or subglacial meltwater by topographical barriers.

The great majority of the channels demonstrate a flow of meltwater in the same direction as ice-flow movement,as illustrated by the striae and friction cracks in their immediate vicinity(Figure 8.1). An example of this type of close relationship between ice-flow and meltwater-flow directions is exemplified by a series of meltwater channels S of Loch Etive in the vicinity of NN070350. This type of relationship implies that the majority of the channels were cut during the later stages of growth of the ice mass to its maximum extent and/or during the early stages of deglaciation.

7.7 Ice-limits and former sea-levels

In many places along Loch Linnhe there is an extremely well-developed raised shore platform and cliffline. Originally the platform was believed to be Holocene in age(Wright, 1911). McCallien(1937), however, suggested that the Holocene epoch was too short to allow the necessary erosion and instead proposed a pre- or inter-glacial age. More recently, Sissons(1974b,1974c) suggested that the raised shoreline was cut in lateglacial times, during the Loch Lomond Stadial, as a result of frost-riving and other periglacial processes. Detailed levelling work by Gray(1978) has shown that the platform is tilted down toward the SW and the pattern of deformation and the gradient accords with Sissons' view that the platform, termed the Main Rock Platform, formed during the lateglacial period.

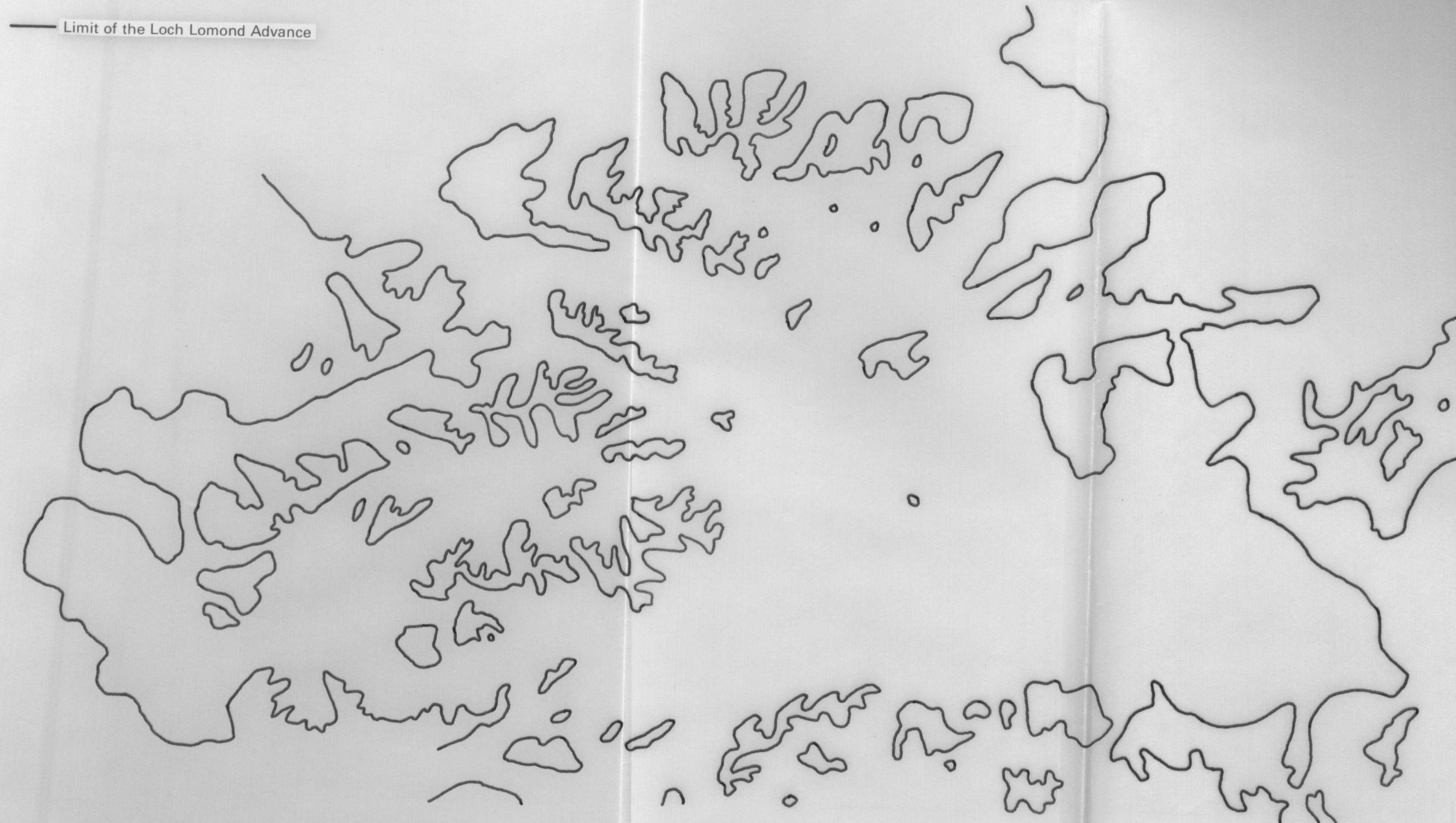
In the Loch Etive area Gray(1972,1975a) has tentatively correlated laminated deposits, either deposited in sea water or in an ice-dammed lake at an altitude of 10-12m O.D.,with a sea-level at about 10-13m O.D. These deposits are regarded as forming during the Loch Lomond Stadial. In addition Gray(1975a) has suggested that the main outwash spreads at lochs Etive and Creran were related to a sea-level below 12-14m O.D. In this area the Main Rock Platform lies at about 10-12m O.D.(Gray,1972).

Although the Main Rock Platform along Loch Linnhe has yet to be levelled N of Shuna(NM915490) the relationships between outwash deposits, believed to have formed during the Loch Lomond Stadial, and the platform are believed to be of some significance in the vicinity of Kentallen(NN008573). At NM977569 a possible narrow, rock platform at an altitude of ca 10m appears to lie beneath deposits of one of the outwash fans. If this interpretation is correct this demonstrates that the fans postdate the formation of the rock platform. Furthermore, a possible cliffline occurs on the landward side of the fans at a number of places. This suggests that the fans were banked up against a pre-existing cliffline. Since similar steep cliffs associated with a rock platform occur to the SW along Loch Linnhe(Gray,1972,1978) for many kilometres then it is assumed that the platform represents an extension of the Main Rock Platform northward along Loch Linnhe.

Furthermore, just beyond the westernmost outwash fan(Figure 7.1) is a clear rock platform and low cliff that extends more-or-less continuously for a kilometre to the SW at an altitude of ca 10m and is devoid of any glacial or fluvioglacial deposits. Beyond the headland at NM962556 a rock platform occurs in places while a cliffline is almost continuous along the edge of Loch Linnhe, without any sign of clear glacial or fluvioglacial landforms,until the entrance of Loch Creran is reached.

Therefore, the evidence cited previously suggests that the outwash fans formed at the end of the Loch Lomond Stadial and were deposited on the pre-existing Main Rock Platform, at or close to the maximum limit of the Linnhe glacier.

Within the ice-limits shown on Figure 7.1 clear rock platforms are difficult to find although a raised cliffline is present in many places(Sissons,1974b). Exceptions include the rock platform at 10-12m O.D. described by Peacock(1977) S of the jetty at Corran Ferry and the narrow rock platform that exists farther S to NN014616. Elsewhere within the ice-limits the rock platform is often mantled by deposits(Gray,1972; Sissons, pers.comm.) and is far less


Limit of the Loch Lomond Advance

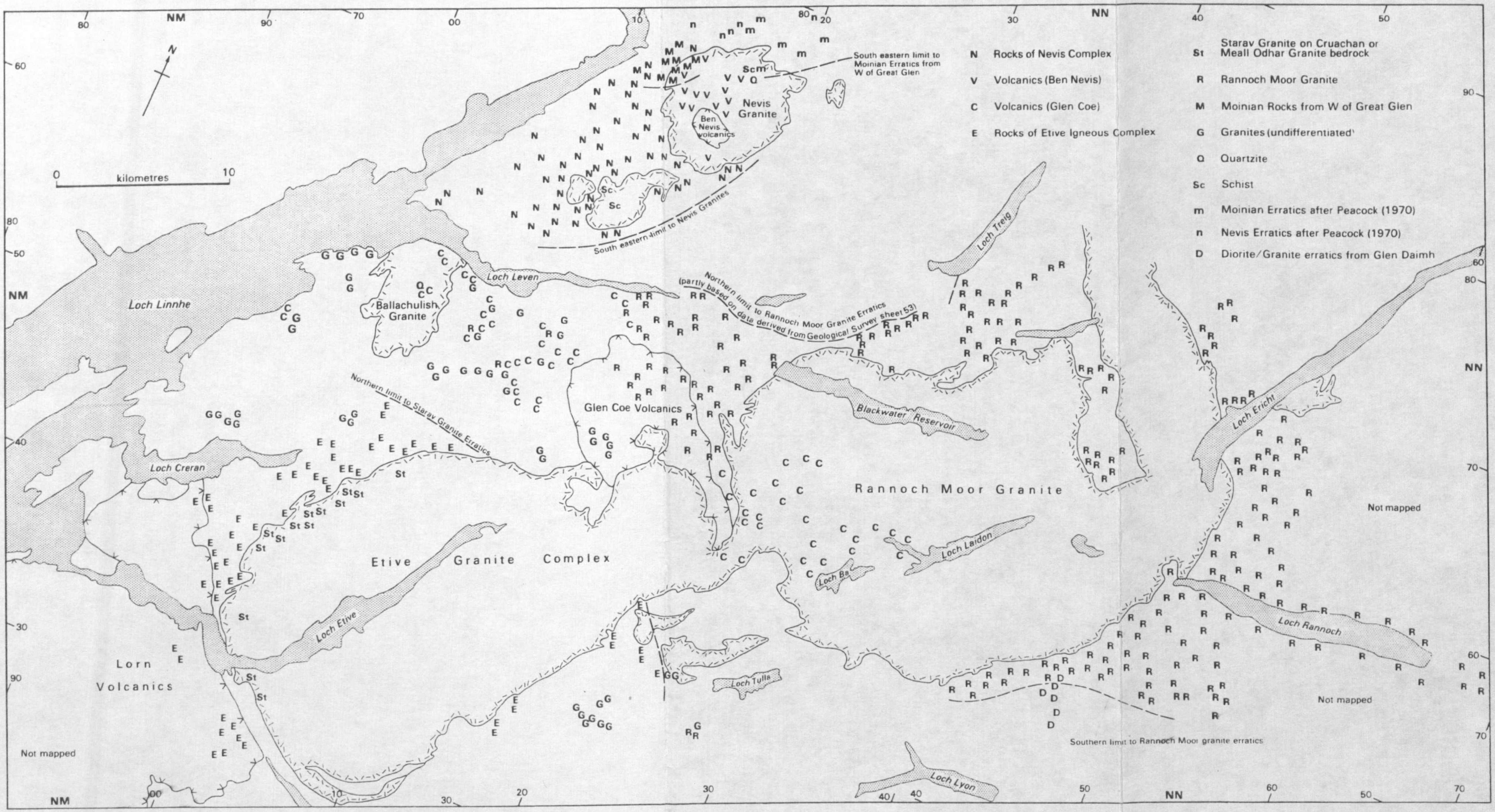


Figure 7.3 Distribution of selected erratics in the study area

Limit of the Loch Lomond Advance
kilometres
NM
NN
Loch Leven
Loch Linnhe
Etive Granite Complex
Rannoch Moor Granite
South eastern limit of Moinian Erratics from W of Great Glen
Starav Granite on Cruachan or Meall Odhar Granite bedrock
Rannoch Moor Granite
Moinian Rocks from W of Great Glen
Quartzite
Moinian Erratics after Peacock (1970)
Nevis Erratics after Peacock (1970)

distinct than outside the limits. This would appear to give added support to the ice-limits described earlier in this thesis.

7.8 <u>The distribution of erratics</u>

The distribution of erratics(Figure 7.3) provides strong supportive evidence for the reconstruction of ice-limits and former ice-flow directions in some parts of the study area.

The northern limit of Rannoch Moor granite erratics mapped by Officers of the Geological Survey(Bailey <u>et al</u>,1960) has been confirmed and extended both to the NE and to the W during the present survey. This limit clearly demarcates ice flowing W from the Blackwater-Rannoch Moor area, from ice derived from the Mamore Forest Range. Mapping of the same granite erratics in the area S of Loch Treig also confirms Sissons'(1979b) observation of a distinct western limit to such erratics in that area. Their distribution implies that a more powerful flow of ice took place from the Mamore Forest Range, than from the Rannoch Moor area,toward the Treig glacial breach.

A sharp change from local Nevis erratics(schists, quartzites and psammites) to Moinian-type erratics at the N end of Glen Nevis demonstrates the confluence of ice from Glen Nevis with ice from W of the Great Glen and again provides an indication of the more powerful of the two streams of ice.

The eastern limit of 'Ben Nevis' granites(taken to include Mullach nan Coirean granite) in Glen Nevis and the Lairigmor valley helps to support other forms of evidence, such as friction cracks and roches moutonnées, of ice-flow toward the W down these valleys.

In some areas a lack of certain exotic erratics (albeit negative evidence) would appear to justify certain conclusions. For example, the absence of erratics that are 'foreign' to glens Nevis, Daimh and Lyon appears to confirm other evidence that these glens maintained higher ice-levels during the Loch Lomond Stadial and were important centres of ice-dispersal to surrounding areas of lower ground. That the ice in Glen Lyon and Glen Daimh was at a higher level than the ice in the adjoining part of Rannoch Moor is demonstrated by a complete absence of Rannoch Moor granite erratics in the two glens.

Furthermore, repeated traversing of the ground for a distance exceeding 12km, from the glacial breach(NN391451) W of Loch Daimh to the lateral moraine(NN550518) S of Loch Rannoch, indicates a relatively sharp change southwards, from ground covered with numerous Rannoch Moor granite erratics to ground covered by locally-derived psammitic boulders(Figure 7.3). This implies that ice derived from Glen Lyon via the glacial breach and from the high ground surrounding Loch Daimh was powerful enough to deflect Rannoch Moor ice to the NE. Such transfluent ice-flow movements from Glen Daimh are supported by an erratic train of dioritic and granitic boulders, extending from their bedrock sources on the S side of the col at NN465482, northward into Rannoch Moor(Figure 7.3). Conversely, farther E beyond the peak of Meall Buidhe(931m) in the vicinity of NN513508 the spread of Rannoch Moor granite erratics within, but not outside, the limit defined by the end moraine at NN514501 demonstrates that ice from the Rannoch Moor ice-cap was more powerful than locally-derived ice spilling over the col at NN522487.

7.9 The distribution of periglacial features

The absence of well-developed periglacial features over large areas of the study area, even though extensive areas of suitable bedrock lie above altitudes of 400 to 500m O.D., implies that either the features were unable to develop because of the pre-existing presence of ice or they were destroyed as the glaciers developed. For example, the distribution of inferred fossil screes, large stone-banked solifluction lobes and terraces and severely frost-riven bedrock show important spatial variations. In the W of the study area thick vegetated scree and strongly frost-riven bedrock occur at altitudes as low as 400-500m O.D. as demonstrated by frost-riven quartzite bedrock and thick fossil screes S of Loch Leven(NN 070574) and in Glen Strae(NN 195335). However, in the Rannoch Moor area no such unequivocal inactive scree or severely frost-riven bedrock was mapped below altitudes of ca 600m O.D. In upper Glen Nevis and in upper Glen Lyon no periglacial features were encountered below altitudes of ca 700m O.D.

Such observations only become explicable when they are related to the form of the reconstructed ice mass described in this thesis and particularly if they are related to glacial limits inferred from glacial evidence. For example, in the Loch Ossian area the mountain slope immediately above the lateral moraine at NN 365686 is covered by a thick accumulation of frost-riven debris, that in places has moved downslope to form solifluction lobes and terraces. This evidence contrasts with the fresh appearance of the till ridges and hummocky moraines below the glacial limit that indicate a lack of major periglacial activity during the Flandrian.

However, in a few localities the relationships between the periglacial and glacial evidence is not always so clear. One notable exception of large stone-banked lobes occurring within the Loch Lomond Advance limit occurs to the N of Ben Nevis(NN 163740). Here amorphous lobes with risers of 1-3m descend obliquely westwards across the slope down to altitudes of less than 600m O.D. Since this is below the inferred glacial limit at ca 640m O.D.(Sissons,1979b; p.101 this thesis) this would seem to negate their use for helping to determine the maximum upper limit to the former Loch Lomond Advance glaciers. However, a possible explanation lies in the steepness of the mountain slope; this steepens sharply toward Ben Nevis and hence this may have favoured continued solifluction following deglaciation.

It was inferred in chapter 2 (p. 52) that smooth debris-strewn slopes were basically fossil periglacial features and thus they should only occur outside the limits of the Loch Lomond Advance. However, in places they descend below the glacial limits, derived from other forms of evidence, and limited movement since deglaciation is therefore inferred. This inference is supported by Ballantyne's(1981) suggestion that 'debris slopes' formed initially during the Lateglacial but have since been modified by microgelivation, wash,deflation and near-surface frost creep.

In a few localities as, for example, at the head of Glen Nevis(NN 226658), S of Loch Treig(NN 285648) and NE of Meall a' Bhuridh(NN 257514) small solifluction terraces with risers of less than 1m, that are inferred to be active at the present time, occur within the limit of the Loch Lomond Advance, implying that minor periglacial activity has continued through the Flandrian(White and Mottershead,1972). Yet such occurrences within the glacial limits described in this thesis are surprisingly uncommon and explanations to explain their paucity are clearly needed.

Sissons(1974c, p.110) has suggested that over much of mainland Scotland the lower limit of present day periglacial activity is probably between 450 and 600m O.D. In the study area many small turf-banked solifluction terraces, that are likely to be active at the present time, do exist down to these altitudes, although they are most abundant at higher altitudes. Yet except for a few examples, including those quoted above, they are almost entirely confined to areas outside the Loch Lomond Advance limits, even though considerable areas within the limits are at altitudes greater than 500-600m O.D.

Since the solifluction terraces and lobes that are inferred to be active at the present time largely coincide with bedrock areas that are severely frost-riven then a strong correlation with areas of well-drained frost-riven debris is implied. An explanation for the general lack of such active solifluction forms within the Loch Lomond Advance limit could be that the composition of the glacial materials is not generally conducive to the development of solifluction processes and landforms. In contrast solifluction terraces in particular are very apparent in areas outside the limits of the Loch Lomond Advance where abundant supplies of fossil frost-riven debris are readily available.

CHAPTER 8

IMPLICATIONS OF THE DISTRIBUTION OF STRIAE AND FRICTION CRACKS

8.1 Introduction

In chapter 3 the methods for identifying and mapping striae and friction cracks were outlined and the problems of inferring correct ice-flow directions from such features were discussed. In this chapter emphasis is given to the reasons for the spatial variability of the glacial markings and to the broad directions of ice-flow they show in the study area.

8.2 The field data

Figure 8.1 depicts all the locations where Officers of the Geological Survey and the writer have mapped glacial markings. The data in Tables 8.1 and 8.2 were obtained by placing a transparent overlay, showing glacial markings and the limits of the Loch Lomond Advance, over a base map depicting the solid geology for the area. A grid of one kilometre squares enabled the area of each rock type to be calculated inside and outside the limit. All calculations were checked for gross errors by obtaining two separate sets of calculations, related to latitudinal and longitudinal directions across the mapped area. Mean values were obtained from the two sets of data and these are the final values shown in Table 8.2. Prior to this, grand totals were checked against independent calculations of the total area, excluding all areas covered by water. Final differences were less than 1.5%. Densities of marking clusters

Figure 8.1 Ice flow direction indicators in the study area

Rock type	NW quadrant		SW, NE and SE quadrants		All quadrants		
	Inside ice limits	Outside ice limits	Inside ice limits	Outside ice limits	Inside ice limits	Outside ice limits	Totals
Volcanic	17	4	19	25	37	29	66
Granite	22	2	158	21	180	23	203
Quartzite	201	21	25	28	226	49	275
Schist and slate	91	10	14	22	105	32	137
Limestone	4	0	0	7	4	7	11
Flags and psammites	69	1	43	32	112	33	145
Totals	405	38	259	135	664	173	837

TABLE 8.1 Summary statistics for all observed glacial marking clusters in the study area

per 10 km^2 were obtained using the formula:

$$\frac{\text{frequency of glacial markings}}{\text{area of rock type}} \times 10$$

It must be emphasised that the total of 837 shown in Table 8.1 refers to clusters of markings and not to individual markings. Thus the arrows shown on Figure 8.1 can only reflect crudely the actual number of glacial markings observed in the study area. For example, the large number of markings that occur on quartzite means that they are under-represented by the data in comparison with the clusters of markings recorded on other rock types. This is exemplified at a number of localities such as on the backwalls of the two corries at NN 209706 and NN 170672, and below the high-level col at NN 267729 at the head of Glen Nevis, where crescentic fractures, striae and chattermarks occur in numbers exceeding ca 100 - 150 on exposed bedrock areas of ca $50m^2$ at each locality. Because of the limitations of scale these are represented by only three or four arrows on Figure 8.1. Yet in many localities with similar-sized areas of exposed rock surfaces, as on granite and volcanic bedrock in upper Glen Coe and on the N side of Ben Nevis, clusters of markings rarely exceed 10 - 20.

The spatial distribution of observed striae and friction cracks can be seen to be highly variable, both inside and outside the ice-limits shown on Figure 8.1. Some of the factors responsible for the variability are :

(i) Variations in rock type across the area (see section 8.4)

(ii) A lack of data from some areas. For example, observed glacial markings are very sparse in the NE quadrant, partly due to a lack of published data and partly because the writer has only mapped in parts of this area. Other areas where a detailed mapping programme has not been undertaken include the N side of the Ben Nevis Range, the Spean valley, parts of the Blackwater valley and much of the area between lochs Ericht and Rannoch.

(iii) The amount of exposed bedrock and as a corollary the extent of superficial deposits. For example, the large areas of exposed ice-scoured rock surfaces that exist in the Western Mountain zone (Figure 7.1) make it possible to record more glacial markings than in areas farther E where abundant superficial deposits mask much of the bed-rock.

(iv) The construction of tracks, roads and railways through the area that expose unweathered bedrock for scrutiny as, for example, in the area between lochs Laidon and Rannoch. In this area striae mapped by Officers of the Geological Survey appear, in many cases, to correlate with the route of the Highland railway and tracks cut across the Moor, suggesting that these relate to unweathered granite areas exposed during construction work at the turn of the century.

8.3 <u>Friction cracks related to topography</u>

Previous research has shown that friction cracks are particularly common on the stoss sides and crests of rock hill-ocks (<u>e.g.</u> Lahee, 1912; Harris, 1943; Okko, 1960). More recently, however, Gray and Lowe (1982) noted that crescentic gouges occurred on a slope dipping at 55° in a down-glacier direction at Marchlyn Mawr in Snow-donia which is not in accord with previous observations.

In terms of micro-topography friction cracks were observed in the western Grampians in a wide variety of topoographical situations ranging from flat,horizontal surfaces to vertical surfaces. Numerically, however, they are frequently most abundant on flat or gently-sloping surfaces either in an up- or down-glacier direction. They are particularly abundant if the surfaces occur in a series of steps that are gently inclined at an angle to ice motion. The lithological and structural conditions most favourable for the formation of this topographical situation occur where smooth, joint-bounded areas of quartzite crop out at an oblique angle to the ground surface. Frequently the crestal edge of the joint-block has been scalloped by numerous crescentic fractures where chips of rock between the fractures were removed by the ice or have been subsequently removed by weathering; as shown by trails of crescentic fractures that terminate at the crestal edge in the form of small, shallow hollows. Numerous examples of this type of topographical situation exist, but the description of one locality will suffice.

Col 210 at NN 267729 in upper Glen Nevis has been eroded into mica-schist but immediately to the E the bedrock changes to quartzite that crops out in the form of ice-abraded ledges with steep, joint-bounded edges. Large numbers of friction cracks, particularly crescentic fractures, exist on the gently-inclined rock surfaces and indicate a flow of ice to the E. Ice overtopping the col had descended the slope to the E and basal debris had clearly come into contact with the quartzite, especially where the edges of the steps projected upwards into the ice.

Friction cracks were observed on a number of very steeply-inclined rock surfaces $>70^{\circ}$. For example, crescentic fractures occur on near-vertical, joint-bounded blocks of quartzite on the lower part of the backwall of corrie 16 at NN 209706 while a particular fine series of crescentic gouges occurs on a steeply-inclined surface of Starav granite near the crest of col 161 at NN 142416.

On a macro-scale friction cracks were observed in many different topographical situations. In corries, although friction cracks exist on several backwalls including those of corries 16,36 and 145(Figure 11.1),they are most numerous on ice-moulded rock exposures on corrie floors and lips such as those in corries 2,3,18,52,72,108,111,113 and 114.

Many friction cracks are to be seen on exposed rock surfaces on valley floors and valley sides, particularly where rock knobs protrude above the general ground surface, although they are at their most abundant on spurs and ridges transverse to ice-flow direction as in upper Glen Nevis at NN 226686 and NN 242684 and on the N side of Glen Kinglass at NN 195426 and NN 175401.

However, there are many localities in corries, on spurs and on valley sides and floors that lack friction cracks, even though extensive areas of ice-moulded rock surfaces are present and similar glaciological conditions are likely to have occurred. Possible explanations of such variability will be considered later in sections 8.4 and 8.5.

8.4 Striae and friction cracks related to rock type and weathering

Friction cracks have been observed on a wide range of rock types. These include quartzite, limestone, shale, coarse sandstone, dolerite, granite, migmatite-granite, granite-gneiss, gabbro, ignimbrite, tuff, rhyolite, andesite and porphyrite (Lahee,1912; Harris,1943; Okko,1960; Thorp,1981a; Gray and Lowe,1982). There appears to be reasonably close agreement that crescentic fractures in particular are most abundant on the fine-grained and medium-grained siliceous rocks(Harris,1943; Virkkala,1960; Thorp,1981a). Significantly Harris(p. 249) suggests that 'Quartzite appears to take and retain these marks(crescentic fractures) better than any other rock'.

Striae have similarly been recorded on a wide range of rock types, although they are often most numerous and deeper on pelitic and semi-pelitic rocks when compared with the fine striae usually observed on psammitic-type rocks(Peacock,1970a; Thorp,1981a).

Chi-square tests of the data in Table 8.2, in which frequency of marking clusters are related to the main rock types, indicate highly significant results at the 0.001 level, both inside and outside the limits of the Loch Lomond Advance. This confirms field observations that the high, spatial variability of glacial markings across the study area is strongly related to rock type variability. One causal factor related to this variability is the differential erasure of striae and friction cracks by weathering processes since deglaciation. For example, mineralogical and textural differences within and between the Cruachan,Meall Odhar and Starav granites, that comprise the Etive granite complex(Anderson,1937), appear to provide useful pointers to explain some of the spatial variability of glacial markings observed in part of the western Grampians.

Rock type	Inside ice-limits			Outside ice-limits			Totals		
	a	n	d	a	n	d	a	n	d
Volcanic	37.9	18	4.75	16.35	4	2.45	54.25	22	4.06
Granite	102.1	22	2.15	35.65	2	0.56	137.75	24	1.74
Quartzite	73.55	201	27.3	31.2	21	6.73	104.75	222	21.19
Schist and slate	123.45	91	7.37	42.3	10	2.36	165.75	101	6.09
Limestone	20.25	4	1.98	16.75	0	0	37.0	4	1.08
Flags and psammites	77.7	69	8.88	1.3	1	7.7	79.0	70	8.86
Totals	434.95	405	9.31	143.55	38	2.65	578.5	444	7.68
df			4			2			4
Significance (Chi-squared test)			0.001			0.001			0.001

a = outcrop area in km^2

n = number of observed clusters of glacial markings

d = density of markings per 10 km^2

TABLE 8.2.a. Glacial marking cluster data related to rock type for the NW quadrant

Rock type	Inside ice-limits			Outside ice-limits			Totals		
	a	n	d	a	n	d	a	n	d
Volcanic	69.5	19	2.73	117.75	25	2.12	187.25	44	2.35
Granite	532.35	158	2.97	103.75	21	2.02	636.1	179	2.81
Quartzite	50.0	25	5.0	24.75	28	11.31	74.75	53	7.09
Schist and slate	138.25	14	1.01	130.5	22	1.69	268.75	36	1.34
Limestone	4.0	0	0	6.75	7	10.37	10.75	7	6.51
Flags and psammites	454.75	43	0.95	240.5	32	1.33	695.25	75	1.08
Totals	1248.85	259	2.07	624.0	135	2.16	1872.85	394	2.1
df			4			5			5
Significance (Chi-squared test)			0.001			0.001			0.001

a = outcrop area in km^2

n = number of observed clusters of glacial markings

d = density of markings per 10 km^2

TABLE 8.2.b. Glacial marking cluster data related to rock type for the SW, NE and SE quadrants.

Rock type	Inside ice-limits			Outside ice-limits			Totals		
	a	n	d	a	n	d	a	n	d
Volcanic	107.4	37	3.45	134.1	29	2.16	241.5	66	2.73
Granite	634.45	180	2.84	139.4	23	1.65	773.85	203	2.62
Quartzite	123.55	226	18.29	55.95	49	8.76	179.5	275	15.32
Schist and slate	261.7	105	4.01	172.8	32	1.85	434.5	137	3.15
Limestone	24.25	4	1.65	23.5	7	2.98	47.75	11	2.3
Flags and psammites	532.45	112	2.1	241.8	33	1.32	774.25	145	1.87
Totals	1683.8	664	3.94	767.55	173	2.24	2451.35	837	3.44
df			4			5			5
Significance (Chi-squared test)			0.001			0.001			0.001

a = outcrop area in km^2

n = number of observed clusters of glacial markings

d = density of markings per 10 km^2

TABLE 8.2.c. Glacial marking cluster data related to rock type for all quadrants

The Starav granite is predominantly a coarse-grained, porphyritic type that crops out in massive, curving sheets on valley and mountain sides, largely as a result of pressure-release jointing. Very few glacial markings were observed on this type of granite, but within the granite occur fine-grained,siliceous veins (aplites). These frequently project 1 - 4 cm above the host rock, are more resistent to chemical weathering, and where of sufficient width(5 - 10cm) may display abundant numbers of excellent crescentic gouges and crescentic fractures. Fairly abundant numbers of friction cracks were observed on the main body of the Meall Odhar granite which is a relatively fine-grained, acid granite(Bailey,1960). The Cruachan granite is medium-grained and varies from a biotite granite to a quartz-diorite. Crescentic fractures were observed at only three locations and all occurred on glacially-polished quartz veins. Crescentic gouges and reversed crescentic gouges were observed at five other localities on the main body of the granite.

Yet very few unequivocal glacial markings were found on Rannoch Moor granite, Ballachulish granite, Mullach nan Coirean granite and fault-intrusian granite, even though extensive areas of exposed Rannoch Moor granite bedrock especially were traversed in search of markings. In contrast great numbers of friction cracks were observed on quartzite rock exposures. For example, 33% of all recorded marking clusters were observed on quartzite(this figure of course includes striae, but these form a relatively small proportion of the total number) even though quartzite bedrock only covers 7% of the total area(Table 8.2,c.). Even this is a gross under-representation of the true number of markings because of the difficulty of recording actual numbers on a base map(p.65). Clearly rock composition related to weathering rates is a significant factor.

The effects of chemical weathering processes on rocks have long been observed and quantitative studies undertaken to assess their role in the disintegration of minerals and in the formation of residual products(Goldich,1938; Butler,1957; Hill and Tedrow, 1961; Smith,1962; Harriss and Adams,1966). Most studies concur with Goldich's(1938) findings that mineral stability is an inverse function of the normal sequence of crystallisation of minerals from an igneous melt (viz Bowen's Reaction Series), namely olivine,calcic plagioclase,augite,hornblende,alkalic plagioclase, biotite,potash felspar,muscovite and quartz, from least to most stable respectively. Thus, the observations relating to the distribution of friction cracks are clearly explicable; their distribution is strongly controlled by the presence of quartz-rich rocks and veins. The stability characteristics of quartz, in resisting chemical weathering processes under cool,humid temperate conditions provides an acceptable explanation for much of the spatial variability of friction cracks(although the same may not be true for striae).

Extensive areas of granite lack such markings because the granite contains fair proportions of minerals such as plagioclase felspar and ferromagnesian minerals that are least stable under present day environmental conditions. The breakdown of these minerals into residual products loosens the more stable minerals (i.e. quartz and potash felspars) that are then removed by surface wash. Micro-spalling, the etching out of micro-relief features such as weathering pits and grooves, and the loose products of granular disintegration on granite surfaces testify to the lowering of granite surfaces and the removal of friction cracks since deglaciation. Their complete absence on weathered surfaces adjacent to aplite veins,that are scored by numerous glacial markings,is not because markings never formed,because in areas where the granite surface has escaped weathering glacial markings occur in abundance on both the aplite veins and on the main body of the granite.

However, the above statements have to be qualified as not all quartzites are characterised by abundant friction cracks. Thin-bedded, impure quartzites generally carry relatively few markings whereas massive, nearly pure quartzites may be scored by so many friction cracks,chattermarks and fine striae that superimposition of one type of marking on another is quite frequent. In addition, surprisingly few markings were observed on the highly-siliceous volcanic rocks, as for example, the strongly ice-moulded rhyolites of the Glen Coe area. The very irregular,knobbly surfaces of ice-moulded bedrock suggest that the tough,brittle volcanics mainly fracture along closely-spaced weaknesses under glacial erosion processes; angular fractures form and angular chips of rock are removed in contrast to the curved fractures and crescent-shaped rock fragments so typical of the smooth joint-blocks of the quartzite.This may provide a possible explanation for the paucity of markings on the volcanic bedrock,relative to the numbers observed on quartzite, although petrographical work and laboratory tests on variables such as the tensile strength of the rock and the presence of in situ stresses might resolve this problem.

Although rock type is of paramount importance, in explaining much of the spatial variability of glacial markings, other factors may be responsible for a scarcity of markings in areas where bedrock of a suitable composition is present(i.e. quartzite) and some of these will be discussed in the following section.

8.5 Glacial markings related to ice-limits

The mapping of Loch Lomond Advance limits within the study area enables the glacial marking clusters to be differentiated into two separate populations that are assumed to relate to separate periods of glaciation. This is based on the reasonable assumption that the Loch Lomond Advance glaciers are likely to have erased the great majority of the markings relating to earlier glacial activity(Flint,1971, p.93).

Field mapping by the writer suggests that glacial markings outside the limits of the Loch Lomond Advance are often poorly preserved and few in number, especially on the former nunataks at high altitudes. This generalisation does not hold for all areas outside the ice-limits, especially if the bedrock is rich in quartz and is at a low altitude. For example, abundant friction cracks occur on quartzite bedrock at three locations W and N of Glen Creran(NN 08 0549, NN057522 and NN 039511) and at several places on Meall Odhar granite surfaces W of Glen Etive(Figure 8.1).

To test that a statistically significant difference exists between the marking clusters on either side of the Loch Lomond Advance limit, densities of marking clusters for each rock type,inside

Rock type	NW quadrant		SW, NE and SE quadrants		All quadrants	
	Inside ice limits	Outside ice limits	Inside ice limits	Outside ice limits	Inside ice limits	Outside ice limits
Volcanic	4.75	2.45	2.73	2.12	3.45	2.16
Granite	2.15	0.56	2.97	2.02	2.84	1.65
Quartzite	27.3	6.73	5.0	11.31	18.29	8.76
Schist and slate	7.37	2.36	1.01	1.69	4.01	1.85
Limestone	1.98	0	0	10.37	1.65	2.98
Flags and psammites	8.88	7.7	0.95	1.33	2.1	1.32
Significance	0.001		ns		0.001	

TABLE 8.3 Mann-Whitney U test using glacial marking cluster densities (per 10 km^2).

and outside the limit, were compared using the Mann-Whitney U test. This was done initially for the NW quadrant(Thorp,1981a) and later for the remainder of the study area and for the combined data(Table 8.3).

A one-tailed test significance level of 0.001 for the NW quadrant and the combined data indicates that numbers of glacial markings are significantly greater inside the limit than those outside the limit. However, differences over a large part of the study area were not significant statistically and this may reflect insufficient data from a number of areas(p.194). Nevertheless, it could be argued that the combined data do show a significant difference between numbers of markings inside and outside the ice-limits. If correct one possible explanatory factor could be the difference in in the length of time during which the bedrock, inside and outside the ice-limits, has been exposed to weathering processes.

If it is assumed that the last ice-sheet completely buried the mountains at its maximum extent(see chapter 13), then weathering of inferred ice-sheet markings is presumed to date from deglaciation. Sissons(1976) used radiocarbon dates, estimates of summer temperatures derived from beetle evidence(Coope <u>et al</u>, 1971), and the presumed age of retreat of Atlantic polar water to the W (Ruddiman and McIntyre,1973) to infer that total deglaciation had occurred in Scotland by 12,500 - 13,000 radiocarbon years ago. There is generally common agreement that the termination of the Loch Lomond Stadial occurred some time between 10,600 and 10,000 years B.P. Consequently a difference of about 2,000 - 3,000 years is likely between the time of exposure to weathering of ice-sheet markings compared with those formed during the Loch Lomond Stadial. Since marking densities within the ice-limits are greater than those outside the limit by a factor of 2(Table 8.3), time alone is therefore rejected as the major factor creating the observed numerical differences inside and outside the ice-limits. A more realistic explanation is that rates of chemical or mechanical weathering, or both, increased during the Lateglacial.

Coope et al.(1971) have suggested from coleopteran evidence that average July temperatures for central lowland Britain fluctuated from 8°C at Late Devensian maximum, to 16°C during the Lateglacial Interstadial, and then fell again to 10°C during the Loch Lomond Stadial. Using firn-line calculations and the presence of fossil frost wedges, Sissons(1980, p.40) has suggested that mean January temperatures were no higher than -9°C in part of the western Grampians during the Stadial.

Thus there are two possible times when severe frost-riving could have affected the bedrock and hence the degree of preservation of the glacial markings outside the Loch Lomond Advance limit. Severe frost-riving could have occurred during Late Devensian ice-sheet deglaciation, but direct evidence for this is lacking. There is more positive evidence for much periglacial activity during the Loch Lomond Stadial(Rose,1975; Sissons,1979d; Ballantyne,1981). Severely frost-riven bedrock at many localities outside the Loch Lomond Advance limit, and not within the limit, testify to enhanced rates of mechanical weathering during the Lateglacial. Whether rates of chemical weathering changed markedly during the Lateglacial is much more difficult to evaluate because of a lack of data at the present time.

8.6 Analysis of directions of glacial markings

The data on the directions of the glacial markings, inside and outside the limit of the Loch Lomond Advance, are presented in the form of rose diagrams with classes at 10^{o} intervals(Figures 8.2 and 8.3). The summary statistics for all marking clusters are shown in Table 8.4.

	n	Resultant vector (Degrees from N)	Vector magnitude	Significance
Ice-sheet glacial markings E of the main ice-shed	36	82.3	92.0	0.001
Ice-sheet glacial markings W of the main ice-shed	137	265.8	83.8	0.001
L.L.A. glacial markings E of the main ice-shed	153	66.5	85.0	0.001
L.L.A. glacial markings W of the main ice-shed	511	259.6	52.3	0.001

L.L.A. : Loch Lomond Advance

TABLE 8.4 Analysis of orientation data(directions of glacial markings) calculated by radius vector methods.

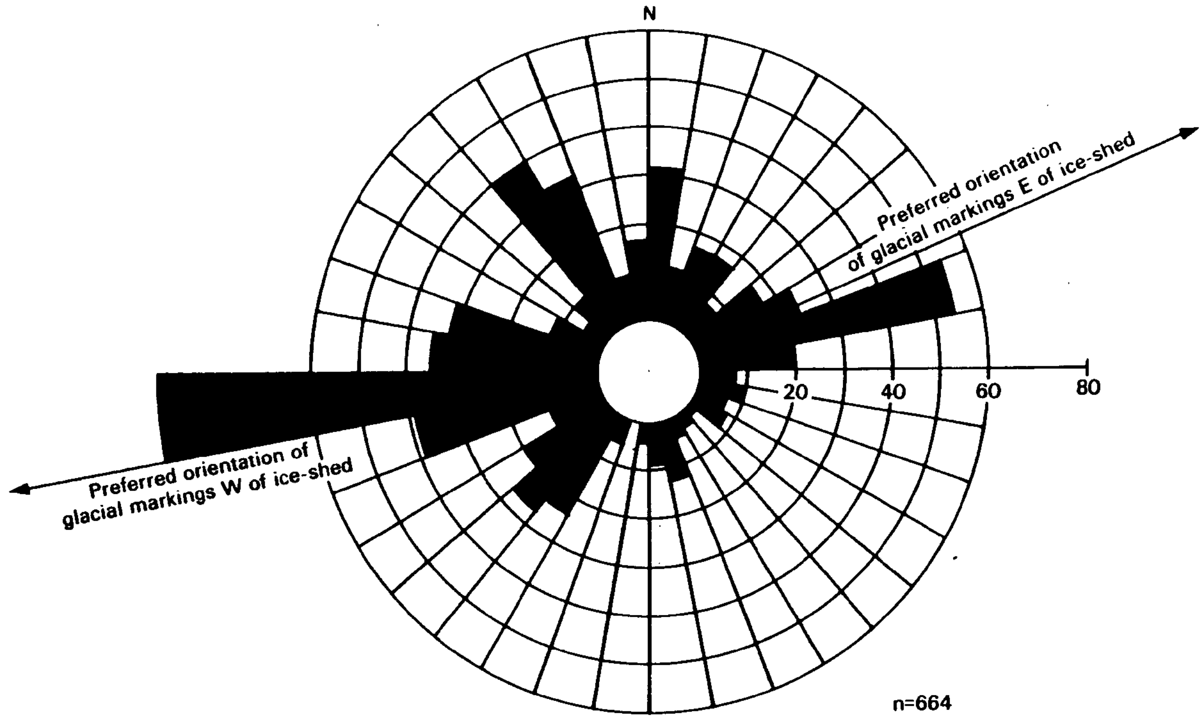


Figure 8.2 Orientation of inferred Loch Lomomd Advance glacial markings

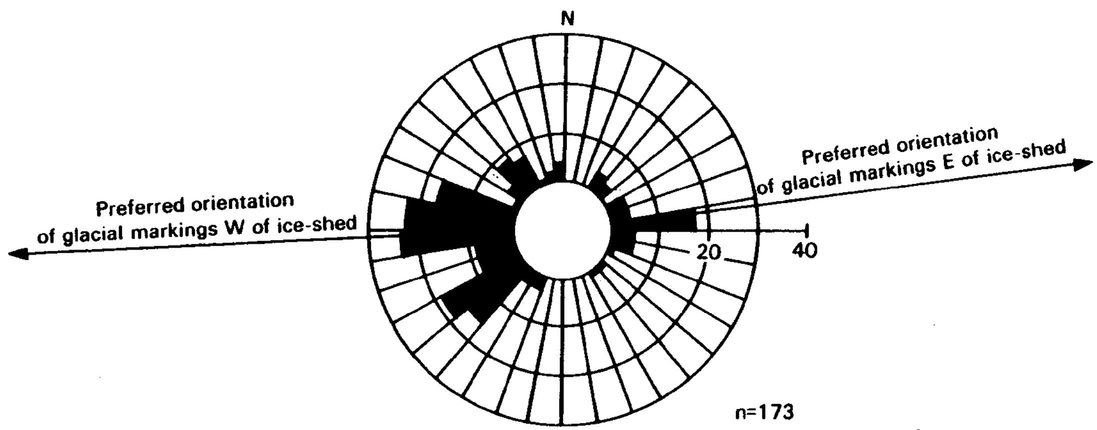


Figure 8.3 Orientation of inferred ice-sheet glacial markings

Although markings are recorded for virtually all classes there is a very strong component of westward-directed markings both for ice-sheet and Loch Lomond Advance markings. This relates to the greater abundance of markings W of the main ice-shed. The westward flow of ice is reflected by the resultant vectors of 259.6° and 265.8° for the Loch Lomond Advance and ice-sheet markings respectively.

The lowest vector magnitude value of 52.2%, for the glacial markings within the Loch Lomond Advance limit and W of the main ice-shed, reflects the large number of corries and tributary valleys in the deeply-dissected Western Mountain zone, where ice flowed in directions at variance with the westward flow of ice down the main outlet valleys. This demonstrates the strong controls exerted by topography in the W on many of the Loch Lomond Advance glaciers. A high vector magnitude of 85% for Loch Lomond Advance markings in the E reflects the far less complex topography of the eastern half of the study area.

In ice-sheet times the topography in the W exerted far less control over ice-flow movements(_cf_ a vector magnitude of 83.8% with a vector magnitude of 52.3% for the Loch Lomond Advance markings). For example, glacial markings on mountain ridges, cols and summits on either side of Glen Coe indicate that ice-sheet flow occurred transverse to the axes of the corries and tributary troughs, although it is likely that more powerful ice-streaming took place along the main westward-orientated valleys.

However, if the markings in the corries and tributary troughs are excluded from the calculations, all the resultant vectors and magnitudes for the Loch Lomond Advance and ice-sheet markings show a close similarity. Thus it would appear that the deep glacial troughs of Leven,Coe,Creran,Etive,Orchy,Rannoch,Ericht, Ossian and Treig performed basically similar functions as outlets draining ice from the Rannoch Moor ice-cap and from the Western Mountain zone both during the Loch Lomond Stadial and the Late Devensian Glacial.

CHAPTER 9

DEGLACIATION OF THE MAIN LOCH LOMOND STADIAL ICE MASS

9.1 Introduction

The way in which the Loch Lomond Advance glaciers retreated from their maximal position has been the subject of some discussion(Peacock,1971a; Sissons,1976,1979c; Gray,1982). Sissons (1976) has noted that in the Scottish Highlands virtually all the clear end moraines are restricted to the terminal zones, that are generally less than several hundred metres in width, of the former glaciers. Exceptions include moraines formed at ice-dammed lake sites such as in Glen Spean(Sissons,1979b,1979c,1981), as shown on Figures 5.2 and 5.4. Such evidence implies that the multiple end-moraines may be partly the result of climatic fluctuations during the intial stages of deglaciation. The absence of end moraines farther up-valley from the maximal limits suggests that following the initial fluctuations of the snout stagnation was rapid. Elsewhere, as for example in Snowdonia(Gray,1982), in the Southern Uplands and the Lake District (Sissons,1979d) some of the glaciers may have retreated actively as 'retreat' moraines occur well within the maximum limits.

The widespread occurrence of hummocky moraine within the limits of the Loch Lomond Advance glaciers could be interpreted as representing widespread stagnation of the ice following initial active retreat. Eyles(1979) and Hodgson(1982) have cast doubts on the use of hummocky moraine as an indicator of downwasting stagnant ice and have suggested that such moraine can form beneath actively retreating ice. Hence the use of this type of morphological evidence to postulate the glaciological conditions operating at the end of the

Loch Lomond Stadial remains questionable.

The timing and mode of deglaciation of the Loch Lomond Advance glaciers has been investigated by Lowe and Walker (1976,1980,1981) using radiocarbon dating and pollen stratigraphy.. Basal organic sediments from sites on Rannoch Moor, the Tyndrum area and from a site near an end moraine at Callander, however, provide radiocarbon dates that range from 11,350± 285 to 8,130± 40 yrs B.P. These clearly fail to provide the degree of resolution needed to determine the pattern of deglaciation of the main Loch Lomond Stadial ice mass. Some of the errors likely to have influenced the accuracy of the radiocarbon dates have been discussed by Lowe and Walker(1980) and Sutherland(1980).

In a recent paper Lowe and Walker(1981) have used biostratigraphic evidence from three sites(Mollands near Callander, Tyndrum at the head of Glen Lochy and Kingshouse on Rannoch Moor) to argue that deglaciation of the main Loch Lomond Stadial ice mass occurred earliest at the periphery of the ice mass(e.g. Mollands) and latest at the centre of the ice mass(e.g. Kingshouse). However, many more biostratigraphic investigations at different sites will be required to verify this hypothesis.

9.2 Morphological evidence along the western sea lochs

On the western side of the main ice mass the rate of retreat of the outlet glaciers terminating in tidewater is likely to have differed from that of the outlet glaciers terminating on land on the eastern side of the ice mass. The retreat of glacier terminii that end in water is generally much more rapid that those that end on land(Flint,1971; Sugden and John,1976; Denton and Hughes, 1981; Paterson,1981).

The rapid retreat of glaciers that calve into tidewater can be related to the combined effects of the relatively high temperature of the water and the effects of the tides. Since a glacier terminus in water cannot stagnate *in situ* as it has a calving front the terminus must respond at once to any decrease in the supply of ice; hence the rapid response in comparison with land-based glaciers. Paterson(1981, p.183) for example, has noted "that large tidewater glaciers seem to behave like ice shelves in that they rely on shoals, headlands or constrictions in the channel to anchor the terminus. A slight initial retreat from such a position results in catastrophic break-up and the terminus retreats rapidly to the next anchor point".

Others(*e.g.* Mercer,1961; Peacock,1971a; Funder, 1972) have also stressed the influence of topography on the precise location of fiord glacier snouts either when advancing or retreating. For example, Mercer(1961) has convincingly shown that at the end of an advance phase the snouts of present-day Alaskan glaciers are almost invariably located at points of local widening, at tributary junctions or at the mouths of fiords, where increased calving will cause the glacier snout to stabilise. He further notes that a land-based glacier will respond to a fall in the firn line by advancing and spreading out to increase the ablation area. A glacier confined to a fiord of constant width, however, will respond to a falling firn limit by advancing, sometimes for a considerable distance, until it reaches a wider part of the fiord and wastage by calving can increase substantially.

As Paterson(1981) has noted(see above) topography

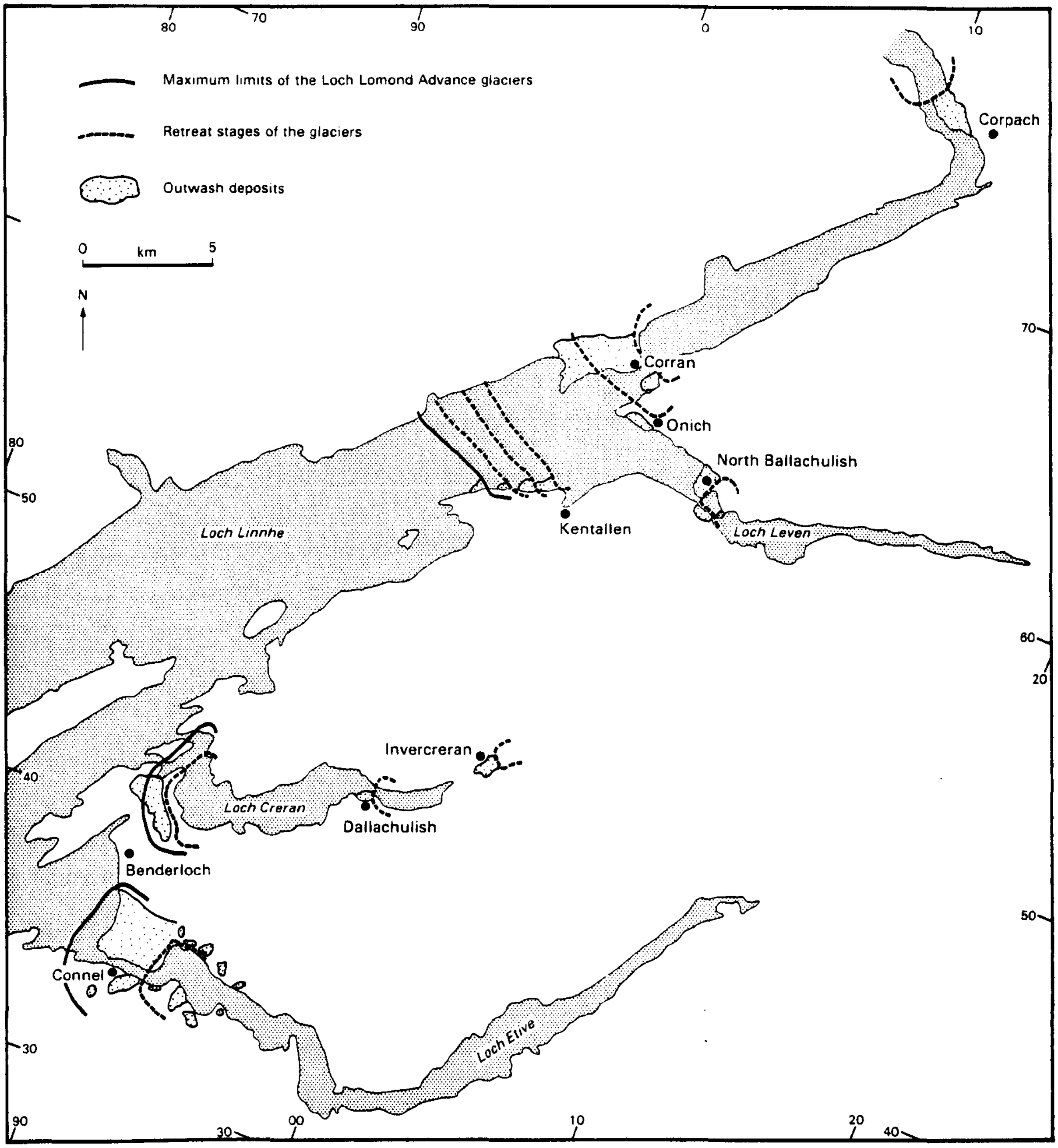


Figure 9.1 Deglaciation features of the Linnhe, Leven, Creran and Etive glaciers

can also be of major importance in the stabilising of calving glacier ice-fronts during a phase of retreat. A marked shallowing or a narrow constriction in the fiord can cause a retreating glacier to stabilise at that point. Peacock(1971a) has used this thesis to argue that, during the deglaciation of the Loch Lomond Advance glaciers in western Scotland, glaciers such as the Linnhe, Creran, Etive and Shiel stabilised at narrow rock constrictions or rock thresholds a short distance from their maximal position. These locations are now characterised by large outwash spreads such as those at Connel, Corran, Benderloch(Figure 9.1) and at the western end of Loch Shiel. Observations by the present writer support this view and elaborate on Peacock's conclusions by referring to additional new evidence.

The maximum limit of the Linnhe glacier, as described previously(p105), is believed to be represented by a clearly defined outwash fan at NM979570(Figure 9.1). The fan projects for <u>ca</u> 250 metres into Loch Linnhe and its surface slopes to the W and SW from a maximum height of +10m at its ice-contact northeastern extremity. The feature was clearly supplied with debris from a meltwater stream flowing under or along the lateral margin of the glacier (Figure 9.2). Identical fluvioglacial landforms have been described by Powell(1981) as forming along the margins of tidewater glaciers in Glacier Bay, Alaska at the present time.

The position of the outwash fan in relation to the varying width of Loch Linnhe is of some significance in the light of the previous discussion. The fan is located where the loch narrows to a width of <u>ca</u> 3.5km; 3km to the SW and 3km to the NE the loch widens to over 5km. Yet, as already stated, Mercer(1961) noted that advancing fiord glaciers in Alaska almost always stabilise at places

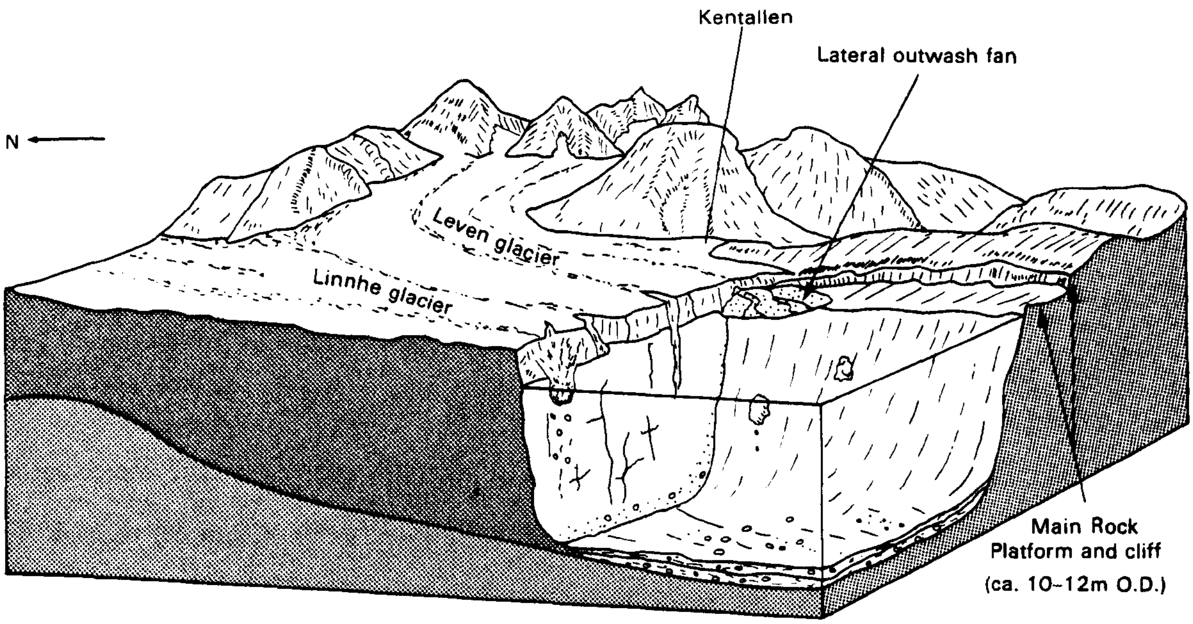


Figure 9.2 Diagrammatic view of the Linnhe glacier at its maximum limit. Not to scale.

where the fiords widen, where losses by calving increase rapidly. The apparently anomalous position of the Linnhe glacier at its maximum position, however, can be explained by changes in the altitude of the firn line. For example, Mercer suggested that the reason that a glacier in Glacier Bay,Alaska stabilised in the eighteenth century in a narrow part of the fiord was because the firn line had begun to rise <u>before</u> the glacier had reached its maximum position. The advance was able to continue because calving losses decreased as the fiord became progressively narrower. A similar situation can be envisaged for the Linnhe glacier with the advancing ice-front reaching Kentallen before the rise in the firn line, but still continuing to advance a further 3km after the rise in the firn line, because of a narrowing of the fiord.

The small size of the outwash fan(<u>ca</u> 0.06km^2) at the maximum limit of the Linnhe glacier may give added support for such a thesis since this suggests that the ice-front only achieved equilibrium for a short time before retreating to the NE.

The ice-front retreated for about 1km before stabilising at NM986574 to form a second outwash fan with an area of <u>ca</u> 0.08km^2, although its morphology is less clearly defined and solid rock crops out in places. Further rapid retreat for about another kilometre before the ice-front stabilised again is indicated by a large outwash fan(<u>ca</u> 0.36km^2) near Ardsheal House(NM995576). Another large outwash fan covering an area of <u>ca</u> 0.25km^2 and signifying further intermittent retreat occurs E of Ardsheal House.

The four outwash fans formed during the retreat of the Linnhe glacier might suggest that climatic oscillations were responsible for the inferred stillstands since the ice-front was

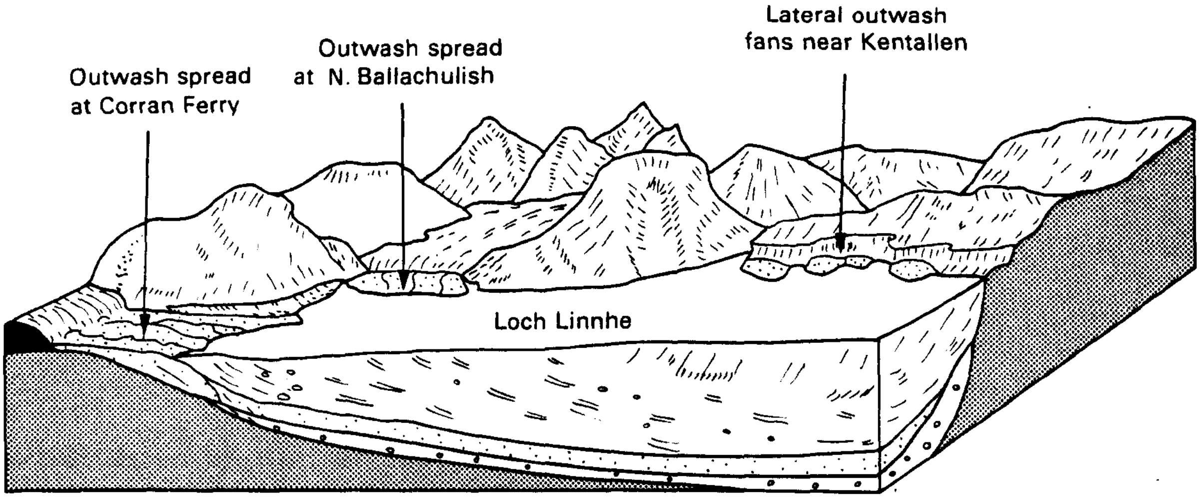


Figure 9.3 Diagrammatic view of the Linnhe glacier during a retreat stage. Not to scale

retreating along a widening fiord and calving would have progressively increased. However, there is no positive evidence in the area so far to suggest that such climatic fluctuations did occur. Alternatively the increasing size of the outwash fans towards the NE might be interpreted as reflecting continuing climatic amelioration, thus accelerating the supply of meltwater delivered to the glacier terminus and by implication the supply of debris. However, this is difficult to verify without corroborative evidence.

Beyond Kentallen where the width of Loch Linnhe increases sharply there is no more evidence of glacier stillstands until the confluence of the Linnhe and Leven lochs is reached(Figure 9.3). This suggests that continuous retreat took place for over 4½km before the Linnhe glacier divided into two and retreated as two independent glaciers along lochs Linnhe and Leven.

A slow retreat or stillstands of the Linnhe glacier are indicated by the outwash deposits at Corran Ferry,and at Onich where the glacier retreated into the narrower northern part of Loch Linnhe and where the calving ice-front was reduced in width from nearly 6km to less than 3.5km. The outwash deposits at Onich formed along the lateral margin of the Linnhe glacier when the outlet via the Abhainn Righ valley(NN024629) was still blocked by ice. Debris was transported along the two meltwater channels at NN025625 and NN 030623 to form the outwash fan at Onich(Figure 9.1). The presence of the fan demonstrates that the Leven glacier must have retreated to at least the eastern edge of the fan at the time of its formation. As the Linnhe glacier retreated to the NE the two meltwater channels ceased to function as outlets for meltwater and the fan at Onich was starved of debris. The ice-front eventually stabilised in the vicinity of Corran Ferry where the loch narrows from 2.6km to 1.8km and calving losses decreased. Equilibrium may have been achieved for a time at this constriction allowing the build-up of the outwash deposits at Corran Ferry (Figure 9.1) before the supply of ice diminished and further retreat continued for 16km along Loch Linnhe (no more substantial deposits of outwash occur until the entrance to Loch Eil is reached(Peacock,1970a)).

The retreat stages of the Leven glacier are recorded by the outwash spreads at North Ballachulish, where the width of the loch is reduced from 1.7km to 0.8km, and possibly near Caolasnacon(NN138611) where a narrow rock constriction occurs. However, the deposits at the latter locality have probably been created by debris brought down by three streams flowing off the mountains to the S.

The maximum position of the Etive glacier is believed to be about 2km W of Connel Bridge(p.117) where significantly the loch widens from 2.6km to 4.1km over the same distance and where increased calving would have caused the advancing ice-front to stabilise at this point.

Retreat of the Etive glacier took place in a narrowing loch enabling the ice-front to achieve equilibrium for a time 2km E of Connel Bridge where the width of the loch at the present time is only just over 2km and where it shallows at the Falls of Lora. This enabled pro-glacial drainage to construct the large outwash spread at the mouth of Loch Etive(McCann,1961,1966; Peacock,1971a; Gray,1975a). Thinning and/or intermittent retreat of the Etive glacier eastwards toward the mountains for 6.5km is recorded by a series of kame terraces(Gray,1975a). No additional outwash deposits were located during the present survey along Glen Etive so it is not known whether there were any further stillstands of the ice-front as the glacier thinned and/or retreated north-eastwards along Loch Etive.

The maximal limit of the Creran glacier is similar to that of the Etive glacier when related to the local topography. The ice flowed through a narrow part of Loch Creran in the vicinity of Dabrannoch(NM938411) and spread out to form a piedmont lobe; the width of the ice-front increased from 3km to 6km, thus probably doubling the calving rate and causing the ice-front to stabilise at the mouth of the sea loch.

The retreat of the Creran glacier to the inner end moraine position(Peacock,1971a; Gray,1975a) and the formation of an outwash spread may relate to climatic amelioration, but this is uncertain. However, the location of outwash deposits farther E into Glen Creran shows that ice retreat was strongly controlled by topography. For example, an outwash fan occurs at Dallachulish(NM 979440) where a rock constriction reduces the width of the valley from 2km to only 0.9km. Finally another stillstand induced by topography is suggested by the outwash train at Invercreran(NN014470) where Gray(1972) levelled the surface as declining from 19.9m O.D. near Invercreran to 13.7m O.D. at Glasdrum. A spur 170m high reduces the width of the valley near the ice-contact slope to less than 0.8km and this reduced the supply of ice to the glacier terminus.

9.3 Morphological evidence on the eastern side of the main ice mass

The glaciers on the eastern side of the main ice mass all terminated on land except for the Ericht glacier. Therefore the pattern of deglaciation of the eastward-flowing outlet glaciers will have differed from those glaciers flowing westward into tidewater with topography having relatively little influence on the precise location of the glacier snout during retreat.

The triple end moraine delimiting part of the Treig glacier and the double end moraine at the maximum limit of the Ossian glacier suggest that climatic fluctuations occurred at the end of the Loch Lomond Stadial(Sissons,1976), but how the glaciers retreated from their maximal position is not known since no clear morphological evidence of additional stillstands was located during the present study. In contrast the retreat stages of the Spean glacier have been fully documented(Sissons,1979b,1979c,1979e,1981) and need not be discussed here.

Only in one area was evidence found from which a sequence of events could be deduced during ice decay. An excellent suite of fluvioglacial landforms occurs to the SW of Loch Ericht (Figure 9.4) where especially favourable topographical conditions existed during the decay of the Ericht glacier for the formation of such features. The area is located where ice from the Rannoch Moor ice-cap flowed to the NE around the southern flank of the Rannoch Forest Range toward the glacial breach of Loch Ericht. This ice stream crossed a shallow valley that at the present time is drained by a small stream flowing to the S(Figure 9.4). During deglaciation the thinning of the ice gave rise to a downwasting lobe of ice within the valley that produced a series of inter-related landforms.

In the initial stage the flow of meltwater toward Loch Ericht is indicated by an ice-directed esker at 1 that extends for ca 1km to a col meltwater channel at an altitude of ca 415m O.D. As the ice thinned a new escape route for the meltwater at a lower altitude of ca 370m O.D. was opened up across the col at

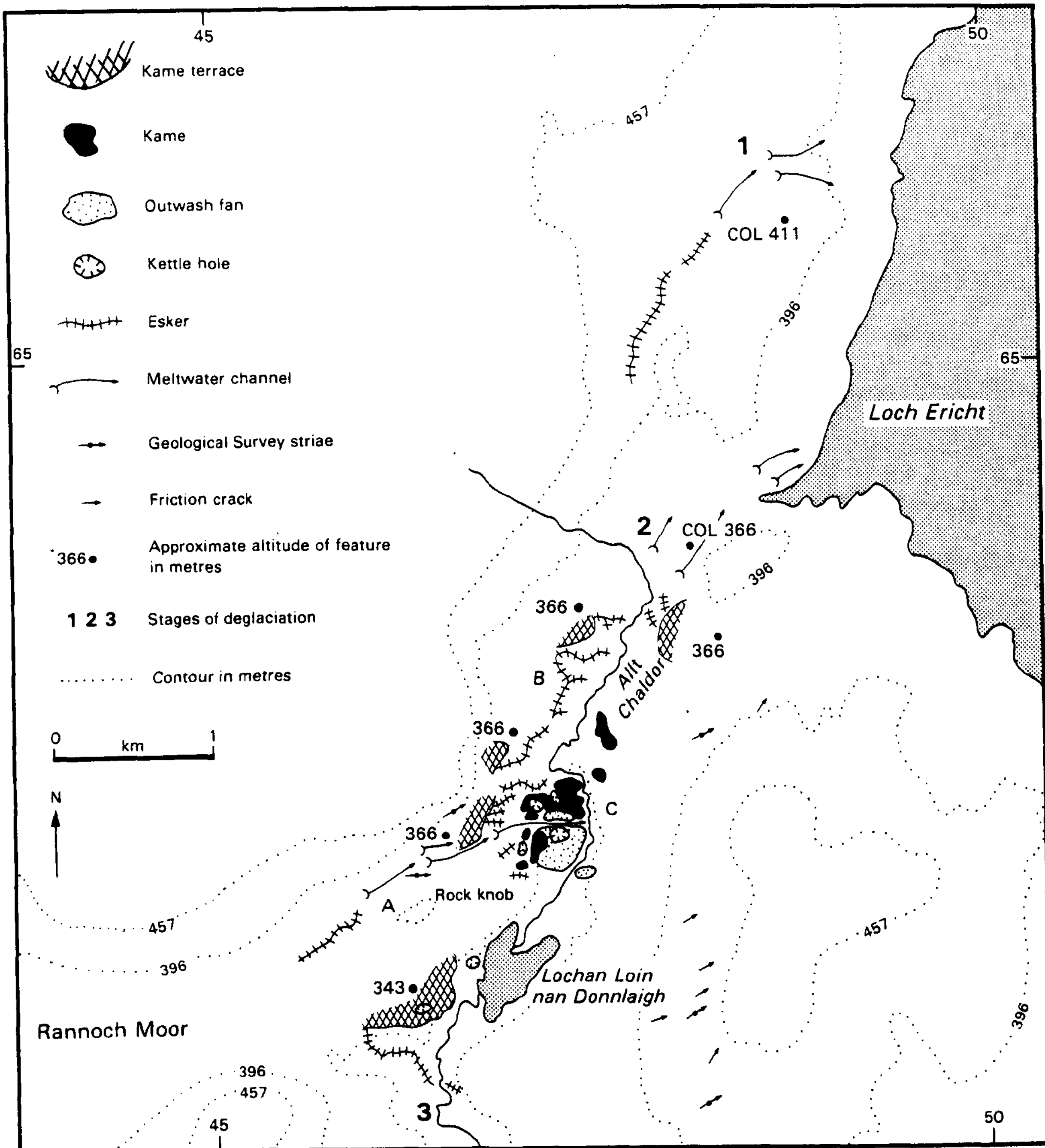


Figure 9.4 Landforms of deglaciation in the area SW of Loch Ericht. For explanation of the reference letters A, B and C see text

2. During the time that this col functioned as the main outlet for the meltwater a series of inter-related eskers,kames,kame terraces, kettles and an outwash fan formed.

Several kame terraces at altitudes of ca 366m O.D. flank the western side of the shallow valley of the Allt Chalder (Figure 9.4) and these were supplied with debris brought by meltwater flowing along a channel between the rock knob at A and the mountain side. The kame terrace at NN473633(just SW of 2) has a surface that slopes away from the ice-contact edge showing that some of the debris was derived directly from the ice. All the kame terraces are associated with subglacially engorged eskers that run oblique to the valley side from the upper edge of the terrace down toward the valley floor. The majority of the eskers are less than 400m in length except for the esker at B. Large flat-topped kames and steep-sided kettles occur in the vicinity of C. Sections in the kames show stratified sandy material, to thicknesses of ca 10m, with very few pebbles and cobbles.

The relationships between the kame terraces,subglacially engorged eskers and kames demonstrate that fluvioglacial deposition took place within and along the margins of a stagnating mass of ice occupying the valley floor. The accordant altitude of the upper surfaces of the kame terraces indicates that deposition was related to an englacial water table(Sissons,1958), the level of which was controlled by the altitude of the col at 2. Deposition in standing water is also suggested by horizontally-bedded fine and coarse sands exposed in a river section at NN474620 on the E side of a low outwash fan that formed between the kames during a late stage of ice decay. Kettle holes in the vicinity of C(Figure 9.4) and within the outwash fan demonstrate the final wasting away of the ice lobe as the ice disintegrated into separate blocks.

The last stage in the deglaciation of the area is indicated by the low kame terrace and esker at 3. The esker is orientated to the SE, parallel to the valley axis, and clearly indicates an advanced stage of ice thinning and disintegration with the orientation of the esker controlled by the local topography and not by the former ice surface sloping down to the NE. This evidence shows that meltwater no longer flowed toward Loch Ericht across the col at 2, but instead escaped through the ice toward Loch Rannoch.

9.4 Conclusions

The details and timing of the deglaciation of the main ice mass in the western Grampians cannot yet be described adequately because of insufficient data at the present time. Nevertheless, it is possible to briefly outline the major problems involved and to draw some broad conclusions from the scanty evidence, although these will be largely speculative.

The precise dating of the end of the Loch Lomond Stadial, that is generally ascribed to a marked climatic amelioration, is still under dispute(see recent summary by Sissons, 1979d, p.202). For example, recent radiocarbon dating evidence(Lowe and Walker,1980,1981; Sissons,1979d) suggests that deglaciation may have begun before ca 10,600 yr B.P. rather than after the generally accepted date of ca 10,300 yr B.P. Thus it is not yet possible to be precise about the time when the glaciers in the study area began a major retreat. Nor is it possible to correlate the retreat stages of the main tidewater glaciers on the western side of the main ice mass since these appear to have been almost entirely controlled by topographic factors and not by climatic factors.

The advance and retreat of glaciers, even over a relatively small area of less than 500km^2, may not necessarily be synchronous. For example, if the interpretation given in section 9.2 of the fluvioglacial evidence along lochs Linnhe,Leven,Creran and Etive is correct this could imply that the Linnhe glacier was still advancing while the other tidewater glaciers were beginning to retreat(cf modern day tidewater glaciers in Alaska(Mercer,1961)), although there are no radiocarbon dates to support this contention.

The response times of different-sized glaciers to climatic fluctuations is known to vary considerably(Sugden and John,1976; Paterson,1981) with the smaller corrie and valley glaciers generally having a much shorter response time than larger glaciers such as those comparable in size to the main outlet glaciers in the study area. Unfortunately, there is no information available yet to show that such different response times occurred between different glaciers at the end of the Loch Lomond Stadial, although it would seem reasonable to suppose that the main ice mass is likely to have responded more slowly to climatic change than the smaller independent corrie glaciers.

The spatial pattern of radiocarbon dates obtained from basal organic matter(Lowe and Walker,1980,1981) within the limits of the main ice mass have so far failed to produce the expected results _if_ it is assumed that the glaciers decayed mainly by backwasting and by retreating into the high ground(_i.e_ the oldest dates near the maximal limits of the glaciers and the youngest dates near the ice-shed, as for example in the Rannoch Moor area). However, this hypothesis may not be correct if the ice mass mainly decayed by downwasting, especially if this occurred in the later stages of deglaciation. The timing of deglaciation in different areas would then largely be a function of the thickness of the ice mass, if all other factors were assumed to be equal, with the youngest dates to be anticipated in areas where the ice was thickest. Figure 12.5 shows that during the Loch Lomond Stadial the thickest ice occurred in the vicinity of Fort William(ca 580m),at the head of Loch Leven (ca 590m), in the Loch Tulla area(ca 500m), in lower Glen Nevis (ca 510m) and in the central Glen Etive area(ca 580m). On this basis Rannoch Moor would not be the last area to be deglaciated (Lowe and Walker,1976,1981) since the maximum thickness of the ice-cap was only just over 400m.

PART 111

PALAEOCLIMATIC INFERENCES

CHAPTER 10

THE PRESENT DAY PATTERN OF PRECIPITATION IN THE WESTERN GRAMPIANS

10.1 Introduction

The effects of mountain barriers on the distribution and amounts of precipitation are imperfectly understood and subject to debate. This is partly because of a lack of weather stations at high altitudes and partly because of the difficulties of accurately measuring amounts of snowfall on windswept mountain summits and ridges(Ballantyne,1981; Barry,1981).

The water vapour content of air decreases quite rapidly with altitude, but in mountain areas this is compensated for by the forced advective ascent of air that intensifies vertical motion in a cyclonic system. Orographic effects tend to increase the frequency,duration and intensity of precipitation particularly by releasing conditional instability and promoting shower activity, especially in Polar Maritime airstreams and in the warm sectors of cyclones.

Many mountain barriers in mid-latitude west coast locations experience 2 to 3 times the amount of precipitation experienced on adjacent lowland areas. In western Britain maximum precipitation is generally recorded to the lee of summits(Barry and Chorley, 1976). This is attributed to the air continuing to rise beyond the crestline so that a time-lag occurs in the precipitation processes after condensation has taken place. The frictional drag of the mountains can create considerable wind turbulence and variability in precipitation amounts over very short distances. For example, the writer has frequently observed pockets of clear descending air over corries in the western Grampians, when the surrounding mountain summits were blanketed by thick rain clouds formed in a moist Atlantic airstream.

Local variations in the intensity and duration of precipitation can also be created by convergences and divergences of airstreams. For example, Meierding(1982) noted that amounts of snowfall are highest in the centre of the Colorado Front Range in the western U.S.A. He attributed the higher snowfall to the funnelling effect on winter storms from the W created by the low-lying Middle Park to the W of the range. Whereas the convergence of air into valley heads can lead to rapid increases in precipitation the divergence of air currents leads to subsidence, a tendency for clearer skies and decreased precipitation.

The various relationships between the factors described above renders any analysis of the spatial pattern of precipitation in a mountain area, such as the western Grampians, a complex task.

In section 10.2 the various methods used in an attempt to disentangle the factors involved will be outlined while the results will be discussed in section 10.3. In order to compare the present day pattern of precipitation across the western Grampians with the distribution of the reconstructed Loch Lomond Advance glaciers(Figure 6.1) it was first necessary to construct isohyetal maps using the different techniques outlined below.

10.2 Methods

The difficulties of constructing reasonably accurate precipitation maps for mountain areas have been well documented by a number of workers(Unwin,1969; Mandeville and Rodda,1970; Pegley,1970; Barry,1981). Many precipitation maps for mountain areas are based on extrapolation from existing weather station data,using empirical altitude-precipitation relationships,with corrections made for local site factors such as windward/leeward locations(Taylor,1980). However, for such a complex mountain area as the western Grampians errors can still occur with such an approach.

Various methods have been applied to the construction of isohyetal maps ranging from simple arithmetical methods such as the arithmetical mean and Thiesson polygons, through various statistical techniques including simple linear and multiple regression analyses, to more sophisticated methods such as trend surface analysis. In this section some of the above-mentioned methods will be applied to the precipitation data obtained for western Scotland and the results compared in section 10.3.

The mean annual precipitation totals for the period 1941 - 1970 were obtained for 124 stations in western Scotland(Appendix D). The mean annual precipitation total for the years 1883 - 1903 for the summit of Ben Nevis was also included since this provided the only record for a high mountain summit in the western Grampians. The areal distribution of the data for the 125 stations is much larger than the study area and extends into the Western Highlands and the SW Grampians. This was done to establish a 'buffer' zone in order to eliminate the problems encountered at the margins of the control area when using trend surface techniques. For this reason only 58 stations are shown on Figure 10.1.

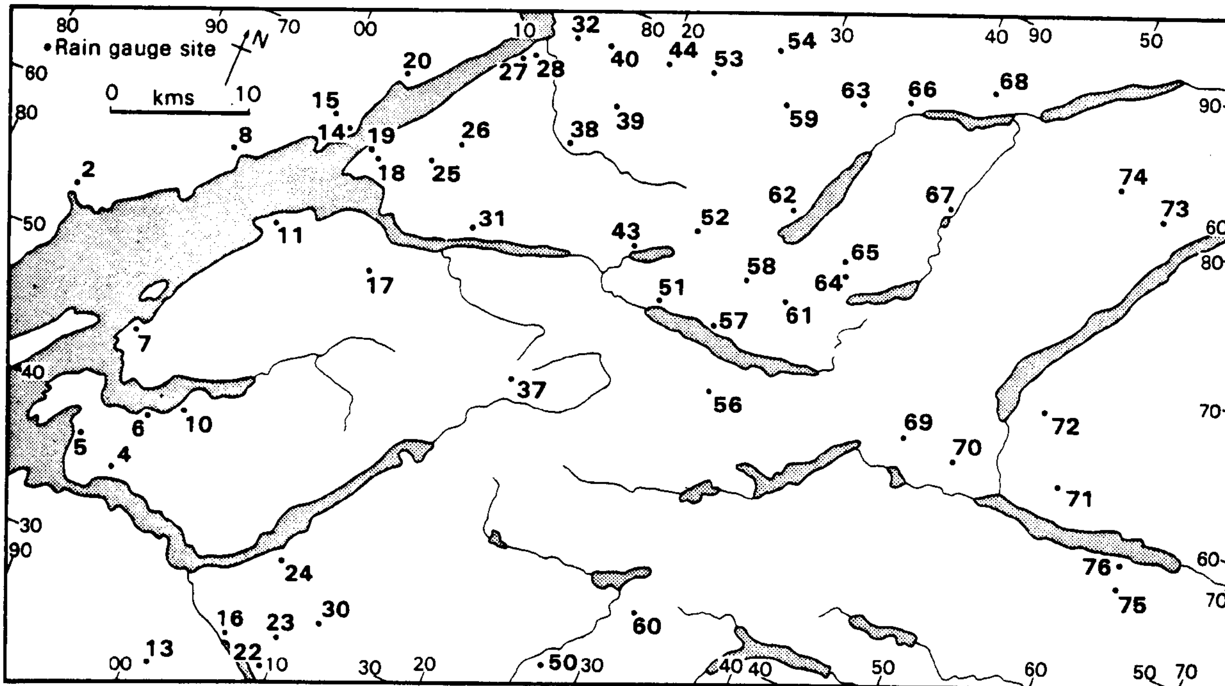


Figure 10.1 Location of rain gauge sites in the study area. These are listed in Appendix D

Area	Methods	Range for subareas (mm yr^{-1} per 100m)	Overall mean (mm yr^{-1} per 100m)	Source
Western Scotland	Pairs of sites	143 - 354	163.0	Author
Western Scotland	Multiple regression	-	191.5	Author
Western Scotland	-	-	253.0	Gloyne(Quoted in Ballantyne,1981)
Eastern Scotland	-	-	83.0	Gloyne(Quoted in Ballantyne,1981)
Southern Uplands	Linear regression	-	240.0	Ferguson(1977)
Southern Uplands	Multiple regression	-	180.0	Ferguson(1977)
Snowdonia,Wales	Linear regression	270 - 622	450.8	Unwin(1969)
Western England	-	120 - 300	-	Salter(Quoted in Barry,1981)
Southern England	-	80 - 150	-	Salter(Quoted in Barry,1981)
U.K.	-	-	119.0	Smith(1976)

TABLE 10.1 Precipitation-altitude gradients in the U.K.

In order to assess approximately the influence of altitude on precipitation totals 7 pairs of station sites, that are less than 5km apart but at different altitudes, were selected from different areas(e.g. 22 & 23, 32 & 40, 38 & 39, 63 & 66, 75 & 76, 114 & 115, 119 & 120 listed in Appendix D) and their precipitation-altitude gradients were calculated. The results indicate that considerable variations in precipitation-altitude gradients exist in western Scotland, ranging from 143 to 354mm yr^{-1} (Table 10.1). However, since these values clearly indicate that other factors are of some considerable importance and that such values can only represent crude approximations of the general precipitation-altitude gradients in western Scotland, simple linear and multiple regression methods were applied to the rain gauge data.

Table 10.2 shows the results of such an analysis for the observed precipitation means against altitude,longitude and latitude. Both longitude and latitude were expressed in arbitrary units N or E of an origin located to the SW of the study area(Appendix D). The matrix of simple linear correlations(Table 10.2.a) shows that the correlations between precipitation and altitude, precipitation and longitude, and precipitation and latitude are significant at the 0.1% confidence level. Table 10.2(b and c) gives the results of sequential multiple regression using the same variables as those used in the linear regressions. The two variable model that provides the greatest improvement over the one variable models is step 4(Table 10.2.b.). As the table shows the addition of latitude as a further variable increases the efficiency of the regression, although the increase in the multiple correlation coefficient from 0.764 to 0.790 is relatively small.

TABLE 10.2 Results of regression analysis

a) Matrix of simple linear correlation coefficients.

	X_1	X_2	X_3	X_4
X_1 Altitude	1.000	<u>0.451</u>	-0.057	<u>0.304</u>
X_2 Longitude		1.000	0.049	<u>-0.489</u>
X_3 Latitude			1.000	<u>-0.274</u>
X_4 Precipitation				1.000

Underlining indicates coefficients that are significant at the 0.1% level

b) Results of multiple regression

Step	Variables entered	R	Rsq	F-Value
1	Altitude X_1	0.304	0.0922	12.6
2	Longitude X_2	0.489	0.2391	38.7
3	Latitude X_3	0.274	0.0753	10.0
4	Altitude, Longitude X_1, X_2	0.764	0.5843	101.3
5	Altitude, Latitude X_1, X_3	0.398	0.1585	9.5
6	Longitude, Latitude X_2, X_3	0.550	0.3021	11.0
7	Altitude, Longitude, Latitude, X_1, X_2, X_3	0.790	0.6241	12.8

F-Value at the 0.1% confidence level, step 7 is 6.85

c) Prediction equations

Step	Equation
1	$X_4 = 1830.67 + 0.911\, X_1$
2	$X_4 = 2632.53 - 12.053\, X_2$
3	$X_4 = 2305.13 - 5.722\, X_3$
4	$X_4 = 2540.44 + 1.975\, X_1 - 19.375\, X_2$
5	$X_4 = 2090.62 + 0.867\, X_1 - 5.378\, X_3$
6	$X_4 = 2861.07 - 11.75\, X_2 - 5.239\, X_3$
7	$X_4 = 2725.769 + 1.915\, X_1 - 18.913\, X_2 - 4.185\, X_3$

In order to objectively assess further the spatial pattern of precipitation across part of western Scotland the data were subjected to trend surface analysis. Figure 10.4 shows the linear,quadratic and cubic trend surfaces based on the data for all the rain gauge stations. However, since the results of the multiple regressions indicate that the data used in the computations incorporate the two main controls of altitude and distance from the Atlantic Ocean, any such trend surface fitted to the data will inevitably include both the regional and orographic components of precipitation. This raises several problems in the interpretation of the results and these will be discussed in section 10.3.

A method adopted to separate the regional precipitation component from the orographic precipitation component was to standardise the rain gauge data to mean sea level by subtracting a correction factor based on the precipitation-altitude gradient (Unwin,1969). The value of 192mm yr^{-1} per 100m obtained from the multiple regression(step 7) in Table 10.2.c. was the value used in the standardisation procedure. This enabled the regional component to be derived from the data. The new set of values was then subjected to trend surface analysis, the results of which are shown in Table 10.5 and 10.7.

10.3 The present day pattern of precipitation in the western Grampians

Inspection of the precipitation-altitude gradients for different areas in the U.K.(Table 10.1) demonstrate a

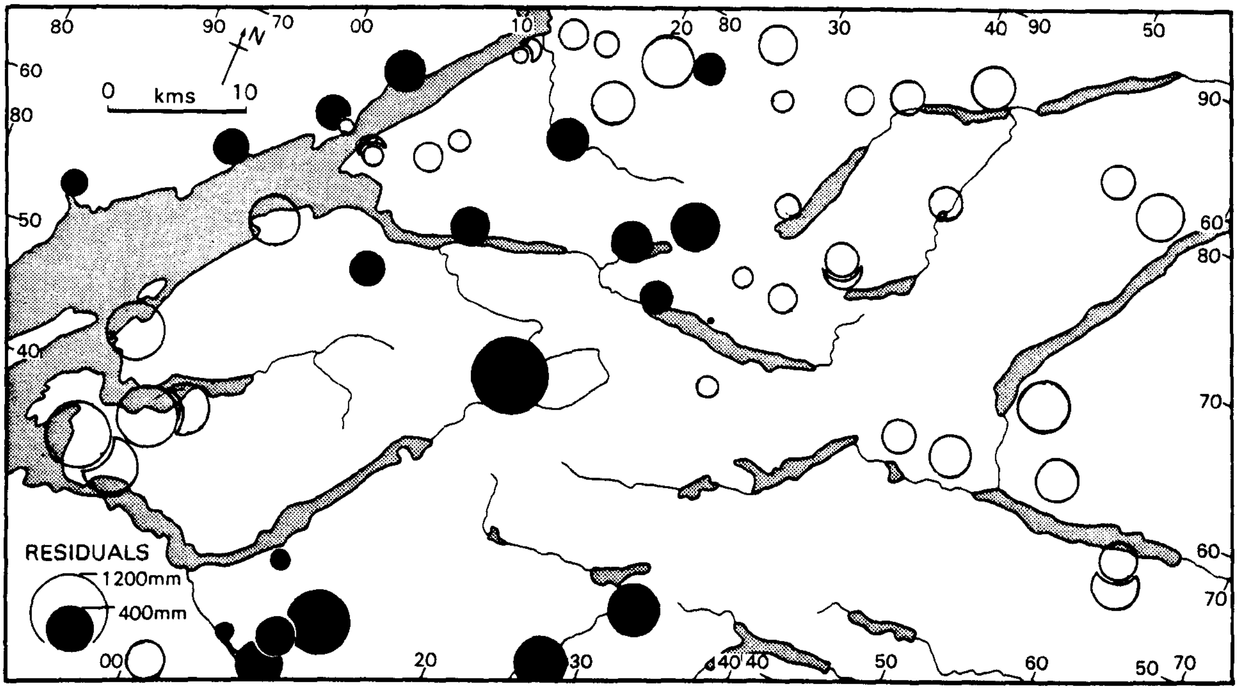


Figure 10.2 Residuals from multiple regression of rainfall on altitude, longitude and latitude.

Positive Negative

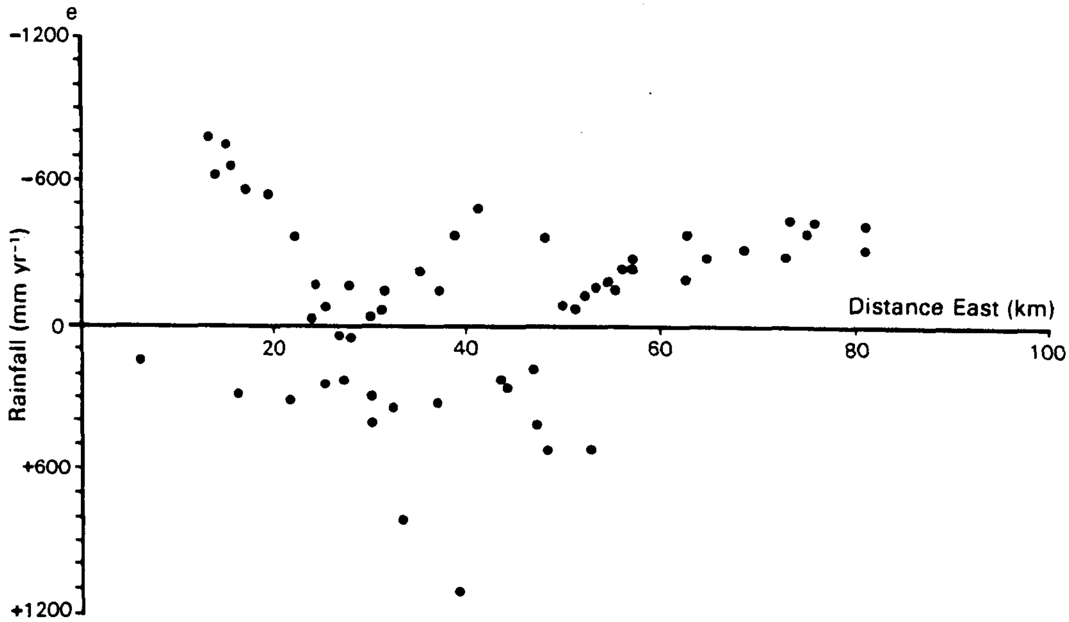


Figure 10.3(a) Scatter diagram of residuals from regression of precipitation on altitude, latitude and longitude

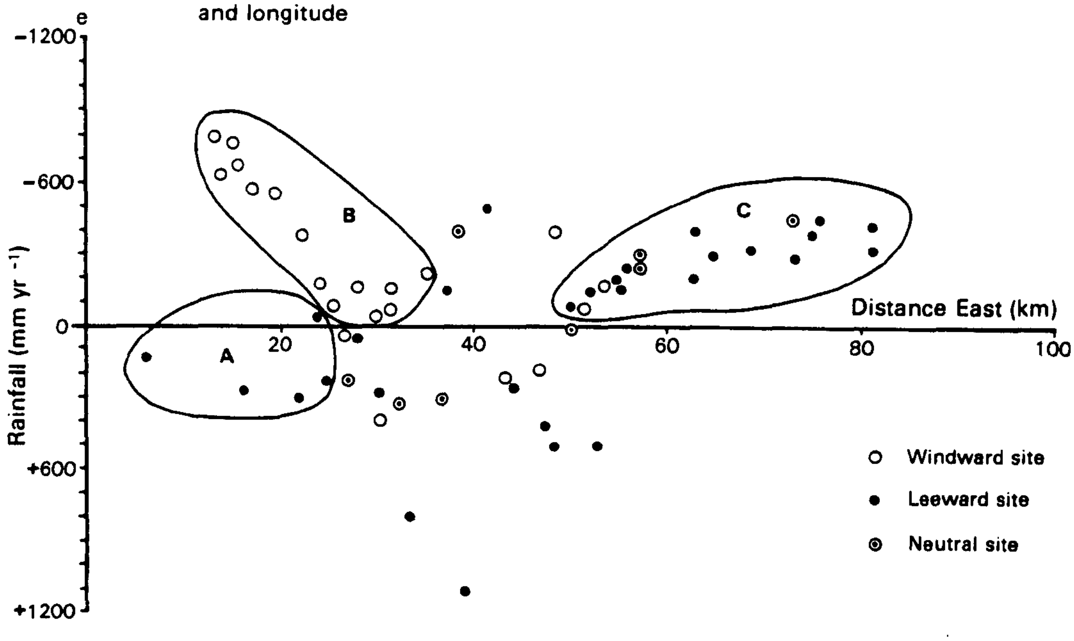


Figure 10.3(b) Scatter diagram of residuals from regression of precipitation on altitude, latitude and longitude

A Low ground on W side of Loch Linnhe
B Low ground on E side of Loch Linnhe
C Area mainly E of Rannoch Moor

wide variability in the results, ranging from a very high value of 451mm yr^{-1} per 100m for Snowdonia(Unwin,1969) to a low value of 83mm yr^{-1} per 100m for eastern Scotland(quoted in Ballantyne, 1981). Nevertheless, it will be noted that there is a fairly good correspondence between the mean values quoted for western Scotland(e.g. ranging from 163mm yr^{-1}per 100m to 253mm yr^{-1} per 100m).

In an attempt to isolate other factors that may have influenced the areal distribution of precipitation in the western Grampians the residuals derived from the multiple linear regression step (7) were plotted on a map of the area (Figure 10.2) and on a scatter diagram(Figure 10.3.a). Inspection of the resultant patterns indicate several distinct clusters of residuals. These comprise a cluster of negative residuals in the area E of Rannoch Moor, a zone of negative residuals along the eastern shore of Loch Linnhe, four positive residuals on the western side of Loch Linnhe and a zone of mainly positive residuals corresponding with the main area of mountains in the W.

In order to assess the influence of topography on precipitation totals each of the 58 sites in the study area was classified according to whether it was to the windward or leeward side of the nearest mountain block. The main moisture-bearing winds were assumed to be blowing from directions between W and SW. Sites that did not clearly belong to either of these two categories (i.e. sites in deep valleys surrounded by mountains) were classified as neutral sites(Figure 10.3.b).

The areal distribution of the residuals shows that precipitation amounts at sites in the lee of the mountains in

Surface	% RSS obtained
Linear	30.5
Quadratic	45.6
Cubic	56.2

Source of variation	Degrees Freedom	% RSS	Mean Square	F	Significance
Total, 125 data points	124				
Due to linear surface with three constants	2	30.5	15.25	26.77	0.1%
Due to residuals over linear surface	122	69.5	0.57		
Due to added quadratic components	3	15.1	5.333	11.666	0.1%
Due to residuals over quadratic surface	119	54.4	0.457		
Due to added cubic components	4	10.6	2.65	6.958	0.1%
Due to residuals over cubic surface	115	43.8	0.381		

TABLE 10.3 Analysis of variance for trend surfaces of precipitation data in the western Grampians..

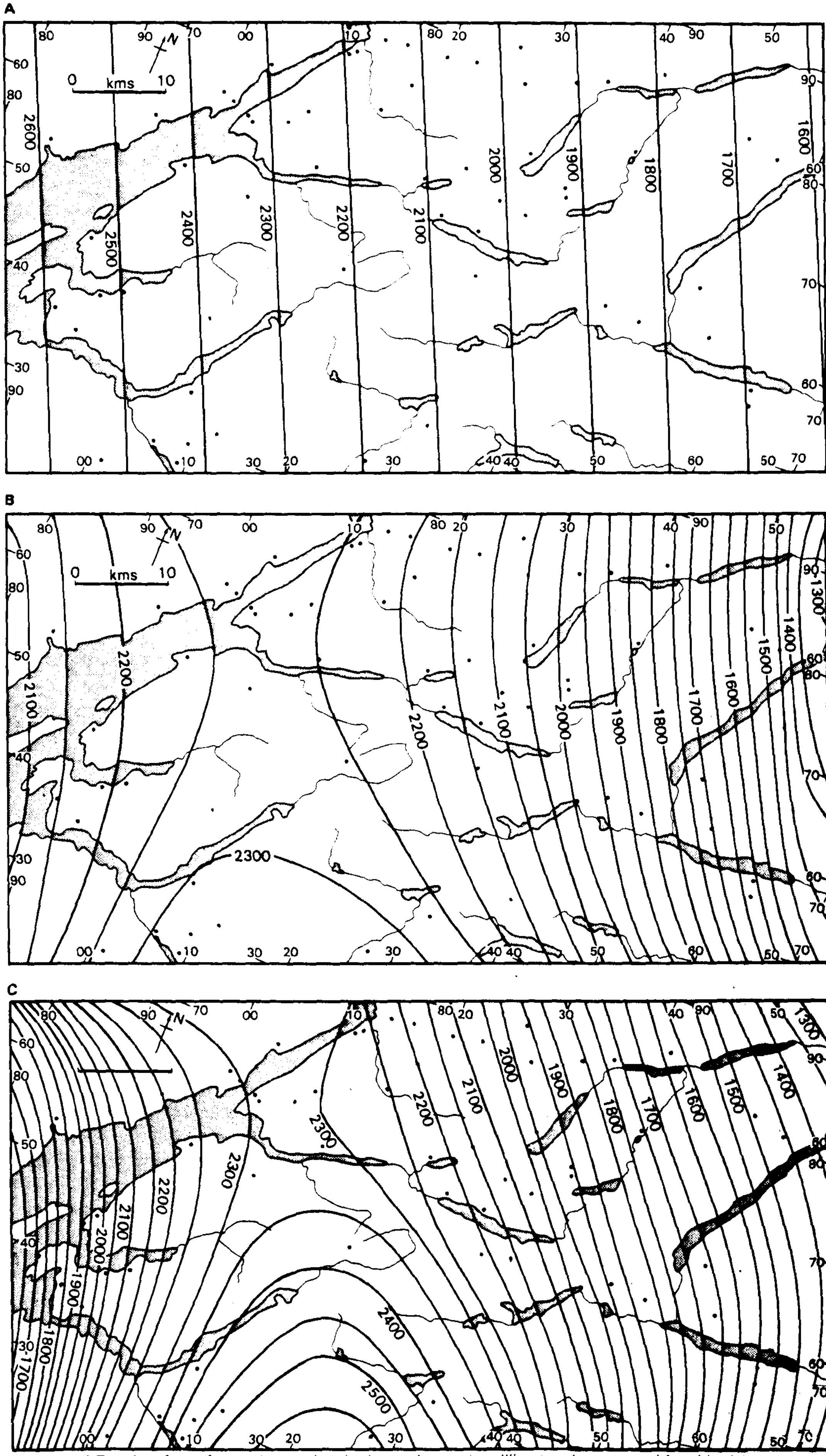


Figure 10.4 Trend surface of precipitation data in the study area in millimetres (uncorrected for altitude)
A Linear trend surface
B Quadratic trend surface
C Cubic trend surface

the W tend to be underestimated by the regression equation while the converse is true for sites E of Rannoch Moor. This suggests that rain shadow effects become more pronounced to the E as the prevailing westerly airstreams become progressively drier. In contrast the cluster of negative residuals along the eastern side of Loch Linnhe reflects the relative lack of orographic influence in amounts of precipitation. However, not all of the large residuals can be explained in terms of a windward/leeward location. For example, station 37 has a very large positive residual of 1135mm because a high precipitation total of 3071mm yr^{-1} occurs at a low altitude of 73m O.D. This anomaly can be explained by the convergence of south-westerly airstreams toward the narrowing valley of upper Glen Etive and bodily uplift of air occurring over the mountains of the Bidean nam Bian and Ben Starav ranges where many peaks exceed 1000m in height.

The linear trend surface shown in Figure 10.4, that has only a moderate fit to the precipitation data, indicates that precipitation declines from a maximum of over 2600mm to less than 1600mm in a WSW to ENE direction across the western Grampians. This simple, highly generalised surface only explains 30.5%(Table 10.3) of the variance in the precipitation data and clearly does not reflect the highly complex pattern of precipitation that exists across the area.

The quadratic and cubic surfaces both show a broad ridge, that in the cubic surface reaches maximum values above 2550mm yr^{-1} in the S and which declines toward the Atlantic Ocean and toward the NE where values are less than 1300mm yr^{-1}. The ridge clearly emphasises the orographic influence and the nearness to the Atlantic Ocean of the Western Mountain zone, extending from Ben Cruachan in the S to the Western Highlands in the NW.

Although the two higher orders provide significant increases in explanation at the 0.1% significance level and the cubic surface explains 56% of the variance in the precipitation data(Table 10.3) there are several major problems related to any conclusions based on these results.

Firstly, there is a notable lack of precipitation data for sites at altitudes greater than 500m O.D. in western Scotland(e.g. there are only 7 in the area covered by the 125 stations and only 4 in the study area). Any trend surface fitted to such data will inevitably bias the results toward the lower precipitation values occurring at altitudes lower than 500m O.D.; thus the maximum value of ca 2550mm yr^{-1} shown in the cubic surface does not accord with the much higher totals of 3000 - 4000mm yr^{-1} known to occur in the Western Mountain zone(e.g. stations 23, 30,37 and 39 in Appendix D). Therefore the surface can only be used in a descriptive sense to indicate where maximum precipitation occurs and it cannot be used in a predictive sense to estimate actual precipitation amounts.

Secondly, the problem described above is compounded by a lack of data from the zone of mountains extending southwards from Loch Leven to Glen Lyon(Figure 10.1). This is likely to have influenced the form of the quadratic and cubic surfaces and helped in the production of a ridge with lower values than occurs in reality.

Thirdly, the fitting of trend surfaces to the

Surface	% RSS obtained
Linear	20.99
Quadratic	23.62
Cubic	28.56

Source of variation	Degrees Freedom	% RSS	Mean Square	F	Significance
Total, 125 data points	124				
Due to linear surface with three constants	2	20.99	10.497	16.177	0.1%
Due to residuals over linear surface	122	79.007	0.648		
Due to added quadratic components	3	2.627	0.876	1.364	ns
Due to residuals over quadratic surface	119	76.38	0.642		
Due to added cubic components	4	4.94	1.235	1.988	ns
Due to residuals over cubic surface	115	71.44	0.621		

TABLE 10.4 Analysis of variance for trend surfaces of altitudes of rain gauges in the western Grampians.

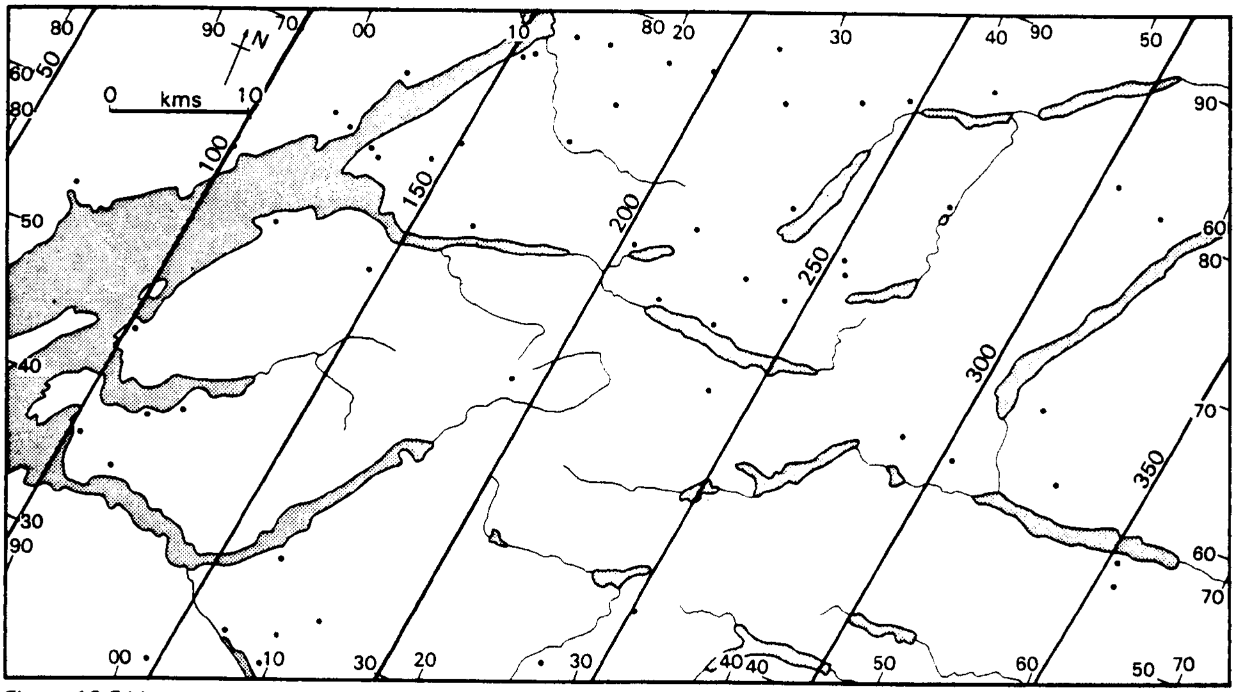


Figure 10.5 Linear trend surface of altitudes of rain gauges in the study area in metres O.D.

altitude of the 125 stations demonstrates a moderate linear trend (Figure 10.5) in the data(the quadratic and cubic surfaces do not provide any significant increases in explanation(Table 10.4)) with the general altitude of the stations increasing in an easterly direction. This means that in trend surfaces such as those shown in Figure 10.4, that are uncorrected for altitude, the precipitation amounts in the E of the study area will be overestimated by an unknown amount in relation to those sites in the W close to sea level.

The pattern of residuals(Figure 10.6) from the cubic surface(Figure 10.4) is broadly similar to that calculated from the multiple linear regression. The main differences occur in the Western Mountain zone where very high positive residuals occur, as for example, for the Ben Nevis station(i.e. 1892mm yr^{-1}) and for station 30 near Ben Cruachan(i.e. 1212mm yr^{-1}). This reflects data from high-altitude stations and emphasises further the problem of fitting trend surfaces to data uncorrected for altitude. To overcome this problem the precipitation data were standardised to sea level(section 10.2).

The trend surfaces depicted in Figure 10.7, that were fitted to the precipitation data standardised to sea level, show broad similarities to those fitted to the uncorrected data (Figure 10.4) although a much steeper decline in the regional component of the precipitation toward the ENE is indicated by the corrected data(e.g. 20mm/km compared with 12.7mm/km for the linear surfaces). In addition the linear surfaces both slope down in a north-easterly direction from maximum precipitation values of 2500-2600mm yr^{-1} in the W. However, the removal of the altitudinal component from the data has led to significant increases in the fit of

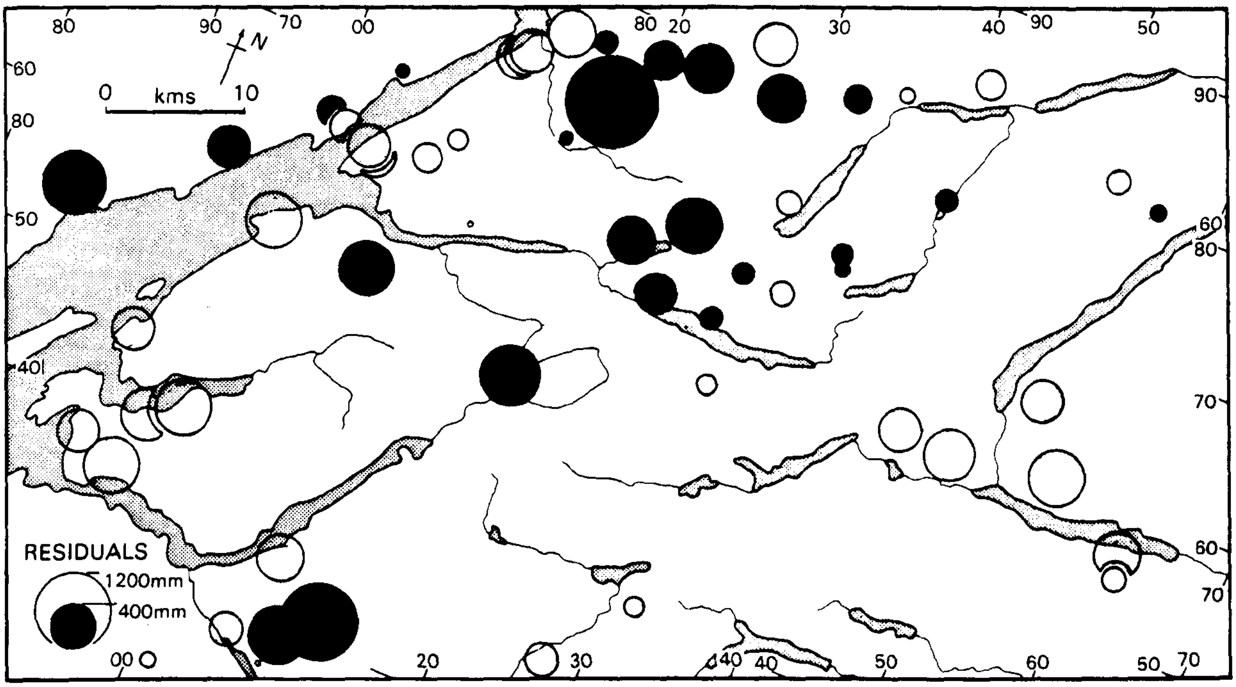


Figure 10.6 Residuals in millimetres of precipitation from the cubic trend surface (uncorrected for altitude)
● Positive ○ Negative

Surface	% RSS obtained
Linear	63.6
Quadratic	75.0
Cubic	81.0

Source of variation	Degrees Freedom	% RSS	Mean Square	F	Significance
Total, 125 data points	124				
Due to linear surface with three constants	2	63.6	31.81		
				106.64	0.1%
Due to residuals over linear surface	122	36.4	0.298		
Due to added quadratic components	3	11.37	3.791		
				18.032	0.1%
Due to residuals over quadratic surface	119	25.0	0.21		
Due to added cubic components	4	6.1	1.527		
				9.285	0.1%
Due to residuals over cubic surface	115	18.9	0.164		

TABLE 10.5 Analysis of variance for trend surfaces of precipitation standardised to sea level.

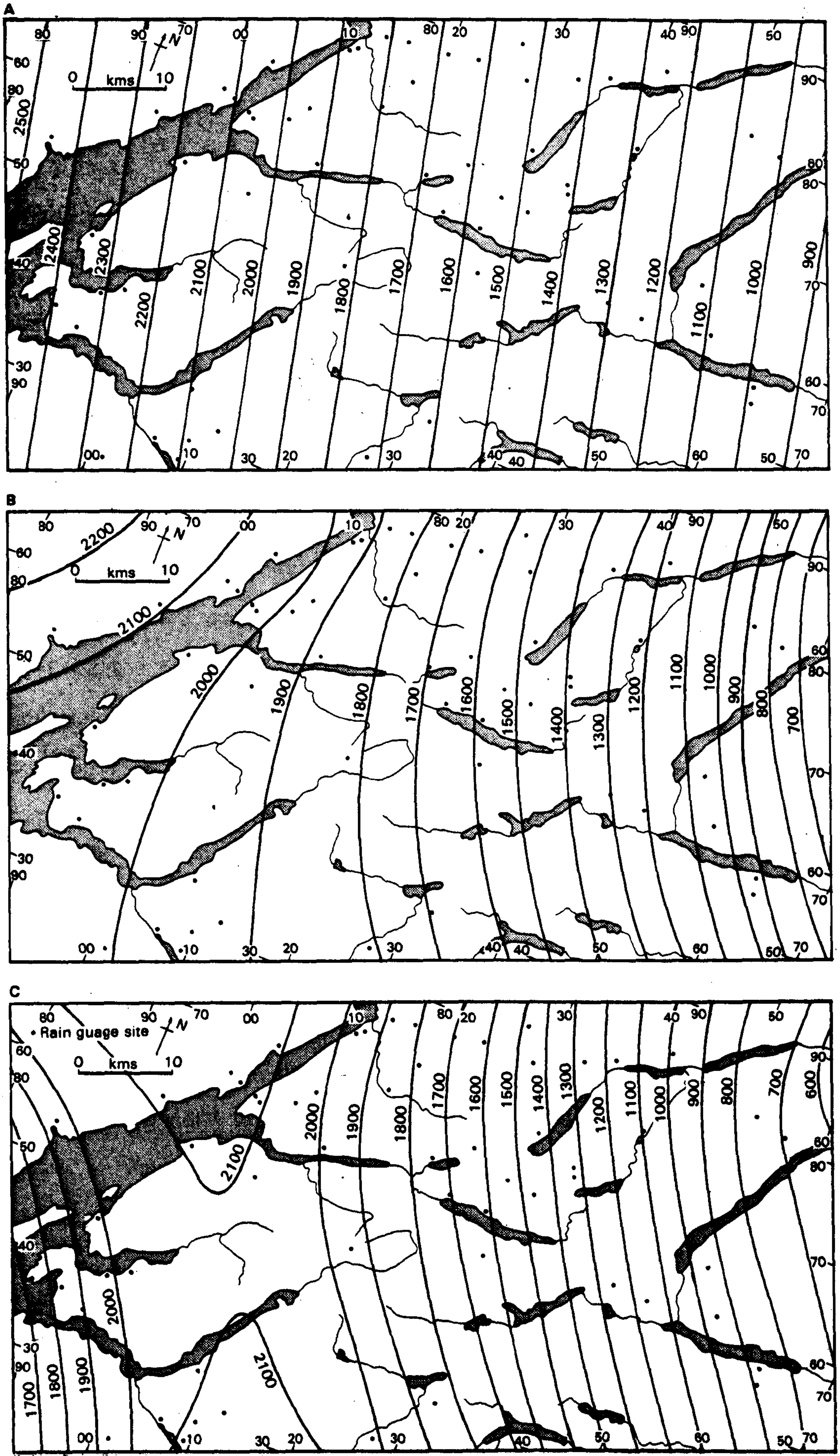


Figure 10.7 Trend surfaces for precipitation standardised to sea-level. Isohyets shown in millimetres.
A Linear trend surface
B Quadratic trend surface
C Cubic trend surface

the trend surfaces shown in Figure 10.7. For example, the level of explanation is increased by a 33.5% reduction in the sums of the squares between the two linear surfaces(Tables 10.3 and 10.5). This reflects, to a large extent, the variations in the altitudes of the rain gauges across the area described earlier.

It is of interest to note that the cubic surfaces are of very similar form with precipitation declining eastward and westward from a broad ridge aligned over or close to the Western Mountain zone. The significance of this spatial pattern of maximum precipitation is discussed particularly in sections 11.7 and 12.5.

An isohyetal map for the study area, drawn subjectively from the information provided by the distributional pattern of the actual precipitation data and from the results of the methods outlined previously, is shown in Figure 10.8. The approach to drawing such a map incorporated the following methods and assumptions:

i) The actual precipitation values placed strict controls on the delimitation of the isohyets, particularly in areas with numerous point-control information.

ii) The cubic trend surfaces were taken as representing reasonably accurately the zone of maximum precipitation extending from Ben Cruachan to the area NW of Loch Linnhe(Figures 10.4 and 10.7).

iii) Since precipitation data are unavailable for sites above 600m O.D. in the study area(except for Ben Nevis) precipitation values for a number of mountains exceeding 1000m O.D. were calculated using the value of 192mm yr^{-1} per 100m from the multiple linear regression equation(Table 10.2.c). The computed values shown in Table 10.6 were used in the drawing of isohyets,particu-

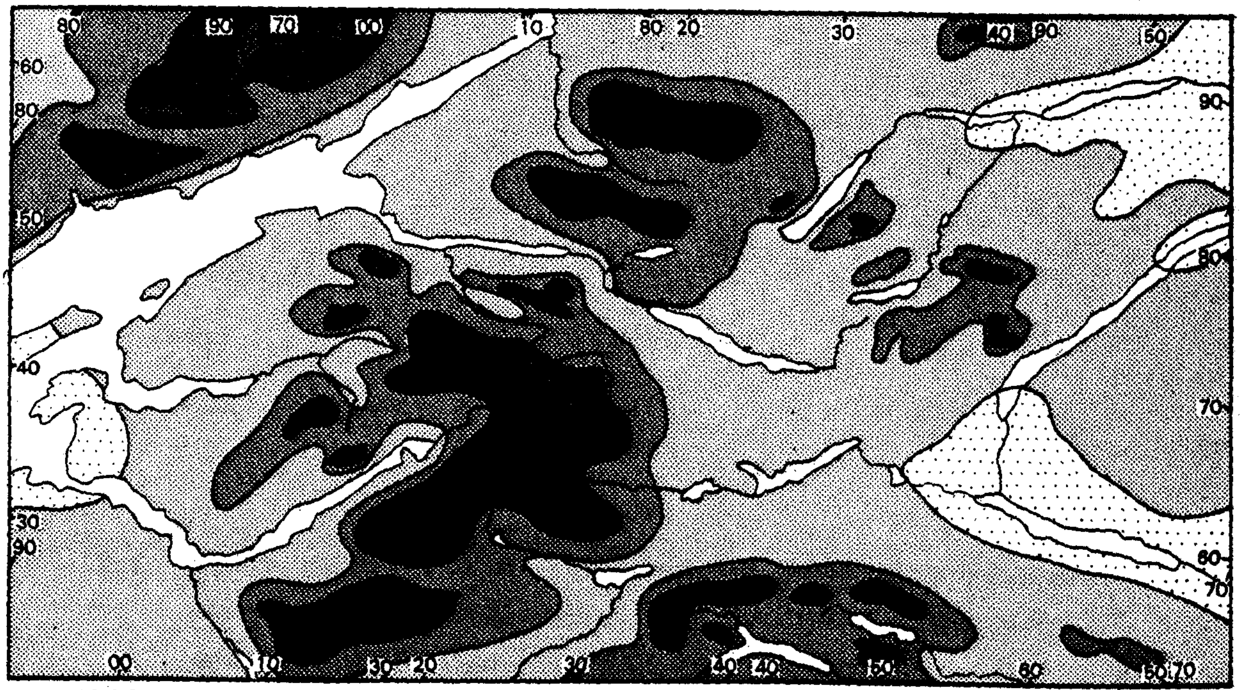


Figure 10.8 Distribution of average annual precipitation in the study area for the years 1941 to 1970. (Compiled from data provided by the Meteorological Office, Bracknell.)

larly on the higher ground where data are lacking.

Mountain	Distance E and N of arbitrary point in SW(km).		Calculated precipitation value (mm yr^{-1})
	E	N	
Ben Cruachan	28	16	4045
Ben Starav	35	28	3862
Bidean nam Bian	35	40	3812
Binnein Mor	43	52	3611
Meall a Bhuridh	48	50	3524
Beinn Doran	54	38	3461
Creag Meagaidh	64	74	3121
Ben Alder	71	58	3056
Carn Gorm	85	36	2883

TABLE 10.6 Precipitation values calculated for selected mountains exceeding 1000m O.D. using multiple linear regression methods. An altitude of 1000m O.D. is assumed in the calculations.

iv) The morphology of the mountain blocks was assumed to strongly control the actual form taken by the isohyets.

The spatial pattern of amounts of precipitation across the area shown in Figure 10.8 reflects the synoptic importance of air of south-westerly origin. The increase in precipitation by a factor of 2 to 3 times from the E side of Loch Linnhe to the Western Mountain zone can be related to numerous depressions tracking E across or near to Scotland and to the orographic influence of the high mountains. Heavy precipitation from warm fronts with S to SE winds,followed by long periods of precipitation associated with winds from SW and W directions is inferred. However, the intensification of precipitation along a warm front created by orographic uplift is likely to be less than for the warm sector of a depression or for a Polar Maritime airstream where a neutral lapse rate can result in the release of conditional instabilty and a dramatic increase in precipitation(Smithson,1969,1970).

Most noticeable are the rain shadow effects of the mountains. In the W, although rain shadow effects are important(e.g. in the Loch Linnhe area) precipitation is intensified and prolonged even on the lee sides of mountains, particularly where the mountains are in close proximity and separated by only narrow valleys. Farther E where the valleys are broader and the mountain blocks more isolated(e.g. the Ben Alder massif) the reductions in amounts of precipitation are more pronounced, partly because of the decreasing water vapour content of the air and partly because large-scale subsidence of air into the Laggan and Rannoch valleys can occur, leading to clearer skies and greatly reduced precipitation from westerly airstreams. In the Loch Laggan and Loch Rannoch areas precipitation totals fall to less than 1200mm yr^{-1}(appendix D, stations 76 and 105).

The contribution to precipitation totals by airstreams from the N and E can be assumed to be low, partly because they are often associated with high pressure conditions and partly because such airstreams will have lost much of their moisture crossing the central and eastern Grampians.

CHAPTER 11

CORRIES IN THE WESTERN GRAMPIANS: AN ANALYSIS OF SELECTED PARAMETERS AND PALAEOCLIMATIC INFERENCES

11.1 Introduction

A well-developed corrie is one of the most easily recognised of glacial landforms with its steep backwall and characteristic armchair shape. Nevertheless, problems of correctly identifying corries in the field or even of finding a common acceptance of a verbal definition of a corrie exist(Evans and Cox,1974). In fact corries form a continuum of features that range from shallow nivation hollows to well-defined corries with tarns, terminal rock bars and rock walls that ascend to surrounding arêtes and horns. Furthermore, a study of 100 corries in Scotland by Haynes(1968) demonstrated that a rock basin and an L-shaped break between the headwall and basin, that are often believed to be representative of a well-developed corrie,were in fact relatively uncommon.

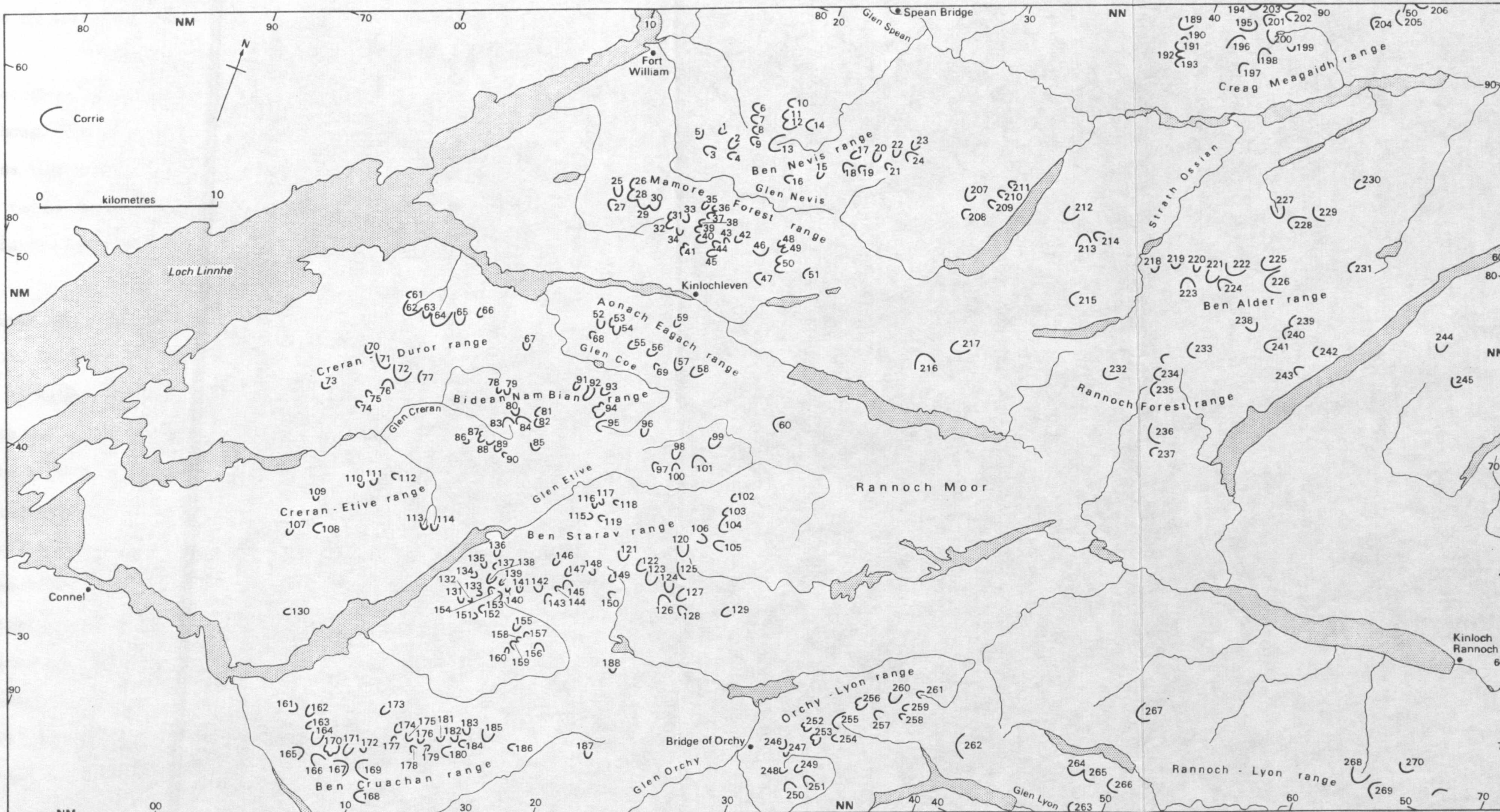
A number of studies of the morphometry of corries have attempted to isolate the different factors involved, such as preglacial relief, lithology and structure,aspect, glacial processes and stage of development(Harker,1901; Battey,1960; Linton,1967; Haynes, 1968; Sugden,1969; Gordon,1977; Olyphant,1977), although difficulties are apparent and the exact relationships between the various factors are still imperfectly understood. For example, Battey(1960) noted that the shape and position of the corries of Vesl-Skantbotn and Veslgjuv-botn in the Jotunheim of Norway were not influenced at all by lithology and structure,whereas Unwin(1973) noted that in Snowdonia the shape and location of a number of corries were strongly influenced by lithology and structure.

More recently, a good deal of attention has been focused on the relationships between past climatic parameters and the formation of corries and corrie glaciers, with increasing use made of trend surface analysis techniques(Andrews, et al, 1970; Unwin,1973; Trenhaile,1975; Meierding,1982). It is this aspect that will largely be explored in the following sections, which deal with the analysis of 271 corries identified in the study area.

11.2 The identification and measurement of corries

Initially, corries were identified on O.S. maps to scales of 1:10,000 and 1:10,560, mainly in the western sector of the area under study. Elsewhere O.S. maps to a scale of 1:25,000 were used to identify corries. All locations of corries were marked on base maps and checked, wherever possible, in the field. Contour patterns and lochans where present were the main criteria employed in the identification based on map evidence. A few small corries, not recognised on the O.S. maps were identified in the field and these were subsequently added to the base maps. The locations of the total of 271 corries identified in the area of study are shown in Figure 11.1.

A major problem in the highly dissected terrain of the western Grampians is that presented by coalescing corries. If several hollows coalesce to share the same threshold, but are separated by distinct spurs they are taken to represent individual corries (Evans,1969). If the spurs only constitute minor buttresses along a corrie wall the feature is regarded as a single corrie. A number of



11.1 Location of Corries in study area. Corries to West of Loch Linnhe not shown.

shallow hollows that lacked the depth and steep sides characteristic of glacially modified hollows were omitted, particularly if they occupied a location at the head of a valley, as these might be strongly influenced by pre-glacial relief.

Approximately 90% of the total number of 271 corries were inspected in the field, usually as far as the foot of the backwall, although some were observed from a distance, but this was generally less than a kilometre. Corries derived solely from map evidence primarily relate to some areas in the NE quadrant and parts of the N side of the Ben Nevis Range.

Various morphological parameters of the corries were measured using either 1:10,000 or 1:25,000 O.S. maps(Figures 11.2 and 11.3). These were:

i) height of corrie floor(H) above mean sea level, taken either from the surface height of a lochan where present or from the base of the backwall where the gradient increases sharply(see below);

ii) maximum width perpendicular to length(W);

iii) maximum length(L) defined as the horizontal distance between the base of the backwall and the lip of the corrie. The lip is identified by the presence of a lochan or where the down-valley gradient increases as indicated by the decreasing distance between the contours;

iv) depth of corrie measured as the vertical distance between the base and crest of the backwall(D);

v) horizontal distance between the base and crest of the backwall(B); and

vi) azimuth of corrie long axis(A), measured in degrees E of grid N.

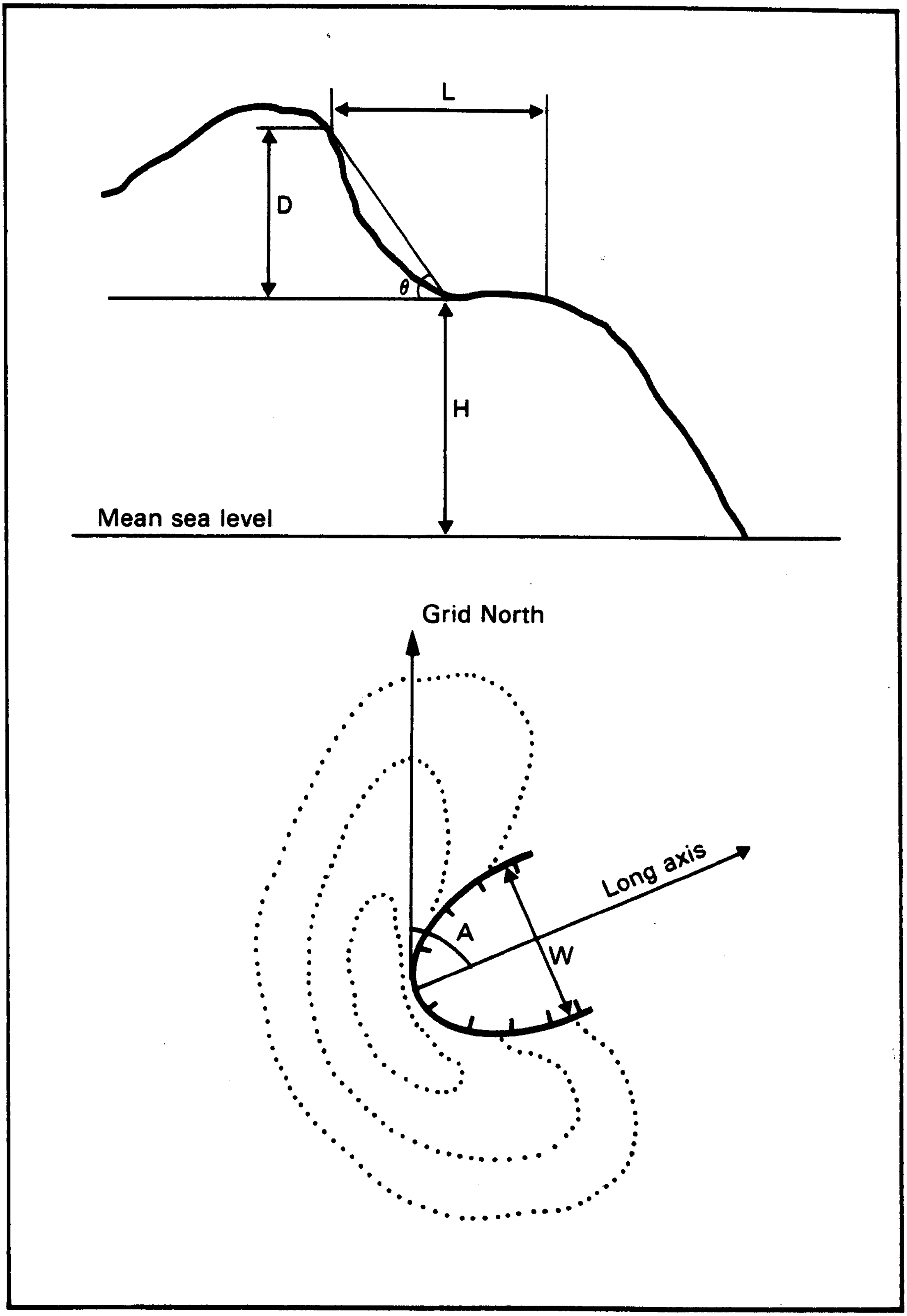


Figure 11.2 Corrie parameters used in the study

A: Azimuth of corrie axis measured in degrees E of grid north
H: Height of corrie floor above mean sea level
W: Maximum width of corrie
L: Corrie length
D: Height of corrie backwall (depth of corrie)
θ: Angle of slope of corrie backwall

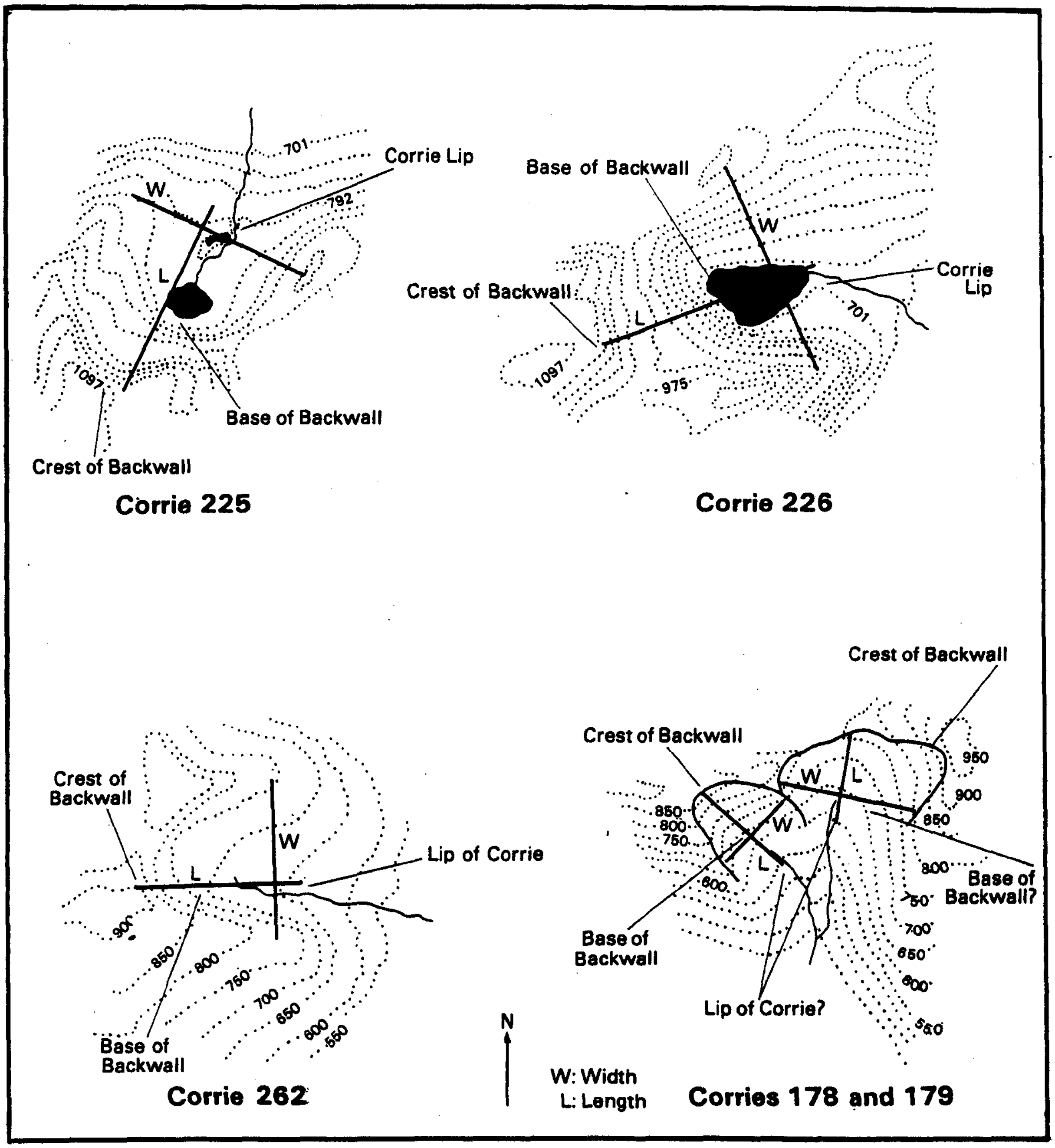


Figure 11.3 Examples of selected corrie parameters measured from O.S. maps to scales of 1:25 000 (Data listed in Appendix B). Corries 225 and 226 represent large well-developed forms; Corrie 262 is a moderately developed corrie and Corries 178 and 179 are small poorly-developed forms.

One measurement problem that has generally been overlooked in previous studies is that of accurately locating the corrie lip,or the break of slope at the base of the backwall,in corries that have outward-sloping floors rather than level floors, particularly if only O.S. map evidence is available(See corries 178 and 179 in Figure 11.3). Inspection in the field usually resolved the problem, but in a few cases the problem remained unsolved. In these cases,and for a small number of corries that were based entirely on map evidence,the altitude of the base of the backwall was taken as the point where the down-slope gradient was at a minimum,as represented by the greater distance between two contours. Similarly the altitude of the corrie lip was taken as the point where the down-slope gradient increases, as shown by the decreasing distance between adjacent contours. However, a degree of subjectivity must inevitably enter into such decisions.

Some of the derived values were used to calculate length/width(L/W) and length/depth(L/D)ratios, the mean angle of slope of the corrie backwall(expressed as the tangent of D divided by B) and the corrie volume(estimated by multiplying L, W and D and dividing by 2).

All data relating to the measured and calculated corrie parameters are listed in Appendix B and are shown graphically as a rose diagram in Figure 11.4 and in the form of histograms in Figures 11.5 and 11.6.

For the purposes of comparison and for some of the statistical tests the total population of 271 corries is divided into two sub-populations. These are divided into western and eastern sub-populations by Eastings grid line NN230. In addition the data are disaggregated into individual mountain ranges in order to detect any possible spatial variations that might exist between sub-areas in the area of study.

11.3 Corrie azimuths

The data on the orientations of corrie axes (Appendix B and Figure 11.4) show that,although they almost span the entire range of directions,a chi-squared test indicates that there is a very strong concentration in the NE quadrant,that is significant at the 0.001 level. This pattern is supported by all the resultant vectors for the thirteen mountain ranges(Table 11.1) and by the overall preferred orientation value of 47.9°E calculated for the 271 corries. This value is similar to the values obtained in the majority of studies of mean corrie azimuths in the Lake District, Wales and Scotland(Table 11.2).

Although the overall preferred orientation is toward the NE the distribution shown in Figure 11.4 is distinctly bimodal with modal classes in the N($0 - 10^{\circ}$) and E($81 - 90^{\circ}$) sectors. In part these distributional peaks may reflect differing controls in different areas. For example, a number of narrow mountain ranges in the western Grampians such as the Mamore Forest Range, the Creran-Duror Range, the Aonach Eagach Range, the Ben Cruachan Range and parts of the Bidean nam Bian Range are aligned in an E - W direction. This pattern of relief appears to have favoured the retention of snow on the N side of the ranges and the development,especially, of N-facing corries. This pattern is exemplified by the line of seven N-facing corries cut into volcanic and quartzite bedrock on the N side of the Aonach Eagach Range. Farther E the individual mountain blocks and

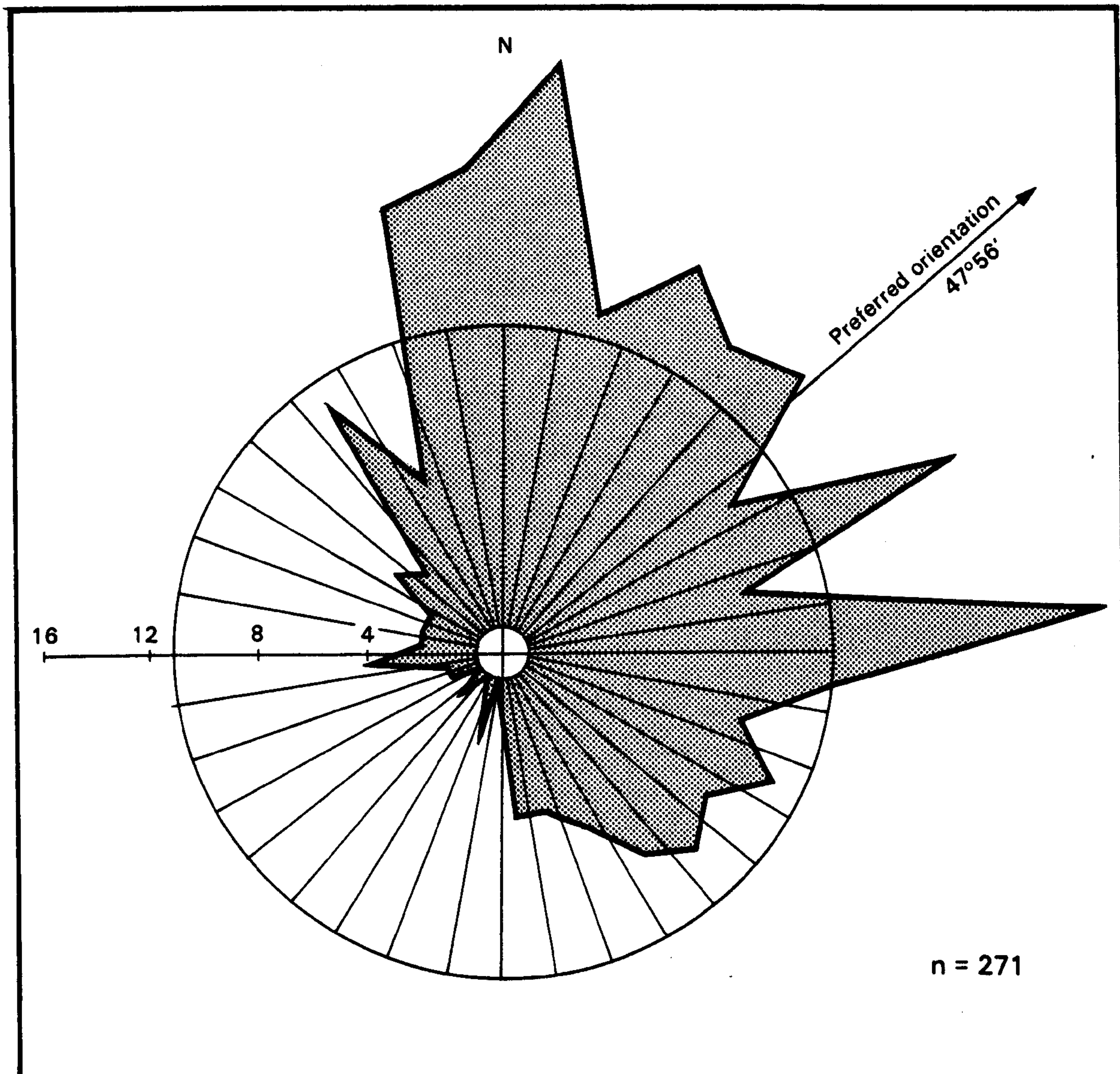


Figure 11.4 Orientations of corries in the study area. The scale represents the number of corries in each 10° sector.

Mountain range	n	Resultant vector (degrees from N)	Vector magnitude	Significance
Western sector				
Ben Nevis	24	61.8	68.0	.001
Mamore Forest	27	44.1	54.6	.001
Aonach Eagach	10	26.9	90.4	.001
Creran-Duror	15	30.1	69.3	.001
Bidean nam Bian	25	40.0	45.9	.01
Creran-Etive	9	23.8	76.9	.001
Ben Starav	51	29.2	31.2	.01
Ben Cruachan	27	33.3	41.5	.01
All corries in W	188	39.5	47.9	.001
Eastern sector				
Creag Meagaidh	18	75.3	41.4	.05
Ben Alder	31	60.2	54.0	.001
Rannoch Forest	8	74.9	72.7	.001
Orchy-Lyon	17	59.4	65.1	.001
Rannoch-Lyon	9	54.8	56.9	.01
All corries in E	83	63.9	56.4	.001
All corries in study area	271	47.9	49.5	.001

TABLE 11.1 Analysis of corrie azimuth data calculated by radius vector methods.

Area	Number of corries	Vector mean (Degrees E from N)	Vector magnitude	Source of data
Torridon(North), Scotland	30	031	60	*Bain
Cuillin,Scotland	52	018	32	Harker,1901
Scotland	347	047	54	Sissons,1967
NW Scotland	437	048	51	*Godard,1965
Scotland	876	044	24	*Sale,1970
Cairngorms,Scotland	30	047	57	Sugden,1969
Kintail-Affric-Cannich,Scotland	231	052	57	Gordon,1977
Western Grampians, Scotland	271	048	50	Author
Cumbria,England	104	062	28	*Sale,1970
Cumbria,England	198	051	55	*Clough,1974,1977
N.Wales	118	059	33	*Sale,1970
S.Wales	15	030	65	*Sale,1970
Snowdonia,Wales	81	048	57	Unwin,1973

* Quoted in Evans(1977)

TABLE 11.2 Mean corrie azimuths from selected areas in the U.K.

and ranges are generally less dissected and in some cases, such as the Rannoch Forest Range, their alignment is in an approximate N - S direction. This has led to a far higher proportion of E-facing corries than in the W which is reflected by the resultant vectors that range from 54.8° to 75.3°E for the five mountain ranges in the east(Table 11.1). A Mann-Whitney U test indicates that there is a significant difference(at 0.005 level) between the mean azimuths of the corries in the western and eastern mountain ranges.

A number of workers(Svensson,1959; Andrews,1965; Derbyshire and Evans,1976; Evans,1977) have noted that the freedom of orientation of corries decreases in a direction normally associated with decreasing amounts of snowfall. For example, Andrews(1965) noted that in the northern Nain-Okak Range of Labrador corries become increasingly restricted to the NE quadrant, farther inland to the W away from the Atlantic Ocean. A similar pattern related to increasing distance from the Atlantic Ocean, with a relatively random pattern of orientation of corrie axes in the W becoming more restricted in the E was observed by Svensson(1959) for the middle Scandes area of Sweden.

Examination of the vector magnitude values(these provide a measure of the degree of clustering of the azimuths around the resultant vector; low values occur when the distribution is scattered widely around the resultant vector while the converse is true for azimuths clustered closely around the resultant vector) for each mountain range indicate that a wide range of values from 31.2 to 90.4% is present(Table 11.2). The very high value(L = 90.4%) for the Aonach Eagach Range reflects the strong controls exerted by topography as described earlier. Most of the vector magnitudes below values of 50%

occur in the W. For example, the Ben Starav Range has the lowest value(l = 31.2%) in the area of study, which may reflect the relative uniformity of the rocks comprising the Etive granite intrusive complex and the varying orientations of the valleys cut into the massif. The lower vector magnitude of 47.9% for the Western Mountain zone in aggregate in comparison with the value of 56.4% calculated for the eastern mountain ranges may also reflect another control, namely that of the general climatic control of decreasing snowfall to the E that led to a consequent reduction in the number of snow traps forming outside the NE quadrant in the eastern part of the study area. However, a Mann-Whitney U test comparing the vector magnitudes of the western and eastern mountain ranges produced a non-significant result, although this result is likely to have been influenced by the varying sample sizes involved. It is probable that local relief and structural effects tend to over-ride the regional climatic controls in some areas, or at least to 'dampen' down such controls in a complex and imperfectly understood way.

11.4 The size and shape of the corries

Various corrie parameters can be utilised to obtain quantitative information on differences in the size and shape of corries, (Andrews and Dugdale,1971; Gordon,1977), although such differences may not necessarily permit a simple analysis of the causative factors involved. The width,length and depth of a corrie, as defined previously (p.260), are three such useful parameters.

Figure 11.5 presents the widths,lengths and depths of the 271 corries in the form of histograms. All three distributions are unimodal and positively skewed with a wide range between extreme

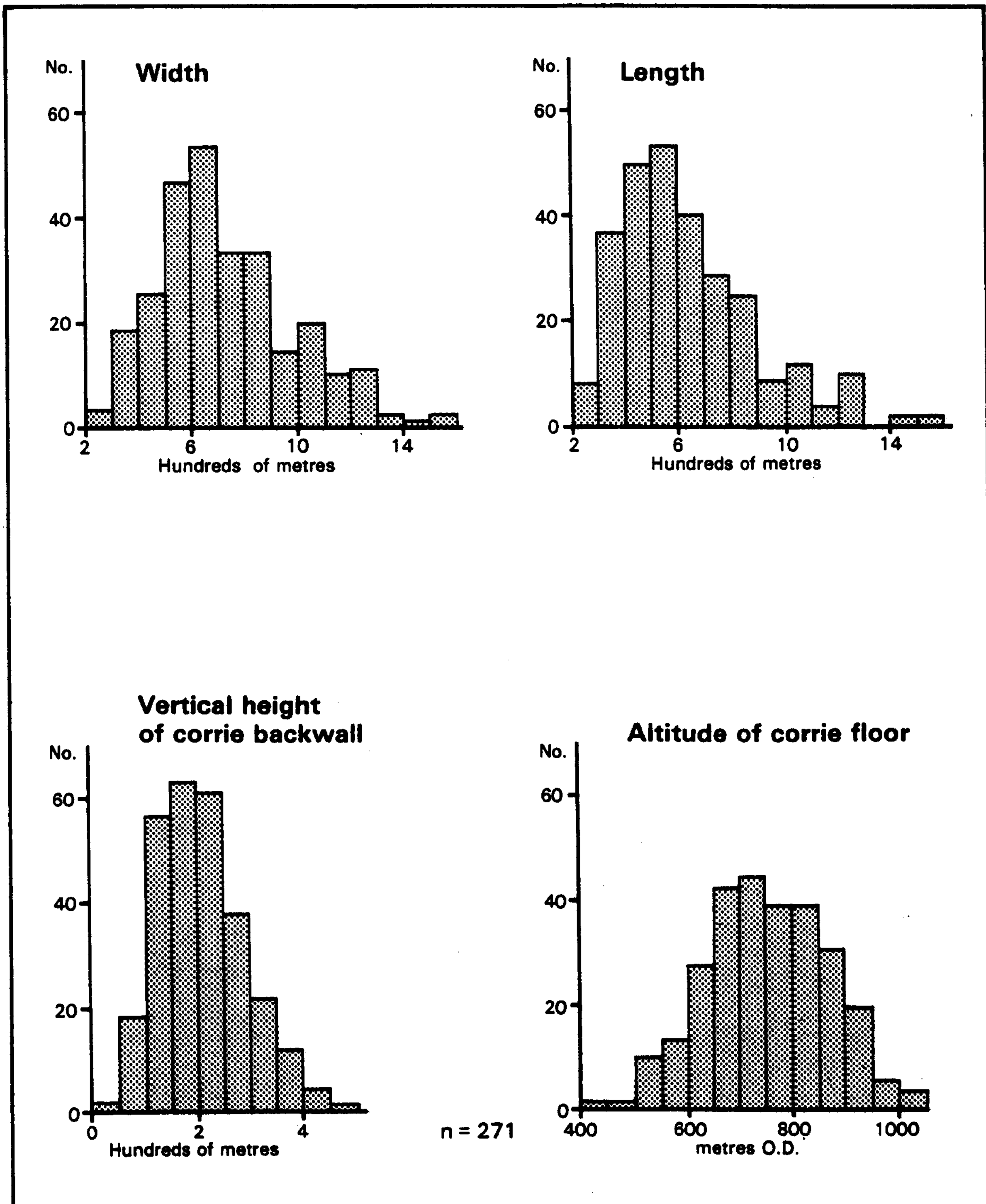


Figure 11.5 Corrie floor altitude, length, width and height of backwall in the study area.

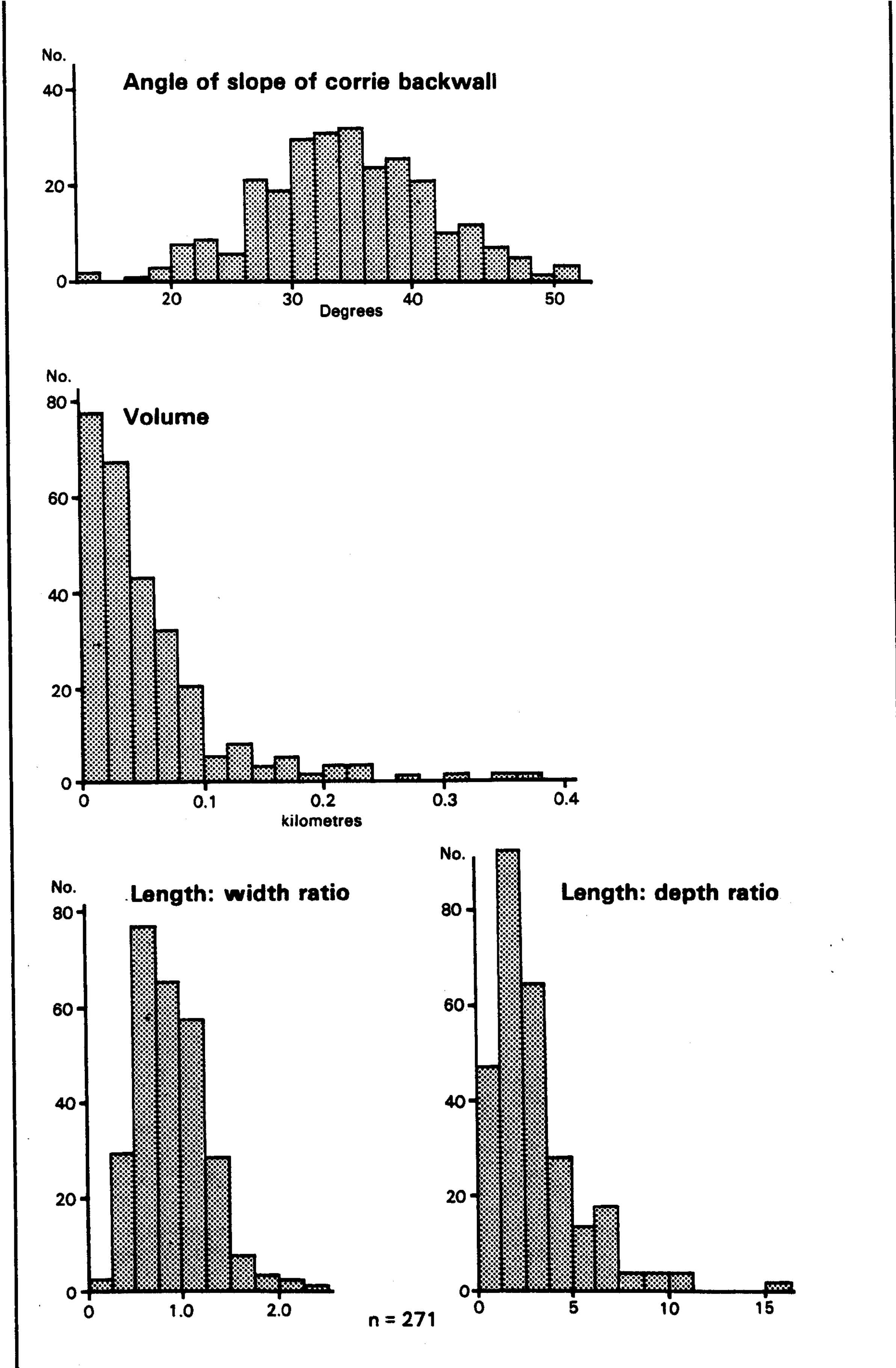


Figure 11.6 Corrie backwall angles, volumes and length/width and length/depth ratios

Mountain range	n	H (m)	W (m)	L (m)	D (m)	θ	L/W	L/D	V (km^3)
Western sector									
Ben Nevis	24	901.3	684.4	584.8	219.1	37.8	0.862	3.159	0.046
Mamore Forest	27	781.9	638.5	537.0	205.8	37.2	0.898	2.760	0.042
Aonach Eagach	10	718.2	683.0	564.0	186.5	35.1	0.842	2.987	0.042
Creran-Duror	15	594.7	642.7	498.0	200.3	33.7	0.832	2.967	0.029
Bidean nam Bian	25	736.3	591.6	456.4	194.6	36.0	0.85	2.670	0.03
Creran-Etive	9	627.7	586.7	593.3	144.9	30.7	1.014	4.212	0.028
Ben Starav	51	716.1	648.8	583.7	191.6	32.2	0.913	3.544	0.042
Ben Cruachan	27	645.19	694.1	725.6	272.2	33.9	1.070	2.798	0.084
All corrie in the W	188	727.9	652.5	573.2	207.3	34.6	0.913	3.115	0.047
Eastern sector									
Creag Meagaidh	18	795.5	870.0	631.1	209.5	33.6	0.749	3.087	0.073
Ben Alder	31	817.8	816.1	717.1	188.3	32.8	0.874	4.229	0.073
Rannoch Forest	8	664.4	1171.3	806.3	197.5	27.1	0.705	4.300	0.115
Orchy-Lyon	17	753.0	705.3	615.9	196.1	35.1	0.968	3.878	0.051
Rannoch-Lyon	9	752.2	936.7	848.9	184.0	31.9	0.914	4.792	0.075
All corries in the E	83	757.0	852.4	700.6	194.9	32.8	0.854	3.977	0.073
All corries in study area	271	736.8	713.7	612.2	203.5	34.0	0.895	3.379	0.055
Significance levels (Mann-Whitney U test)		ns	.001	.005	ns	.1	ns	.01	.01

TABLE 11.3 Summary statistics of corrie parameters: mean values. Values above the overall mean are shown underlined.

values(W = 200-1500m; L = 250-1500m; D = 38-456m). In all three histograms the mean value and the modal class show a close correspondence.

Disaggregation of the data according to the 13 mountain ranges(Appendix B and Table 11.3) indicates that the mean values of W, L and D vary by a factor of approximately 2. The particularly high mean value of 1171.3m for corrie widths in the Rannoch Forest Range reflects the difficulties of accurately defining the limits of relatively poorly-defined corries along the long rockwall on the eastern side of the range.

The mean values of widths and lengths are similar to those recorded by Graf(1976) for corries in the Rocky Mountains that lack glaciers at the present time, although he found that corries that contain glaciers today generally have greater mean widths. In contrast Gordon(1977) recorded generally lower widths for corries in the Kintail-Affric-Cannich area of Scotland(Table 11.4). The mean depths of corries in the western Grampians tend to be less than in both NW Scotland(Gordon,1977) and in the Rocky Mountains, where mean values of corrie depths for individual mountain ranges vary from 200 - 574m (Graf,1976). In the area of study the overall mean depth of the corries is <u>ca</u> 200m while extreme values range from 38 to 456m. The low mean value of 144.9m(SD = 32.73m) for depths of corries in the Creran-Etive Range may be a result of intense over-riding by ice under ice-sheet conditions(p.322). The mean widths and lengths of corries in the E are significantly greater than those in the W at levels of 0.001 and 0.005 respectively(Mann-Whitney U test). Possible reasons for these differences will, however, be deferred until some of the other corrie parameters have been described.

	Range	Mean	Median	Mode
Corrie floor altitude(m)	400-1036	744.4	743.5	700-749
Width(m)	200-1500	713.7	660.0	600-699
Length(m)	250-1500	612.2	570.0	500-599
Depth(m)	38-456	203.5	198.0	150-199
Backwall angle (degrees)	12-53	34.0	34.0	34-35.9
Length : width (ratio)	0.28-2.0	0.895	0.860	0.6-0.8
Length : depth (ratio)	1.13-13.5	3.379	2.907	2.0-3.0
Volume(km^3)	0.002-0.336	0.055	0.037	0-0.02

A. Corries in the western Grampians. n = 271

	Range	Mean	Median	Mode
Corrie floor altitude(m)	106-1006	673.5	688.0	600-650
Width(m)	130-2250	586.5	530.0	300-400
Length(m)	100-1840	625.1	540.0	200-400
Depth(m)	92-670	275.6	265.0	150-200
Backwall angle (degrees)	18-73	44.6	45.0	42-45
Length : width (ratio)	0.37-4.05	1.12	1.05	1.0-1.2
Length : depth (ratio)	0.84-4.96	2.21	2.07	1.8-2.0
Volume(km^3)	0.001-0.489	0.064	0.036	0-0.025

B. Corries in the Kintail-Affric-Cannich area(Gordon,1977). n = 231

TABLE 11.4 Comparisons between corrie parameters for the study area and the Kintail-Affric-Cannich area of NW Scotland.

Mean values of the angle of slope of the corrie backwalls range from 27.1^{o} in the Rannoch Forest Range to 37.8^{o} in the Ben Nevis Range, with an overall distribution that approaches a normal distribution very closely(Figure 11.6). The individual backwall angles listed in Appendix B are in fact mean angles,since they include the generally less steep slope forming the lower part of the backwall(Haynes,1968),as well as the steep upper part of the backwall. Hence the angles appear less than might be anticipated.

The derived values are similar to those recorded by Haynes(1968) for corrie backwall angles in different areas of Scotland and to those recorded by Gordon(1977) for corries in NW Scotland, except that a number of backwall angles exceeding 60^{o} were recorded by Haynes and Gordon, whereas none greater than 55^{o} was measured in the study area. However, this difference is likely to relate to differences in the measurement and compilation of the data, rather than to inherent differences in corrie shapes as,for example,Haynes divided the angles into two groups(1968,Figure 4, p. 225) comprising backwall and backwall-foot angles that makes direct comparison difficult.

Examination of the mean values of backwall angles on a range-by-range basis shows that angles are generally greater in the western mountain ranges, as for example in the Ben Nevis, Mamore Forest and Bidean nam Bian ranges. The difference in mean backwall angles between the western and eastern mountain ranges is significant at the 0.01 level(Mann-Whitney U test) and this may reflect the contrasts in lithology between E and W across the study area. In the W some of the steepest backwalls are associated with volcanic and quartzite rocks, as exemplified by the high, N-facing,vertical rock-walls(corries 1 and 2) below the summit of Ben Nevis. Farther E the relatively uniform psammites and granulites give rise to generally less steep backwalls. The effects of lithology are also apparent in the lower mean backwall angles recorded for the Creran-Etive and Ben Starav ranges(e.g. 30.7° and 32.2° respectively). Many of the corries in these areas have been cut into Cruachan and Starav granites that display very strong pressure-release jointing sets. In many cases the jointing sets have influenced the angle of slope of the backwall and of the corrie floor, particularly where the joints dip down-corrie at angles mainly between 10° and 30°, a feature noted by Haynes(1968) for corries in other granite areas such as Arran and the Cairngorms. This would seem to explain why corries 113,145,148,152,153,156 and 158 all have backwall angles of less than 25°.

A considerable range exists in the size of corries in the area of study, as demonstrated by corrie volumes that range from 0.002 to $0.336km^3$(Table 11.3 and Appendix B). However, 88% of the corries have volumes less than $0.1km^3$ and this is reflected by a mean value of $0.055km^3$ for all the 271 corries. Furthermore, some of the largest corries may not represent true corries since they mainly lie at the heads of valleys and may be of more complex origin. Such an example is corrie 200 that lies at the western end of a short glacial trough in the Creag Meagaidh Range and is overlooked by steep crags and by several perched corries at higher altitudes. The corries with the lowest mean volumes of below $0.03km^3$ occur in the extreme W in the Creran-Duror and Creran-Etive ranges; farther E the mean size of the corries increases to over $0.07km^3$(Table 11.3). This contrast in corrie volumes between the western and eastern mountain ranges is statistically significant at the 0.01 level(Mann-Whitney U test).

The reasons for the differences in the mean sizes the corries are difficult to explain as there many uncertainties. Embleton and King(1975) have suggested that some of the most important variables will be rock type, the size of the mountain mass on which the corrie develops and the length of time during which the corrie formed. Others(Sugden and John,1976; King,1980) have emphasised climatic differences,as well as the duration of glaciation, with the possibility that the largest corries may develop in areas of high altitude, high precipitation and where corrie glaciers are highly active. For example, King(1980) has pointed out that the corries in Alaska are greater that those in the San Juan Range of the Southern Rocky Mountains by two orders of magnitude : differences that are ascribed to the contrasting oceanic and continental climates of the two areas. Graf(1976) noted a similar regional pattern in the Rocky Mountains in the mean size of the corries, with the width and depth parameters being greatest in the northern ranges of the Rocky Mountains.

The greater mean widths,lengths and volumes of the corries in the eastern sector of the study area could relate to the decrease in snowfall with increasing distance from the Atlantic Ocean. It must be noted though that this proposal is virtually opposite to the observations of Sugden and John(1976) and King(1980) described above. However, an additional factor over-riding those cited by the above authors may have been operative in the western Grampians. In the E fewer favoured sites for the retention of wind-blown snow would have been present(p.256) and this may have allowed the corries to develop to a larger mean size than in the W, where the development of numerous corries adjacent to each other interferred with their mutual growth in size and generally restricted the size of the area from which snow was blown into the corrie. For example, fifteen corries occur on the flanks of Ben Starav with mean volumes of only $0.031km^3$. Such differences do not apply to the corries of the Ben Cruachan Range where mean widths, lengths and volumes are similar to those in the E. The reasons for this are unknown at the present time.

Sugden(1969) attributed different-sized corries in the Cairngorm mountains to differing age periods of development. However, there is no positive evidence in the study area to suggest different generations of corries based either on size or on height above sea level.

The planimetric shape of corries may be described quantitatively by the length/width ratio. A ratio of unity indicates an equidimensional corrie. A corrie with a high L/W ratio will be long and narrow, while a low L/W ratio will reflect a short, wide corrie.

Mean L/W ratios are less than unity for all the mountain ranges except for the Creran-Etive and Ben Cruachan ranges. This pattern is similar for glacierized corries, but not for empty corries in the Rocky Mountains at the present time(Graf,1976). These data demonstrate that, despite highly complex geology, varying degrees of glacial dissection and erosional modifications by over-riding ice-sheets, corries in the western Grampians tend to have a similarity of plan.

The length/depth ratio, referred to elsewhere as the length/height ratio(Andrews,1965; Embleton and King,1975; Graf, 1976) or as the length/amplitude ratio(Gordon,1977) is another shape parameter that can be used to describe the corrie in a vertical section. The smaller the ratio the steeper and higher the backwall is relative to the length; thus where the ratio is small the corrie is more protected from insolation and provides a better shelter for snow accumulation. Manley(1959) claimed that well-developed corries have L/H ratios lying between values of 2.8 and 3.2.

As shown in Figure 11.6 the L/D ratios for the 271 corries in the western Grampians show a high degree of variability with extreme ratio values ranging from 1.13 to 13.5(cf the lower variability recorded by Gordon(1977) for NW Scotland(Table 11.4)). Nevertheless, on a range-by-range basis many of the mean values lie between the limits proposed by Manley, particularly in the western half of the study area. Toward the central Grampians L/D ratios are higher(significant at the 0.01 level) and this may reflect the changes in topography from the deeply,dissected terrain in the W to the broader mountains and plateau-like terrain characteristic of the central Grampians.

11.5 Trend surface analysis

Trend surface analysis techniques generally using computer 'package' programs have been applied to a number of spatially distributed geographical variables(Chorley and Haggett,1969; Smithson,1969; Unwin,1969; Gray,1972,1974b,1978; Doornkamp,1972; Shakesby,1976). Applications that have been specifically related to the varying altitudes of corrie floors include those by Petersen and Robinson(1969), Unwin(1973) and Trenhaile(1975).

The computer programs use standard regression techniques to fit a best-fit surface to spatially distributed values. Input data are required in the form of three variables(Xi, Yi, Zi) where Xi and Yi are locational co-ordinates for a data point and Zi is the corresponding data value. Computation separates each observation into two components, namely a regional trend and local effects (residuals). Large-scale processes generate the regional or trend component which is a function that behaves predictably. The local component operates over smaller areas and also represents random fluctuations and errors of measurement superimposed on the large-scale patterns. In a trend surface local values are damped down and 'sacrificed' to the overall trend.

The aim is to fit trend surfaces of increasing complexity by least squares methods from 1st order(linear) through 2nd order(quadratic) and 3rd order(cubic) to higher order surfaces, sometimes as high as octic(Bassett and Chorley,1971). In practice programmes are rarely run beyond the 3rd order since such higher order surfaces are extremely demanding in their data requirements while some computer programmes do not compute trend surfaces beyond a 3rd order surface. Furthermore, as Unwin(1975) has pointed out it is difficult to imagine any <u>a priori</u> theory that might predict surfaces of an order higher than the quadratic whilst having fitted such surfaces it is equally difficult to come up with any <u>a posteriori</u> theory to account for them.

The basic shapes generated by a quadratic equation comprise a dome,ridge,tongue or saddle. Surfaces generated by higher orders become progressively more complex.

The degree of 'fit' or explanation of a trend surface can be expressed as the % reduction in the total sum of the squares(%R.S.S). This is analogous to the coefficient of multiple regression and can be converted into equivalent correlation coefficients. To test that the obtained fit is not simply a chance phenomena on the two sources of variation, namely the regional trend surface and the residuals, can be separated out to form a variance ratio, F according to the relation:

$$F = \frac{\%RSS \, / \, df_1}{(100 - \%RSS)/df_2}$$

df_1 = number of constants - 1

df_2 = number of sample size - 1 - df_1

Similarly, F ratios can be used to test that a higher order surface is a significant improvement on a lower order surface.

More recently Meierding(1982) has suggested the use of root mean square error(R.M.S.E.), that is residual error from the surface, as a more suitable statistic for expressing absolute data variation than ' percent variance explained'.

Although trend surface analysis has proved to be a useful tool in the analysis of spatially distributed data the method does have a number of limitations. These can be summarised as:

i) There needs to be a sufficient number of data points; at the very least as many as there are constants to be estimated in the equation.

ii) The method assumes that spatial continuity exists, yet data such as the altitudes of corrie floors are discontinuous and located at points in space, although this difficulty can be overcome by regarding the observed corrie floor altitudes as a 'sample' of the continuous climatic snowline where it intersects the ground surface.

iii) In many trend surface studies it is not possible to have control over the sample locations. For example, the altitudes of corrie floors are located at fixed points and if those points are poorly distributed the results can be distorted. Spurious domes, saddles,etc can be produced where there is poor data control. Where data points are in clusters or in lines the estimate of the true %RSS obtained can be unreliable and conventional significance tests may be rendered invalid(Norcliffe,1969; Doornkamp,1972).

iv) Poor data control at the edges of the mapped area can lead to spurious slopes as a result of extrapolation from the controlled area. A 'buffer' zone of data points outside the control area can be established to eliminate this particular problem.

v) The shape of the boundary surrounding the data can affect the form taken by the trend surface. For example, elongated areas tend to produce contours that are aligned parallel to the longer side(Shakesby,1979).

vi) Spatial autocorrelation is almost always present in the residuals and this results in the estimates of the 'unexplained' variance being smaller than they should be which in turn alters the F-distribution in its tabulated form(Unwin,1975).

vii) The selection of the trend surface that 'best' fits the data presents a difficult problem that is not easy to resolve. Some authors, such as Robinson(1970), suggest that trend surfaces should be used as simple descriptive devices for spatial generalisations that might suggest a posteriori theory. Other operators have stopped when a higher order surface provides no significant increase in explanation or when autocorrelation has been eliminated. Chayes' (1970) suggestion that F ratios should be used at each order of increasing complexity and the computing terminated when no significant increase occurs, has been adopted by some workers(Gray,1972; Unwin,1973). It is also the method adopted below in this thesis.

11.6 Previous studies related to trend surfaces of corrie floor altitudes

Various studies(Linton,1959; Porter,1964; Sissons, 1967; Flint,1971) have used the altitudes of corrie floors to construct surfaces that are related to the altitude of a composite Pleistocene snowline that will lie below the former orographic snowlines. More recently trend surface analysis has been employed by a number of workers(Robinson et al, 1971; Andrews et al, 1970; Unwin, 1973; Trenhaile, 1975; Meierding, 1982) as an objective method for contouring corrie floor altitudes. However, there are a number of problems that relate to the varying altitudes of corrie floors that need to be discussed first.

Some workers such as Flint(1971) have suggested that the theoretical surface will not duplicate any single snowline of the past, but instead will relate to former firn lines at different levels created by the ascent and descent of the snowline during the onset and decline of a glacial period. Other workers(Unwin,1973; Trenhaile,1975; Embleton and King, 1975) have stressed that the variability in the altitude of corrie floors over a small area is more likely to relate to local factors such as the pre-glacial relief, geology and aspect. A further complicating factor is that most corries are likely to be composite in origin since they will generally be reoccupied by ice during each glacial. In areas such as the Alps and the Rocky Mountains some corries continue to be occupied by glaciers even during the present warm interglacial(Graf,1976) while many corries remain devoid of ice. Even during cold periods not all the corries may necessarily be occupied by glaciers. For example, during the Loch Lomond Stadial a number of corries failed to nourish glaciers(Sissons,1977a,1977b,1977c,1979a,1979b; Thorp, 1981b). Thus the distributional pattern of corrie glaciers can vary considerably through time and especially as between glacial and interglacial periods. Any surface that is based on the altitudes of corrie floors can only be a crude, composite approximation of the former firn lines. Nevertheless, various studies of the altitudes of corrie floors in areas as far apart as the western U.S.A. (Porter,1964), Tasmania(Petersen and Robinson,1969) and Scotland (Robinson et al,1971) demonstrate that these rise inland from the coast and appear to duplicate a similar rise in firn lines at the time the corries were being formed.

Various sampling methods have been applied in an attempt to overcome some of the problems outlined above. In order to eliminate possible errors created by varying aspects some studies have only sampled N-facing(Porter,1964) or N to E-facing corries (Flint,1971). However, Trenhaile(1975) and Meierding(1982) have suggested that the firn lines of N-facing corrie glaciers may be at least 100m lower than those facing S because of the incidence of lower values of direct insolation. Other workers(Embleton and King, 1975) have suggested that corries strongly influenced by geology or by pre-glacial relief should be eliminated in the sampling procedure, but this can result in highly subjective decisions being taken. Meierding(1982) in a study of glacial equilibrium-line altitudes in the Colorado Front Range only sampled corrie-floor altitudes of the lowest north-to-east facing corrie that appeared to have fully contained a glacier on each 24,000 quadrangle. However, this sampling method is still strongly subjective and it resulted in only a small sample of 12 corries that was poorly distributed (Meierding,1982, Figure 4a).

Some studies have used all the corrie floor altitudes in a given area(Petersen and Robinson,1969; Andrews et al, 1970; Unwin,1973; Trenhaile,1975), but the more corries that are sampled the higher will be the derived firn line gradient. This arises because there is a lower limit to the formation of corries but not an upper limit except that imposed by topography. Thus it is possible that this sampling procedure will produce a surface that can be strongly influenced by the distribution of the height of the land. For example, Unwin's(1973) linear and quadratic surfaces and linear residuals(Figures 3 and 4) relating to the corrie-floor altitudes of 84 corries and to the altitudes of 81 corrie backwall crests in Snowdonia show a marked similarity of form.

11.7 Corrie floor altitudes in the western Grampians and palaeoclimatic inferences

Trend surfaces to a 3rd order were computed using the data for altitudes of corrie floors(Appendix B). The altitudes of corrie floors of 31 corries additional to those listed in the appendix were included in the computations. These values were derived from the area W of Loch Linnhe in order to provide improved control over the surfaces in the NW quadrant. None of the three surfaces gives impossibly low or high values except for a small part of the cubic surface where the values $<$400m are suspect. Figure 11.7 shows the best-fit linear,quadratic and cubic surfaces for the 303

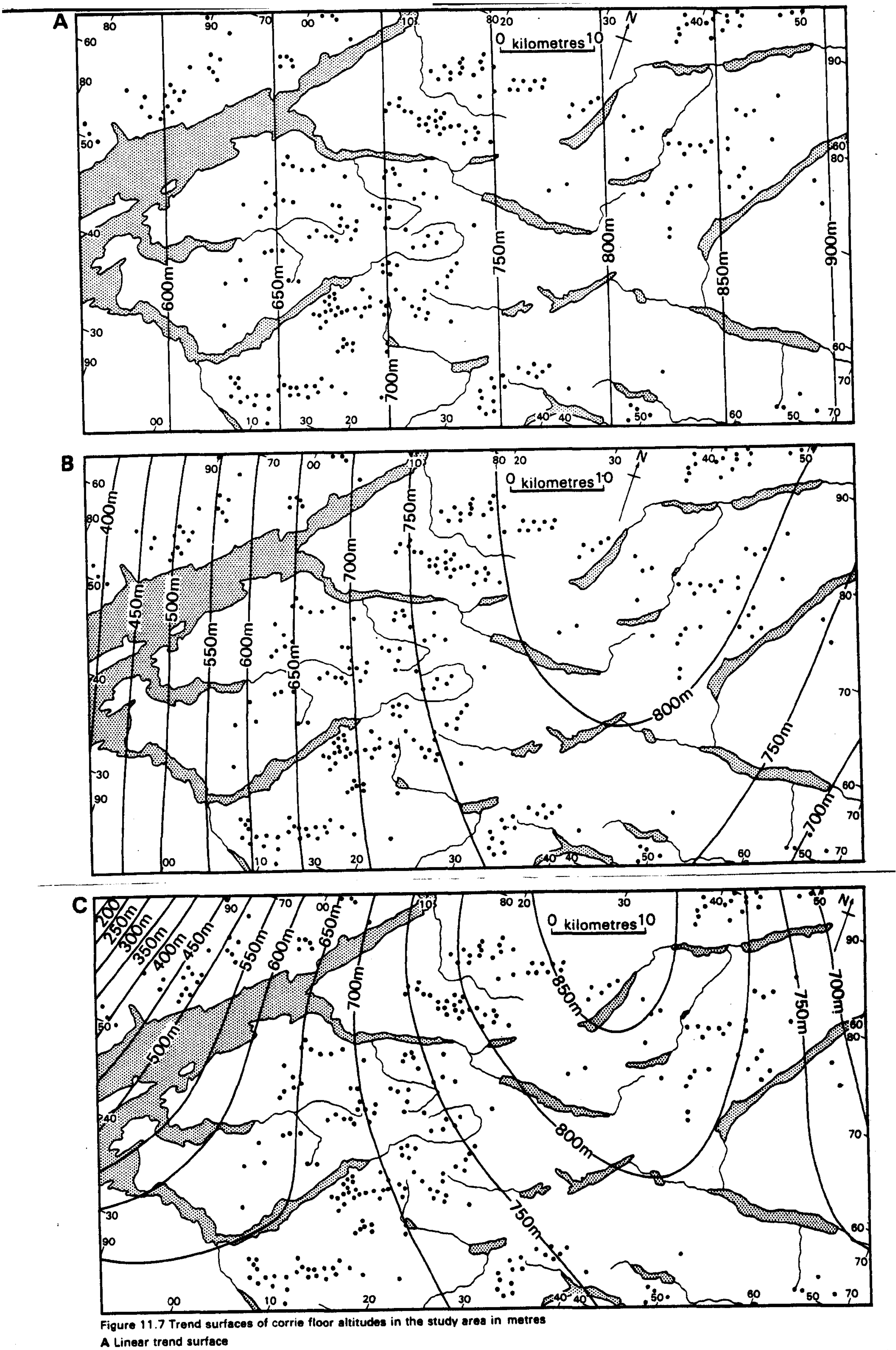


Figure 11.7 Trend surfaces of corrie floor altitudes in the study area in metres
A Linear trend surface
B Quadratic trend surface
C Cubic trend surface

Surface	% RSS obtained
Linear	31.9
Quadratic	45.5
Cubic	49.1

Source of variation	Degrees Freedom	% RSS	Mean Square	F	Significance
Total, 303 data points	302				
Due to linear surface with three constants	2	31.9	15.95	70.033	0.1%
Due to residuals over linear surface	300	68.1	0.227		
Due to added quadratic components	3	13.6	4.533	23.705	0.1%
Due to residuals over quadratic surface	297	54.5	0.184		
Due to added cubic components	4	3.6	0.9	5.181	0.1%
Due to residuals over cubic surface	293	50.9	0.174		

TABLE 11.5 Analysis of variance for trend surfaces of altitudes of corrie floors.

corrie-floor altitudes in the area under study. The results are shown in Table 11.5 which provides the % RSS values and a complete analysis of variance of the surfaces using the method outlined by Unwin(1975).

The linear surface has a moderate fit that trends from less than 600m O.D. in the WSW to over 900m O.D. in the ENE. This simple surface, with a gradient of 4.31m/km, explains 31.9% of the variance in corrie-floor altitudes(Table 11.5).

The quadratic surface is in the form of a ridge aligned in a N - S direction that is centred over the eastern half of the area under study. The gradient of the surface is steeper on the Atlantic side of the ridge than on its eastern side. The fit has improved to give a value of 45.5% which constitutes a substantial quadratic trend in the data. Although the cubic surface gives a further improvement in the fit, which is significant at the 0.1% level, it adds little variation to the quadratic surface. Moreover, the surface in the far W is likely to be partly in error because of the poor distribution of data points close to the western margin of the control area. The ridge in the quadratic surface is emphasised further in the cubic surface and reflects, particularly, the generally high altitude of the corries in Ben Nevis and Ben Alder ranges(Table 11.3).

Figure 11.8 shows the residuals over the linear, quadratic and cubic surfaces for the western Grampian data. Only residuals with values exceeding 100m are shown in Figure 11.8.

The overall pattern of the residuals from the linear surface is irregular, but there is a strong tendency for the clustering together, in particular, of the positive residuals.

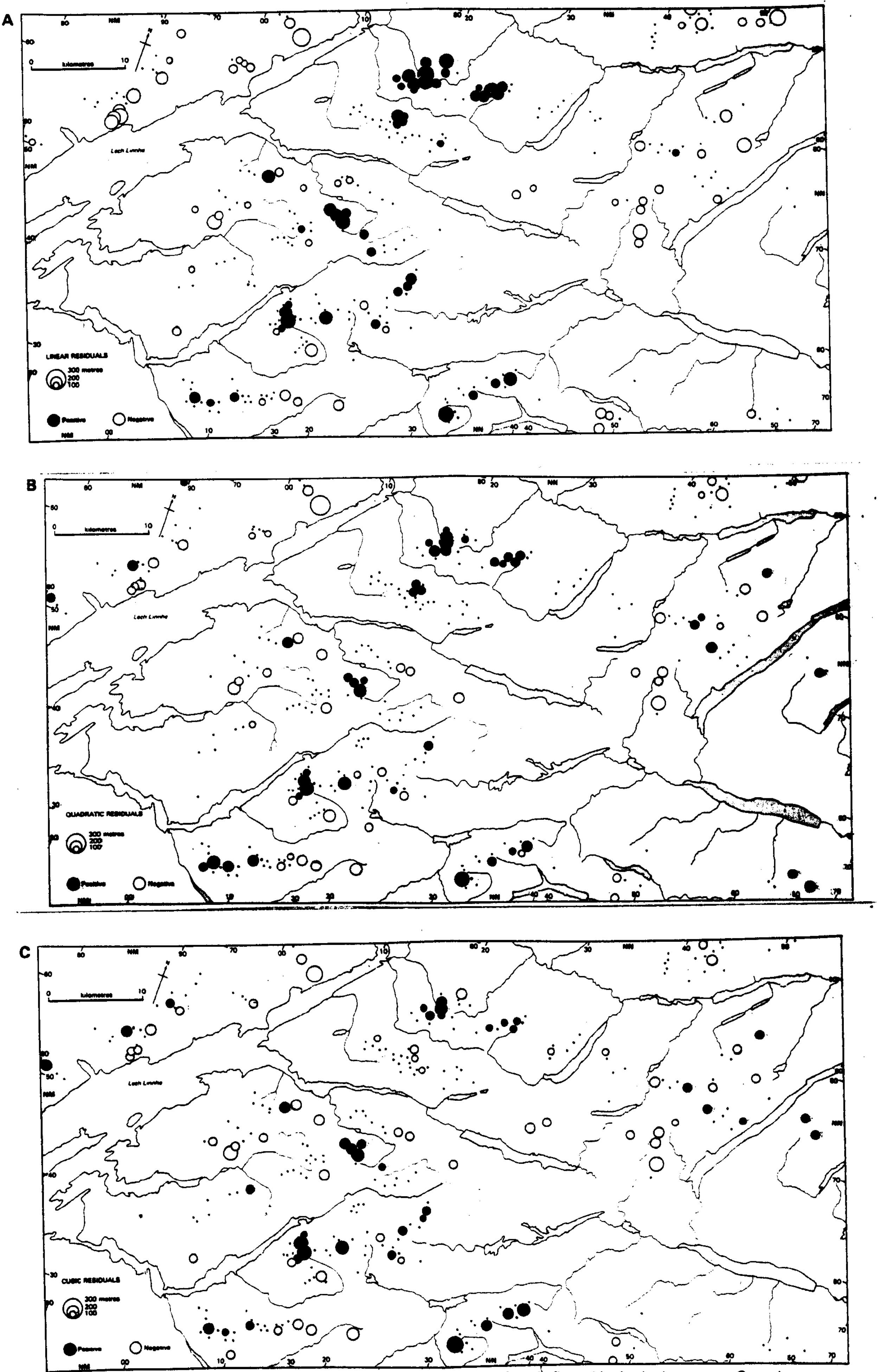


Figure 11.8 Residuals from the linear, quadratic and cubic trend surfaces of corrie floor altitudes in the western Grampians.

A Residuals from the linear surface

B Residuals from the quadratic surface

C Residuals from the cubic surface

The great majority of the positive residuals are clustered around mountain peaks with summits greater than 1,000m O.D. The largest cluster of positive residuals occurs along the Ben Nevis Range with 17 values exceeding 100m. In the Mamore Forest Range four high values are clustered around Sgurr Mhaim(1098m, NN165607) and one below Binnein Mor(1128m, NN212663). In the Bidean nam Bian Range the largest cluster of four such values occurs around Bidean nam Bian(1141m, NN141542) itself. Nine high positive values occur in the Ben Starav Range, but four of these are clustered around the peak of Ben Starav (1078m,NN126426). Elsewhere, most of the remaining large,positive residuals occur in the Ben Cruachan and Orchy-Lyon ranges.

Negative residuals tend to be less clustered than the positive residuals, although two clusters each with four large values occur W of Loch Linnhe below Sgurr na Eanchainne(730m,NM997-659) and Fuar Bheinn(765m,NM853563).

Overall, a clear pattern is apparent with the Western Mountain zone, from the Ben Nevis Range to the Ben Cruachan Range, dominated by numerous large, positive residuals. W and E of this zone the large residuals are almost entirely negative values.

The pattern of residuals over the quadratic surface is similar to that shown for the linear surface except that the values of both the negative and positive residuals have generally been reduced. A number of large positive values are also present E of Rannoch Moor that are not shown in Figure 11.8a.

Residuals from the cubic surface still show minor autocorrelation suggesting that local factors are significant. For example, clustering of large,positive residuals persists below the peaks of Ben Nevis,Bidean nam Bian and Ben Starav. Overall, however,

Area	Linear trend surface		N^o of corries	% R.S.S.	Source
	Gradient m/km	Direction of slope			
Tasmania	4.9	NE to SW	379	50.5	J.A.Petersen and G.Robinson(1969)
Okoa Bay, East Baffin Island	17.0	SW to NE	49	59.3	J.T.Andrews, R.G.Barry and L.Drapier(1970)
Home Bay - Nudlung, East Baffin Island	10.7	SW to NE	42	56.98	ditto
Ekalugad Fiord, East Baffin Island	6.0	NW to SE	36	32.8	ditto
Scottish Highlands and Islands	-	-	175	50.2	G.Robinson, J.Petersen and P.Anderson(1971)
Snowdonia	13.3	NE to SW	84	42.0	D.Unwin(1973)
Southern British Columbia and Alberta	2.8	NE to SW	1244	34.8	A.Trenhsile(1975)
Front Range, Rocky Mountains	4.26	SW to NE	12	109*	T.C.Meierding(1982)
Western Grampians	4.31	ENE to WSW	303	31.9	Author

* Root mean square error

TABLE 11.6 Linear trend surfaces based on the altitudes of corrie floors in different areas of the world.

both the negative and positive residuals show a more random pattern of distribution that suggests that any regional trend within the residuals has been largely eliminated.

It is instructive to compare the results for the western Grampians with the results obtained by other workers for other areas using similar data. The dip and fit of linear surfaces computed from similar data has been tabulated for other areas in Table 11.6. The direction of dip towards the SW and a gradient of 4.31m/km for the linear surface fitted to the western Grampian data are very similar to those computed for Tasmania(Petersen and Robinson,1969) and the Canadian Cordillera(Trenhaile,1975). These gradients are very much less than the gradient of 13.3m/km produced for Snowdonia by Unwin(1973) and the gradients of 6m/km to 17m/km calculated by Andrews et al(1970) for corrie-floor altitudes in the east Baffin Island area. The reasons for the differences are not readily apparent and further similar studies are required from many more areas before possible explanations may emerge.

Robinson et al(1971) chose the cubic surface intuitively as the surface most likely to represent the regional trend of 175 corrie-floor altitudes in the Scottish Highlands and Islands. Comparison between the best-fit cubic surface shown in Figure 11.7c(this thesis) and their cubic surface(Figure 1) for the same area of western Scotland indicates a broad correlation with, in each case, the surface declining in altitude from the NE to the SW. However, their surface is based on very few data for the western Grampians and fails to highlight the many corries at altitudes greater than 800m O.D. in the NE quadrant of the study area.

The rise in the altitude of the corrie floors to the ENE in the western Grampians, as depicted by the linear trend surface(Figure 11.7a) is paralleled by a decrease in precipitation at the present time that is almost identical in direction(cf Figures 11.7a and 10.4a). This suggests that the atmospheric circulation at the time the corries were being formed was similar to that of today, with the main snow-bearing winds blowing most frequently from the SW. The cubic surface fitted to the corrie floor altitudes indicates a more complex situation and emphasises the high corries in the Ben Nevis and Ben Alder ranges and in the mountains to the N of these two ranges. This pattern implies a strong precipitation shadow existing to the N of the Ben Nevis and Ben Alder ranges during the time the corries were excavated. This view is supported by the very low precipitation values of ca 200 - 300mm yr^{-1} calculated for the Spey valley(this area lies just a few kilometres to the NE of the study area) during the Loch Lomond Stadial by Sissons(1979c). The cubic surface also correlates well with the high equilibrium firn lines calculated,for the NE quadrant,on the basis of 226 former Loch Lomond Advance glaciers in the Highlands and Islands(Sissons,1979c). The declining altitude of the cubic surface toward to the E from the dome over the Ben Nevis and Ben Alder ranges may reflect, in part,the increasing importance of snow-bearing winds from the S and SE(Sissons,1979c), but counterbalancing this tendency would have been the increasing dryness of the prevailing south westerly winds as they crossed the Western Mountain zone. However, caution is necessary with this interpretation since the area E of the Ben Alder Range is close to the margin of the control area and also it is not known by how much topography has influenced the form of the cubic surface.

11.8 Conclusions

Corries reach their optimal development in the western part of the study area and steadily decrease in number to the E, even though the area of land above 600m increases in the same direction(Figure 1.2). Theoretically, the number of corries in the E should be greater than in the W given the larger potential areas above 600m from which snow could be blown and yet the converse is true. Thus it must be assumed that such favourable topographic conditions were more than counterbalanced by decreasing amounts of snow toward the central Grampians.

The orientations of 271 corries in the study area demonstrate a strong preference for directions facing between N and E. A higher proportion of N-facing corries than is usual for mid-latitude areas in the northern hemisphere reflects several mountain ranges that are aligned in an E - W direction(Bailey et al,1960; Sissons,1967).

Although the corries show considerable variations in shape and size, caused in part by variations in lithology and structure and in the size of the area from which snow was blown into the corrie, the great majority indicate a unity of form that relates to common glacial processes. However, the angle of slope of the back-wall, in particular, appears to be strongly influenced by the jointing sets in the bedrock with the steepest angles occurring in strongly-jointed quartzite and volcanic bedrock, while some of the most gently-sloping backwalls are frequently associated with the granites.

The mean size of the corries increases toward the E while their orientation tends to be increasingly restricted to directions facing between N and E (albeit a weak trend). If these trends are not artifacts of the way in which the corrie parameters were measured they may indicate that snowfall decreased in a similar direction during the time of formation of the corries. Stronger support for this view is provided by the trend surfaces to a 3rd order fitted to the corrie-floor altitude data. Residuals from these surfaces indicate considerable variations in the altitudes of the corrie floors, owing to local geologic and topographic effects. Nevertheless, there is a distinct regional trend in the data that in the cubic surface accounts for 49.1% of the variance in the corrie-floor altitudes. This reflects the general decrease in snowfall toward the NE quadrant of the study area, both in the past and at the present time.

The use of corrie parameters as surrogates for past climatic trends across the western Grampians can only provide at best generalised information. To obtain more detailed palaeo-climatic information it is necessary to use data based on the form and orientation of reconstructed glaciers: it is this aspect that is discussed in the next chapter.

CHAPTER 12

PALAEOCLIMATIC INFERENCES BASED ON GLACIER RECONSTRUCTIONS

12.1 Introduction

In recent years considerable effort has been devoted to reconstructing past climates based on former glacier outlines,as for example in the western U.S.A.(Porter,1977; Pierce, 1979), the Italian Alps(Porter and Orombelli,1982), New Zealand (Porter,1975) and Baffin Island(Andrews et al,1970). In Scotland much of the work carried out on palaeoclimates based on glacier reconstructions has been by Sissons and his co-workers(e.g.Thompson, 1972; Sissons,1974a,1977b,1979c,1980; Sissons and Sutherland,1976; Robinson,1977; Ballantyne and Wain-Hobson,1980; Cornish,1981).

In this chapter some of the methods first pioneered by Sissons will be applied to the reconstructed glaciers in the western Grampians and attempts will be made to make specific inferences about the climate of the Loch Lomond Stadial.

12.2 Methods

Areas and volumes of the three small,discrete glaciers in the W of the study area(Figure 12.1) were calculated using a grid of squares with sides equivalent to 100m while the areas and volumes of the large transection glaciers were calculated from a grid of squares with sides equivalent to 250m. These grids were superimposed onto maps of scales of 1:25,000 or 1:50,000,depending on glacier size,and on which ice-limits and glacier surface contours had been drawn.

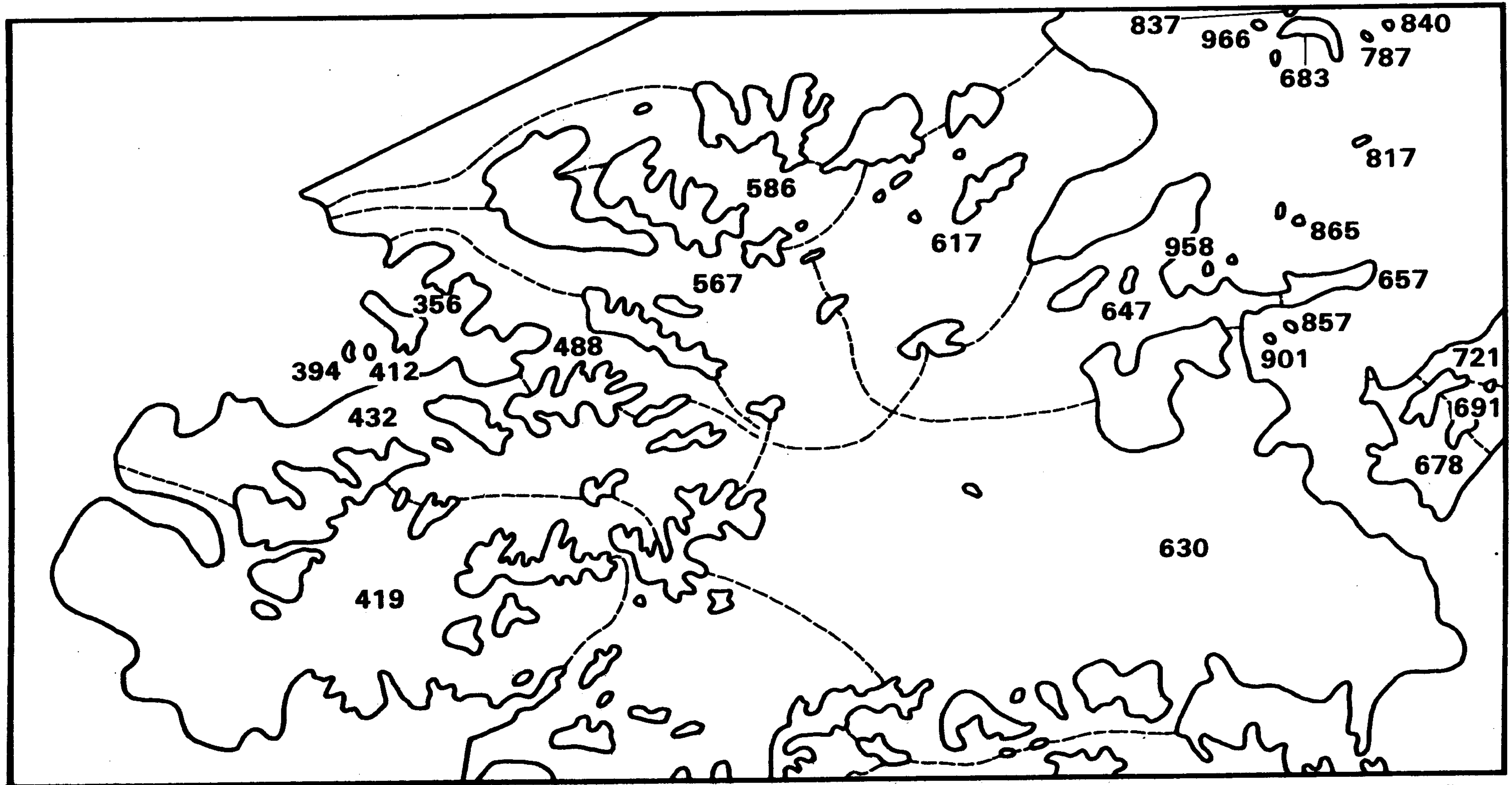


Figure 12.1 Firn line altitudes in the study area. Values for small independent glaciers in the east after Sissons (1979c, 1980 and unpublished)

Glacier	Glacier area km^2	Glacier volume km^3	Maximum thickness of ice m	Glacier weight 10^9 tonnes	Minimum altitude m O.D.	Minimum altitude at snout m O.D.	Equilibrium firn line m	
Etive	297.0	60.6	580	54.54	-75	-20	419	Western side of ice mass
Creran	167.55	29.6	360	26.64	-25	-5	432	
Coe	54.5	11.4	460	10.26	0	-	488	
Leven	177.15	41.8	590	37.62	-50	-	567	
Nevis	83.2	16.0	510	13.5	20	-	586	
Linnhe *+	90.0	27.65	580	24.89	-125	-85	-	
Orchy-Strae +	173.0	41.7	500	39.53	-	-	-	
Sub-totals	1042.4	228.75		206.98				
Treig	158.7	30.3	500	27.36	125	255	617	Eastern side of ice mass
Ossian	90.6	20.3	310	18.27	318	318	647	
Rannoch	578.0	136.1	430	122.49	86	206	630	
Lyon +	68.0	12.3	480	11.07	180	180	-	
Spean * +	90.0	27.65	560	24.89	0	150	-	
Sub-totals	985.3	226.65		204.08				
Totals	2027.7	455.4		411.06				

* The area and volume of the Linnhe-Spean glacier were divided into two equal parts. + Part-glacier only

TABLE 12.1 Data relating to the main ice mass of the Loch Lomond Stadial in the western Grampians.

Equilibrium firn lines were calculated for the former glaciers using the area/altitude distribution of each glacier, after the methods described in detail by Sissons(1974a)

The areas, volumes and firn line values for the glaciers forming the main ice mass in the western Grampians are shown in Table 12.1. The measurements for the three small glaciers in the W are provided in Table 12.2 while the firn line values for the independent glaciers in the NE quadrant provided by Sissons (pers.comm.) are shown on Figure 12.1.

Glacier	Glacier area (km^2)	Glacier volume (km^3)	Maximum thickness (m)	Minimum altitude at snout (m)	Equilibrium firn line (m)	V/A ratio
Duror	7.75	0.7439	215	30	356	0.0960
Dubh	0.972	0.0564	150	185	394	0.0580
Salachan	0.363	0.0068	75	250	412	0.0188
Totals	9.085	0.8071				

TABLE 12.2 Data relating to three independent glaciers west of the main ice mass.

The size of the measured glaciers varied greatly ranging from corrie glaciers less than 0.5km^2 to parts of the main ice mass, such as the Rannoch glacier, that exceeded 500km^2. In total the main ice mass covered an area greater than 2,000km^2 in the study area.

The reconstructed ice mass presented problems when attempts were made to delimit the areas of individual outlet glaciers, since inter-connections occur between the glaciers as a result of divergent and convergent flows of ice. For example, ice in upper Glen Etive diverged westward into Glen Creran via the deep glacial breach of Glen Ure(NN065475). The difficulty was to decide where to draw the dividing line between ice flowing W into Glen Creran and ice flowing to the SW along Glen Etive. The demarcation line was eventually drawn by using the evidence provided by ice-direction indicators(Figure 8.1) that indicated that a large proportion of the ice in upper Glen Etive flowed into Glen Creran. This was caused by the build-up of ice in lower Glen Etive as a result of ice flowing in from Glen Kinglass and from the corries on the Ben Starav and Ben Cruachan ranges. Similar types of evidence were used to demarcate the boundaries between confluent glaciers in other areas, such as at the head of Glen Coe where ice diverged to the SW down Glen Etive, to the W down Glen Coe and to the NW toward Loch Leven.

In the Loch Linnhe area the main problem was to delimit the differing proportions of ice contributed to the Linnhe glacier from the Glen Nevis, Loch Leven and Glen Coe areas and from the mountains to the W of Loch Linnhe. The method adopted was to use the size of the source area of each contributing glacier to provide an approximate control on the widths drawn on the Linnhe glacier. For example, the combined source areas of the Nevis and Coe glaciers, calculated when they were at their maximal extent, approximate to $72km^2$. Since the source area of the Leven glacier, with a value of ca $112km^2$ is greater by a factor of $1\frac{1}{2}$ it is assumed that the contribution of ice from the Loch Leven area to the Linnhe glacier was proportionally similar and demarcation lines were drawn on this basis on the outline of the Linnhe glacier. Inspection of the position of medial moraines on a number of glaciers in southern Spitsbergen confirm that the method adopted is a reasonable one(for example, see the Penckbreen glacier on sheet B11 or Vestre Torellbreen on sheet B12).

The subsequent spatial pattern of the equilibrium firn lines(Figure 12.1) appears to support the method used above for the increase in firn lines northward from Glen Coe(488m) to Glen Nevis(586m) is one that would be anticipated from the overall pattern of firn lines calculated for the whole of the Scottish Highlands (Sissons,1980). Further support for such a pattern is derived from the trend surface analysis of corrie floor altitudes for the area described earlier(e.g. section 11.7). Ideally, to act as a check on the derived firn lines, it would have been useful to obtain firn lines for the areas of the main ice mass lying to the W and NW of Loch Linnhe, but these were unavailable at the time of writing.

Equilibrium firn lines could not be calculated for the Orchy,Lyon and Spean glaciers since either parts of or the whole of their source or snout areas lay outside the study area. The delimitation of the areas of the glaciers for which firn lines were calculated are depicted on Figure 12.1.

12.3 Analysis of former glacier parameters

Best-fit trend surfaces to a 3rd order surface were fitted to the 25 firn line values(Figure 12.2). The X and Y co-ordinates for the data points were obtained by noting the position of the firn line for each glacier on Figure 6.1 that shows

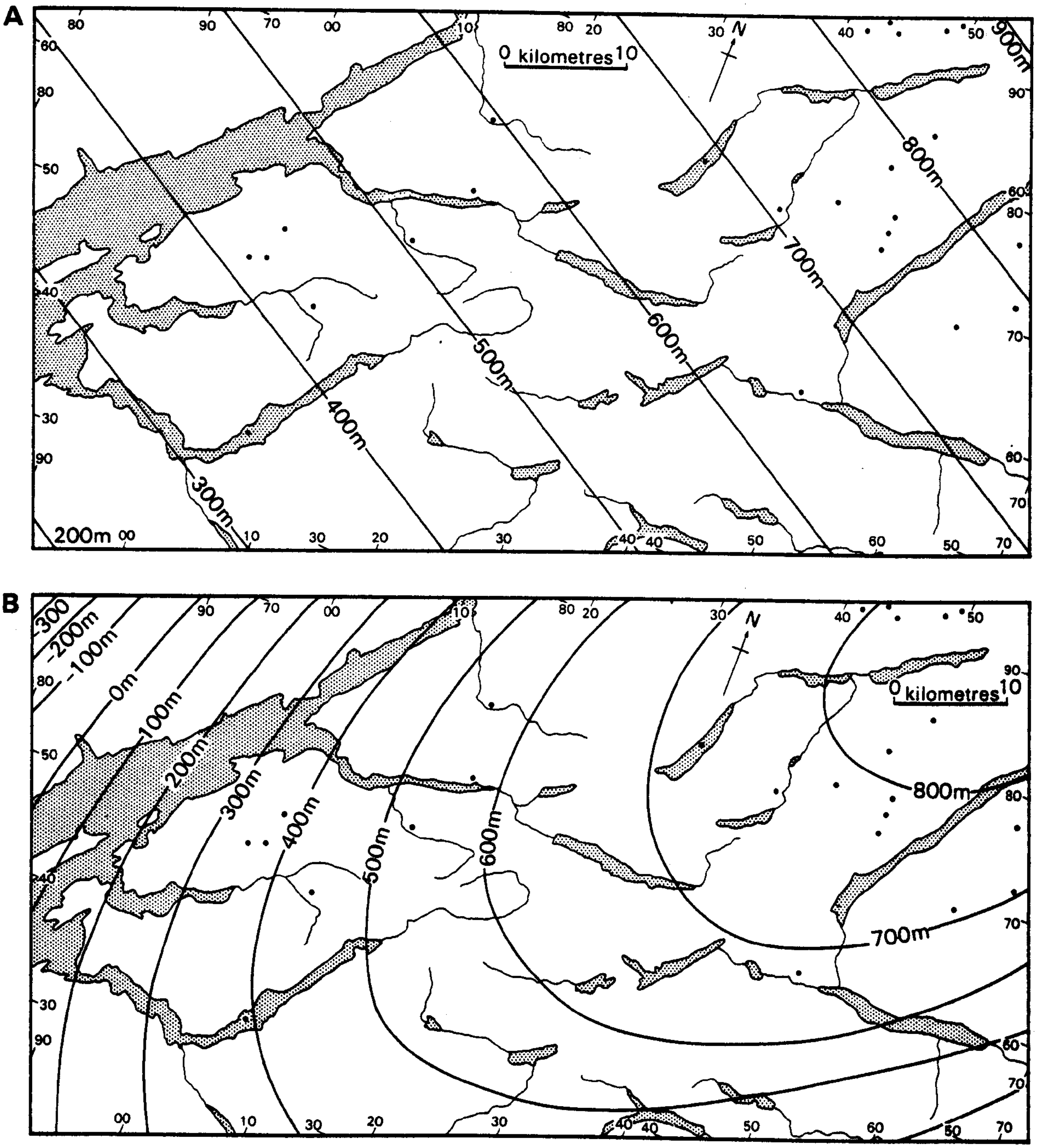


Figure 12.2 Trend surfaces of firn lines in the study area in metres
A Linear trend surface
B Quadratic trend surface

Surface	% RSS obtained
Linear	75.5
Quadratic	80.0
Cubic	83.9

Source of variation	Degrees Freedom	% RSS	Mean Square	F	Significance
Total, 25 data points	24				
Due to linear surface with three constants	2	75.5	37.75	33.887	0.1%
Due to residuals over linear surface	22	24.5	1.114		
Due to added quadratic components	3	4.5	1.5	1.425	ns
Due to residuals over quadratic surface	19	20.0	1.053		
Due to added cubic components	4	3.9	0.975	0.909	ns
Due to residuals over cubic surface	15	16.1	1.073		

TABLE 12.3 Analysis of variance for trend surfaces of equilibrium firn line of the Loch Lomond Advance glaciers in the study area.

the form and surface contours of the reconstructed glaciers. The results are summarised in Table 12.3 that provides % RSS values and a complete analysis of variance for the three surfaces. However, the results are strongly affected by several problems relating to the data. Individual data points are poorly distributed, are relatively few in number and in some areas are strongly clustered. This results in quadratic and cubic trend surfaces with impossibly low values in those areas lacking point-data control, as in the NW quadrant(see the quadratic surface in Figure 12.2). The two higher order surfaces do not provide a significant increase in explanation over the linear trend surface(Chayes,1970). In addition a NE - SW trending ridge is computed in the higher order surfaces that is partially spurious since it slopes steeply to the SE and WNW. This does not accord with the known distribution and general altitude of the ice mass in these two areas(Thompson,1972; Sissons,1979b; author, unpublished mapping). The errors arise because of the lack of firn line data in the north-western and southern parts of the study area.

The linear trend surface with a % RSS value of 75.5(Table 12.3) demonstrates a very high fit to the data and indicates a marked linear trend ranging from 300m O.D. in the SW to 900m O.D. in the NE at a gradient of 7.8m/km. This is the trend surface that best fits the rather limited data and which relates reasonably well with the known form and extent of the main ice mass in the western Grampians. This surface compares well with Sissons' map(1980, Figure 4) of the regional firn lines for the whole of the Scottish Highlands(see Figure 12.3).

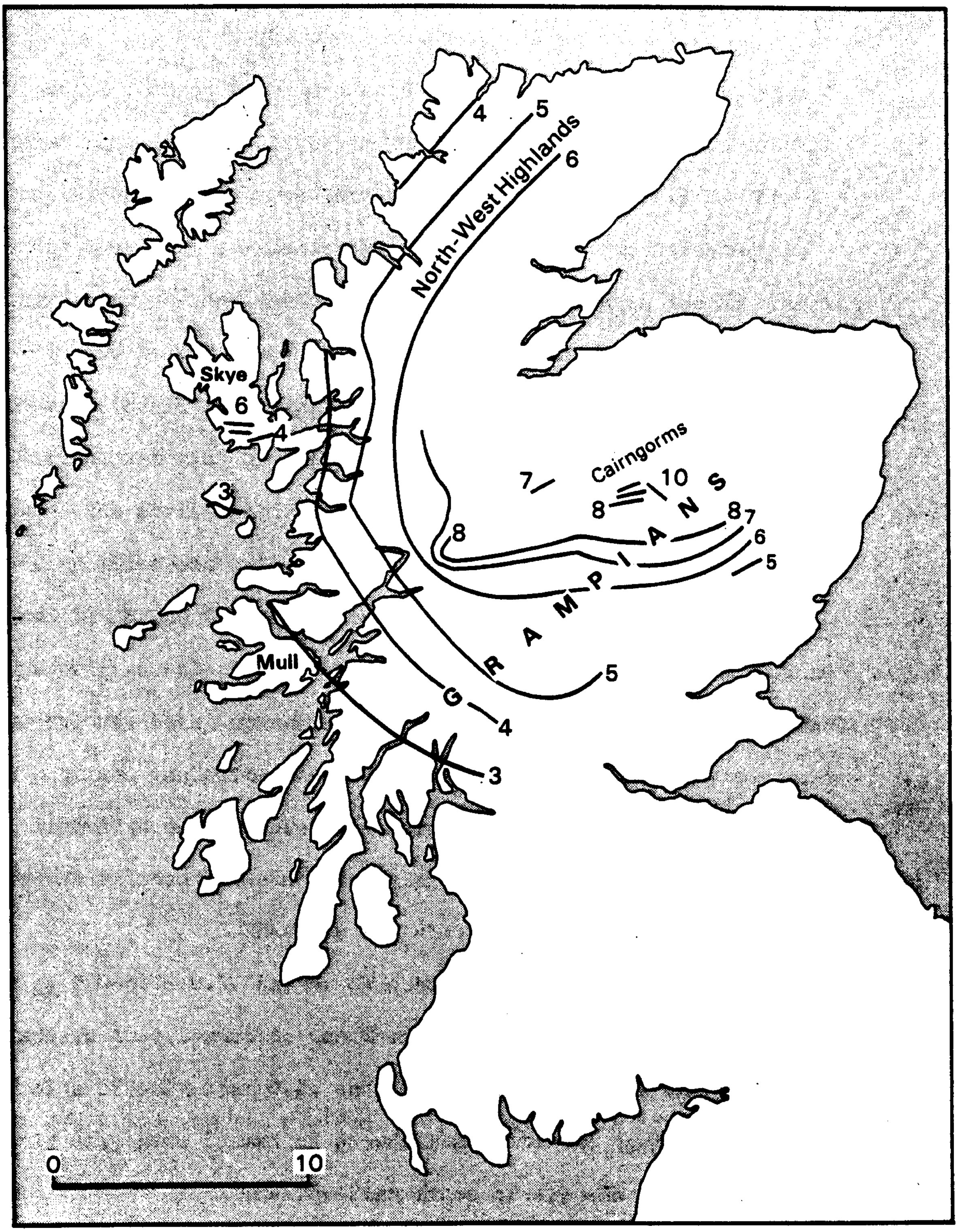


Figure 12.3 Regional firn lines (hundreds of metres) for Loch Lomond Advance glaciers in the Highlands and Inner Hebrides after Sissons (1980).

The increase in firn line values in a NE direction from _ca_ 400m to over 900m over a distance of _ca_ 65km cannot be attributed to major temperature differences across the area, since temperatures were unlikely to have varied to that extent over such a relatively short distance. Furthermore, the discrete,north- and east-facing corrie glaciers in the NE quadrant would have been favoured by lower values of direct insolation,less summer rainfall and lower air temperatures because of their high altitude, yet many of the glaciers in this area failed to extend beyond the corrie floor(Sissons,1979b). Such regional variations in firn lines can only be explained by invoking very much heavier snowfall in the SW than in the NE part of the study area. This type of regional pattern of precipitation during the Loch Lomond Stadial has been emphasised by Sissons(1979c) for a number of areas in Scotland including the north-western Highlands,Skye, the central Grampians and the south-eastern Grampians, although local differences existed between these areas.

That ice was able to develop at altitudes as low as _ca_ 200-300m O.D. in the SW quadrant is indicated by the two small glaciers that formed to the E of low rock walls, below mountain summits of only 654m O.D. and 560m O.D(Figures 5.2 and 5.3). These could only have formed if snowfall amounts had been high in the W.

The low firn lines of 419 and 432m for the Etive and Creran glaciers provide further support for heavy amounts of snow in the W, particularly as large areas of these glaciers faced to the SW where direct insolation would have been at a maximum.

Since the firn lines were calculated for large transection glaciers when they had reached their maximum extent (i.e. the Etive glacier had a surface area of 297km^2 and extended for a distance of 40km from its maximum limit to its main source area in the Ben Starav Range) then the values represent mean values for the many different-sized glaciers that are likely to have built up in the earlier stages of the Loch Lomond Stadial. For example, Sissons(1977b,1977c) has shown that there were marked differences in the firn lines of former glaciers in the NW Highlands and on Skye over very short distances of only a few kilometres. He ascribes these differences to local factors such as the amount of snow blown and avalanched onto the glaciers,glacier aspect and the snow shadow effects of individual mountain blocks. Such marked variations in firn lines for adjacent glaciers are absent for the main ice mass. The firn lines for the main outlet glaciers show a very consistent SW - NE-trending pattern across the main ice mass, ranging from a minimum of 419m for the Etive glacier in the SW to a maximum of 647m for the Ossian glacier in the NE. It is inferred that snow blowing and avalanching from the nunataks is unlikely to have been a very significant factor in creating differences between adjacent glaciers when the glaciers were at their maximum extent(this of course would not have been true in the earlier stages of the stadial when many discrete corrie and valley glaciers would have existed and snow blowing areas would have been a significant factor in their development). Such an inference is based on the relatively small area occupied by the nunataks as a proportion of the total area of the ice mass,i.e. 293km^2 compared with 2,028km^2 (Table 12.1).

In addition the influence of the snow shadow effects of the mountain ranges within the main ice mass is likely to have been reduced as the height of the surface of the ice mass increased, especially in the ice-divide areas of Glen Nevis, Rannoch Moor and Glen Lyon. Therefore, the main factor accounting for the increase in the firn lines in a NE direction across the ice mass must relate to the regional decrease in amounts of snow and this is reflected by the very marked best-fit linear surface with a % RSS value of 75.5 described earlier.

The contrasts in amounts of snowfall across the ice mass are further supported by the asymmetry of the ice mass, with the snouts of the former outlet glaciers on the E side terminating at relatively high altitudes ranging from 150 to 318m O.D (Table 12.1), while the snouts of all the westward-flowing glaciers terminated at or close to sea level.

Calculations of the volumes of the individual transection glaciers show that the total volume of the ice mass in the part of the western Grampians within the study area exceeded $450km^3$ (Table 12.1), with values for individual glaciers ranging from $11.4km^3$ for the Coe glacier to $136.1km^3$ for the Rannoch glacier. An isopach map of ice thicknesses constructed for the main ice mass (Figure 12.5) shows that maximum thicknesses were attained on the W side of the ice mass where the outlet glaciers flowed into the sea lochs. For example, in the Loch Leven, Loch Etive and Loch Linnhe areas ice thicknesses were all close to 600m. In the Rannoch Moor area ice thicknesses averaged between 300 and 400m and the basin contained about 25% of the total volume of ice in the study area.

The surface areas and volumes of the glaciers on either side of the main ice-divide are very similar in aggregate (Table 12.1). Since the glaciers on the western side of the ice mass mainly faced in directions between W and S they would have been subject to high insolation values and high ablation rates. The converse would have been true for the glaciers on the eastern side of the ice-divide as these faced in directions between E and N. This indicates that at the time of their maximum extent the glaciers on either side of the ice-divide had achieved an approximate similar balance with high accumulation rates on the glaciers W of the ice-divide being counterbalanced by high ablation rates. Conversely, E of the ice-divide a similar-sized mass of ice owed its development to the balance between lower accumulation rates and lower ablation rates. This could have interesting implications for estimating the pattern of growth of former Scottish ice-sheets if such a balance between the western and eastern sides of the ice mass could be extrapolated into areas covered by the last ice-sheet (p.327), particularly if field evidence supports a similar balance for the ice mass of Loch Lomond Stadial age in the Western Highlands.

In contrast to the ice mass local factors were much more important in the NE quadrant, where some of the small corrie glacier firn lines (Sissons, pers.comm.) show considerable deviations from the linear trend surface (Figure 12.4). For example, in the NE quadrant several positive and negative residuals lie between values of 100 and 226m, whereas in the remainder of the area only the residuals of the Etive and Treig glaciers exceed 60m (i.e. residuals of 75m and -66m respectively). Such deviations can relate to a number of interacting factors. Older reports assume that firn lines are primarily controlled by the effects of air temperatures on summer ablation (Leopold, 1951; Andrews, 1972; Flint, 1971). More recent work (Sissons and Sutherland, 1976; Evans, 1977; Paterson, 1981; Meierding, 1982) has

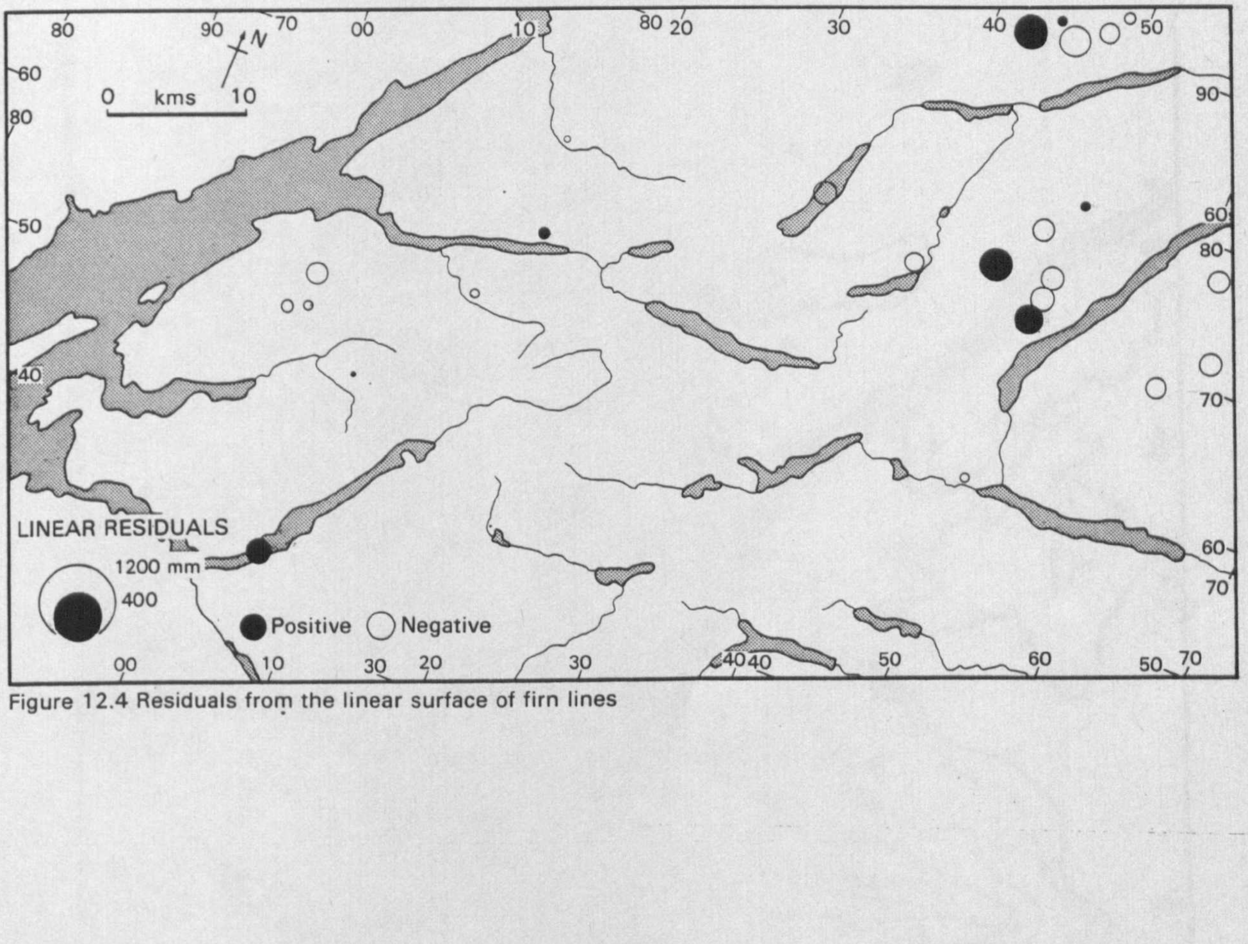


Figure 12.4 Residuals from the linear surface of firn lines

Figure 12.5 Isopach map of the main ice mass in the western Grampians.

demonstrated that firn lines are also influenced or related to other variables. The most important include snowdrift input from avalanching and snow blowing areas, the spatial distribution of cloud, warm advection due to wind, the surface gradient of the glacier and the proportion of direct radiation reflected from its surface, and the proportion of rain to snow.

All of the small discrete glaciers in the NE quadrant faced between N and SE, except for the lower part of the Ardair glacier(Figure 5.4). Consequently amounts of insolation reflected from their surfaces were likely to be low. Many of the glaciers have considerable areas of ground at high altitude to the W or SW from which snow could be blown onto the glaciers(Sissons and Sutherland,1976) and yet the majority of the glaciers were very small ($<1km^2$). This suggests that the glaciers were strongly influenced by direct insolation and implies that clear sky conditions occurred more frequently than in areas farther to the SW. In addition the spatial pattern and altitude of the glaciers suggests that precipitation must have been very low,with glaciers only forming in locations with the most favourable topographic factors.

12.4 An analysis of corries unoccupied by ice during the Loch Lomond Stadial

Graf(1976),in a study of 319 corries containing glaciers at the present time and 240 corries that were unoccupied by ice in the Rocky Mountains of the U.S.A.,reached the conclusion that in a marginal glacial environment the morphology of a corrie was a strong factor in determining the presence or absence of a glacier. He concluded that the optimum glacier location was a large corrie facing NE with its width greater than its length,high steep walls, a pass located to the windward and a mountain peak to the SW. However, one important variable that was not included in the analysis was the importance of the size of the area from which snow could be blown into the corrie(Sissons and Sutherland,1976). Nevertheless, his conclusions provide a useful basis on which to work.

The great majority of the corries in the SW quadrant are believed, from the field evidence, to have contained glaciers at the time of the Loch Lomond Advance. Exceptions include corries that are either small features($<0.02km^3$ in volume) as along the Creran-Etive Range(e.g. c112) or they are shallow corries that lack steep rock walls(e.g. c67 and c77). Thus a major factor controlling the development of corrie glaciers in the SW quadrant appears to have been the shape and size of the corrie. Farther to the NE larger and deeper corries, that from field evidence remained unoccupied by ice during the stadial, occur at altitudes as high as +900m O.D. as for example corrie 35 in the Mamore Forest Range and corries 20 and 22 in the Ben Nevis Range(Sissons,1979b). If the lithological and glacial/periglacial evidence has been interpreted correctly it appears that an additional factor is over-riding that of corrie morphology.

Although the development of a corrie glacier is the result of the complex interaction between a number of variables, the dominant factor preventing their development in large,deep corries must be largely climatic. In order to test this hypothesis corries in the study area were grouped into those that were occupied by ice and those that were unoccupied by ice during the Loch Lomond Stadial, based on clear evidence in the field. Corries with equivocal evidence or those that were identified on map evidence alone were excluded from the tests. For this reason the number of corries involved totalled 145 and not the 271 listed in Appendix B. The corries were further placed into four categories, those facing to the NW,NE,SE or SW, to eliminate differences created by aspect. Five morphological parameters were selected for each corrie(Appendix B), namely width,length,depth,backwall angle and the altitude of the corrie floor to test whether there were any significant statistical differences in morphology between glacierized and non-glacierized corries. Discriminant analysis and minimum distance cluster analysis methods were applied to the data and comparisons made between the two results(a full description of the methods used is given in Appendix C).

The discriminant analysis method indicated that the five corrie parameters could be used to predict whether a corrie was glacierized or non-glacierized with about 75% accuracy. However, as the minimum distance classification method did not indicate any natural groupings it suggests that the previous method imposed an artificial structure on the data. The results, therefore, suggest that important variables, in addition to those of morphology, are likely to have influenced the development of corrie glaciers during the Loch Lomond Stadial.

Thus the regional decrease in snowfall toward the NE would seem to provide a general explanation for the increasing number of corries that failed to nourish glaciers, especially in the NE quadrant, but on a local scale it still fails to explain why on an individual mountain block some corries contained ice while others lacked ice, even where the corries were of a similar size and shape.

For example, a number of well-developed corries that lacked glaciers face directions between N and NW, as along the Mamore Forest Range (e.g. c25 and c35), W of Loch Treig(e.g. c207), along the Ben Nevis Range(e.g.c20 and c22) and N of Loch Laggan(e.g. c189 and c194) yet nearby corries contained glaciers. This implies that local factors, especially the snow shadow effects of individual mountains, were of considerable importance in some areas, and that in particular the N and NW sides of some high mountains received only low amounts of snow. This supports the suggestion that the main snow-bearing winds were blowing from directions between SW and SE during the Loch Lomond Stadial(Sissons,1979c).

12.5 The pattern of precipitation during the Loch Lomond Stadial

A relationship is apparent between the ENE rise in the average altitude of the corrie floors and the decrease in present day precipitation across the study area(Table 12.4.).A similar but crude relationship is apparent between the firn line data and the precipitation data(Table 12.4). These relationships imply that precipitation during the Loch Lomond Stadial decreased in approximately the same direction as at the present time.

Comparison between the linear trend surface of corrie floor altitudes and the linear trend surface fitted to the firn line data(Table 12.4) indicates a similar broad correspondence, although the firn lines rise in a more north-easterly direction than the corrie floor altitudes. However, the glacier firn lines reflect the effects of many inter-related factors(as do the corries), as for example the varying importance of direct insolation to different glaciers(Sissons and Sutherland,1976). Correcting for direct insolation would be necessary, for example, to compare more accurately the linear trend surface of firn lines with the general direction of decrease in precipitation across the area. Furthermore, the poor distribution of firn line data(p.303) in the study area must necessitate a cautious approach when interpretating such results. Nevertheless, the relationships implied between the three variables of corrie floor altitudes, firn lines of the former glaciers and the distribution of present day precipitation suggest that the various climatic parameters that operated in the past did not greatly differ from those that operate at the present time, at least in broad terms. In detail there would have been important differences with, for example, the snow shadow effects of individual mountains,and mountain ranges such as the Ben Nevis Range accentuated under the colder air temperatures of the Loch Lomond Stadial.

Type of data	Direction of dip of surface (degrees E of Grid N)	Gradient	Orientation of surface
Firn lines	210	7.78m/km	NNE to SSW
Corrie floor altitude	250	4.31m/km	ENE to WSW
Precipitation (uncorrected for altitude)	68	12.7mm/km	WSW to ENE
Precipitation (corrected for altitude)	78	20.0mm/km	WSW to ENE

TABLE 12.4 Linear trend surface data based on firn lines, corrie floor altitudes and present day precipitation in the western Grampians.

If the cubic trend surface fitted to the precipitation data(Figure 10.4) is superimposed on the map depicting the main ice mass(Figure 6.1) it will be seen that the zone of maximum precipitation(this is also the zone that contains the maximum number of corries) occurs a few kilometres W of the main ice-divide. This relationship supports the evidence and arguments outlined in section 12.3 for much higher amounts of snow in the western part of the study area during the Loch Lomond Stadial.

Sissons and Sutherland(1976) and Sissons(1979c) have suggested that precipitation gradients(i.e. the decrease or increase in precipitation in both a horzontal and vertical direction) were considerably steeper during the Loch Lomond Stadial than at the present time(e.g. cf Figures 4A and 4B, p.34, Sissons and Sutherland,1976). Such an assumption seems a reasonable one when applied to the glaciers that existed in the study area during the stadial. For example, the large tidewater glaciers in the W advanced along sea lochs for considerable distances during their development(30km and 32km for the Etive and Linnhe glaciers respectively). Losses by calving into tidewater would have been substantial in addition to the adverse factors of direct insolation described in section 12.3. Furthermore, such losses would have been aided by a relative sea level ca 10-12m higher than at present (Gray,1972,1975a). Yet only 30 to 50km to the NE of the maximal limits of the tidewater glaciers large, well-developed corries at altitudes greater than 900m O.D. failed to nourish glaciers(section 12.4). Such contrasts imply that precipitation gradients from the SW to NE were much steeper than at the present time. This view is supported by the steep increase in firn lines in a NNE direction

(Table 12.4). These spatial contrasts in amounts of precipitation would have been accentuated by the build-up of the main ice mass to surface altitudes of +700m O.D in the path of the main snow-bearing winds from the SW. For example, the Laggan valley, lying in the lee of the ice mass and in the shadow of the Ben Nevis and Ben Alder ranges is likely to have received well below its present precipitation amounts of 1000 - 1500mm yr^{-1}; it may well have been as low as 400 - 500mm yr^{-1} similar to the low precipitation suggested for the Spey valley by Sissons(1979c).

12.6 Temperatures during the Loch Lomond Stadial

Present day mean monthly temperatures in the coastal area of Loch Linnhe range from 4.4°C in January to 13.8°C in July giving a mean annual range of 9.4°C(Cruickshank and Jowett,1972). By comparison records kept for the years 1883 to 1903 on the summit of Ben Nevis(Hann,1912) illustrate the major influence that altitude has on temperatures in mountains of the western Grampians. With mean temperatures of -3.8°C and 5.4°C for the coldest and warmest months respectively and a mean annual temperature of only -0.2°C,conditions closely parallel those to be found in subarctic areas.

The existence of a mass of ice exceeding 2,000km^2 in the western Grampians during the Loch Lomond Stadial clearly demonstrates that temperatures during this period must have been considerably lower that at the present time. One way to estimate Pleistocence air temperatures is to apply modern lapse rates to the vertical difference between the equilibrium firn lines of past and present glaciers(Porter,1977). Since no glaciers exist in

Scotland at the present time modern analogues must be sought for in areas of similar latitude and topography. Such an area is S Norway. For example, Sissons(1979c) has used Liestøl's curve, which relates accumulation and summer temperatures(May - Septemer) at the equilibrium firn lines of Norwegian glaciers, to gain some indication of summer temperature during the stadial. If Sissons' view that precipitation amounts in the stadial were roughly similar to those of today but distributed differently(see Sissons,1979c,1980), then average precipitation values of ca 3000 - 4000mm yr^{-1} can be assumed for the mountains in the source areas of the Etive glacier (Figure 6.1). Reducing these figures by 20 - 25% to allow for summer rain(Manley,1959) provides precipitation values in the range 2250 - 3200mm yr^{-1}. On Liestøl's curve this is equivalent to a summer temperature(May - September) in the range 3.0 - 4.0°C at the equilibrium firn line. Using a lapse rate of 0.6°C per 100m and the equilibrium firn line value of ca 400m for the Etive glacier a mean sea level temperature of 5.4 - 6.4°C for May to September is indicated. By inference this suggests a mean July temperature of ca 7.5°C. Since these temperatures relate to a time when the glaciers had responded to a slight climatic amelioration actual temperatures during the stadial would have been lower by perhaps 0.5 - 1.0°C (Sissons,1979c). Using an equilibrium firn line value of ca 600m and precipitation values of 2500 - 3000mm yr^{-1} for the Ben Nevis area comparable mean May - September and mean July temperatures would have been -1.5°C and -0.5°C respectively for the summit of Ben Nevis.

The difference of ca 7.0°C between the present mean July temperature of 13.8°C at mean sea level and the calculated value of ca 7.0°C for the stadial gives some idea of the magnitude of the depression in summer temperatures. Differences in the mean January temperature between those of the stadial and those of today are likely to have been even greater. For example, Sissons(1980) used the existence of permafrost, implied by the presence of fossil frost wedges in western Scotland, to suggest a mean annual January temperature no higher than -9.0°C in part of the western Grampians. Thus if this view is correct it suggests that mean January temperatures were depressed by at least 13.0°C in the Loch Linnhe area during the Loch Lomond Stadial.

CHAPTER 13

COMPARISONS BETWEEN PATTERNS OF GROWTH OF THE LOCH LOMOND ADVANCE GLACIERS AND THE DEVENSIAN ICE-SHEET : GLACIOLOGICAL AND PALAEOCLIMATIC IMPLICATIONS

13.1 Introduction

This chapter has three primary aims. Firstly, to present the field evidence for the existence of former ice-sheet activity, presumed to be Devensian in age, in the study area. Secondly, to reconstruct the former ice-flow directions and ice-divides of the ice-sheet using striae,friction cracks,ice-moulded bedforms and the distributions of erratics, Thirdly, to assess the relationships between the patterns of growth of the Loch Lomond Advance ice mass and the Devensian ice-sheet and the palaeoclimatic implications of such relationships.

Although a large ice mass existed in the western Grampians during the Loch Lomond Stadial sufficient evidence of earlier ice-sheet activity survives on some of the nunataks and beyond the Loch Lomond Advance ice-limits to enable reconstructions of the former ice-flow patterns to be made.

13.2 The evidence for ice-sheet activity

Since the ice-limits of the Loch Lomond Advance have been mapped in the study area(chapters 2 to 5) it can be assumed that any glacial features existing outside those limits will relate to earlier glacial activity, especially the Devensian ice-sheet.

The mapping of erratics in many parts of the area

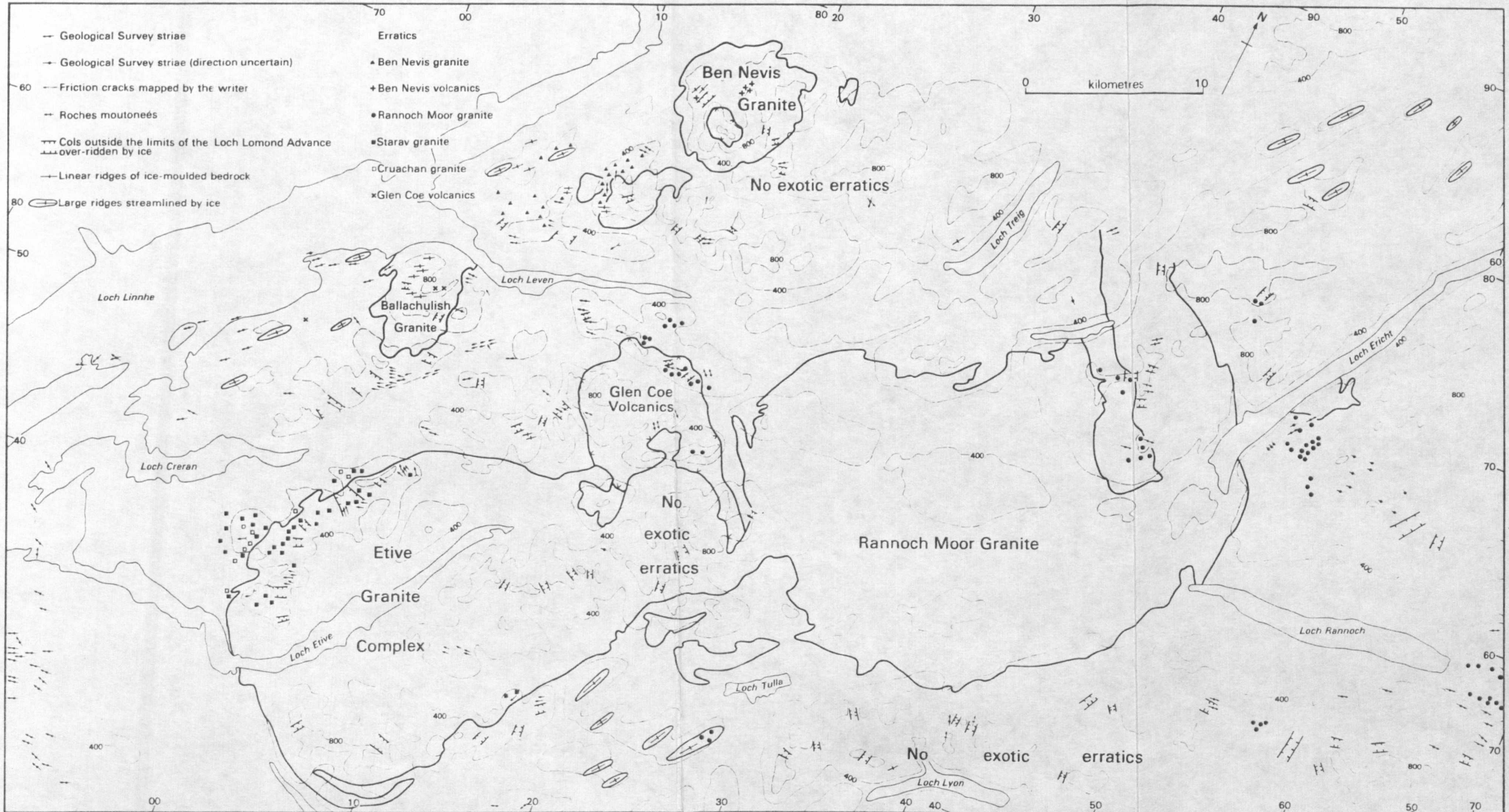


Figure 13.1 Striae, friction cracks, ice moulded landforms, erratics and cols over-riden by ice related to the ice-sheet in the western Grampians. Glacial evidence within the limits of the Loch Lomond Advance not shown.

broadly confirms the observations of earlier workers(Hinxman et al,1922; Bailey et al,1960; Peacock,1970a; Sissons,1976) that the study area formed the main accumulation area of the ice-sheet centred over the western Highlands. Most of the important ice-direction indicator erratics that have been mapped are shown in Figure 13.1.

In the SW numerous granite erratics from the Etive intrusive complex exist outside the outcrop area, with large boulders of the distinctive Starav porphyritic granite resting on Meall Odhar and Cruachan granite surfaces and with all three varieties of granite spread across areas of metamorphic or volcanic bedrock. Starav granite erratics are especially abundant on high ground along the Creran-Etive Range and many hundreds of such boulders litter the frost-riven rock outcrops to altitudes of +800m O.D. as along the summit ridge of Creach Bheinn(NN025422). Many broad, ice-shaped interfluves and numerous cols at altitudes between ca 500 and 700m O.D. that breach the range also testify to former powerful flows of ice across the range in a W or SW direction toward Loch Linnhe.

Farther N the spread of volcanic erratics from the Coe igneous complex and of granite erratics from the Rannoch Moor intrusion indicates that the Leven trough was a major outlet for a large flow of ice from the main ice-divide to the E, although the pattern is complicated in places by glaciers of the Loch Lomond Advance that flowed in directions that differed from the ice-flow directions of the ice-sheet(p. 40). This flow is supported by the evidence on the flat-topped granite summits of the mountains SW of South Ballachulish(NN035582 to NN028558). The bedrock, although it is frost-riven with deep,open joints in many places, has been strongly shaped by the ice-sheet in a W or SW direction at altitudes between 600 and 800m O.D.

The distribution of erratics from the Ben Nevis and associated igneous intrusions, outside the limits of the Loch Lomond Advance, shows that ice from Glen Nevis was deflected to the SW across the mountain summits at the western end of the Mamore Forest Range and across the mountains at lower altitudes near Loch Linnhe. This deflection was probably a result of ice from W of the Great Glen occupying much of the upper part of the Loch Linnhe area. The same mass of ice from the W was probably responsible for the deflection of ice from the Ben Nevis massif,north-eastward toward Glen Spean(indicated by andesite erratics on the high ground to the NE of Ben Nevis, see p.101).

In the eastern part of the study area the mapped distribution of Rannoch Moor granite erratics, outside the ice-limits, confirms earlier work(summarised in Sissons,1976; p. 73-74) of ice-flow taking place toward the E and NE from Rannoch Moor.

A flow of ice toward the SW from Rannoch Moor down glens Strae and Orchy in ice-sheet times is more difficult to justify from the evidence of erratics, mainly because of the many varieties of granite that crop out in the area. Nevertheless, a small number of what appear to be Rannoch Moor granite erratics were mapped along the summit ridge of Ben Inverveigh(NN276388) at altitudes of <u>ca</u> 600 - 620m O.D. suggesting that such an ice-flow direction did take place. In addition the many hills and low mountains in the Glen Orchy area streamlined by ice toward the SW would

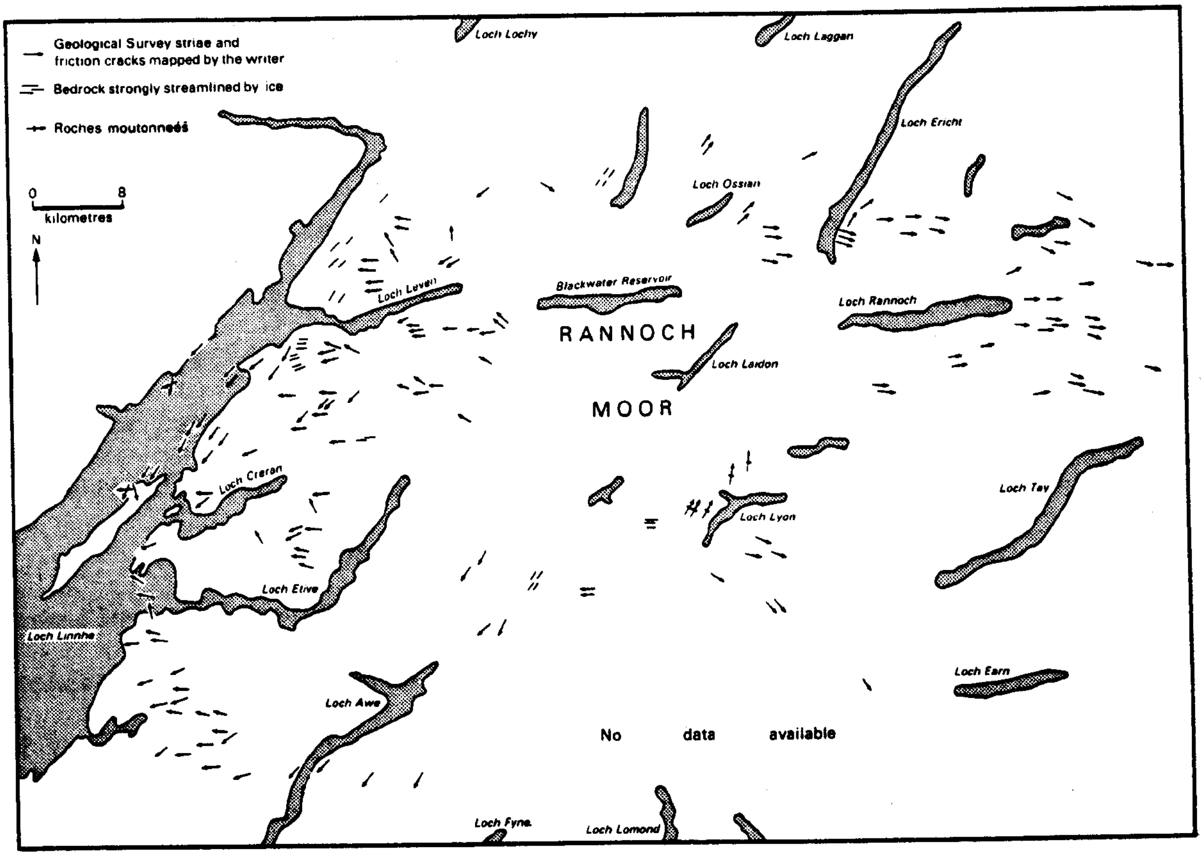


Figure 13.2 Ice direction indicators relating to the ice-sheet in the western Grampians

appear to give added support to such an ice-flow direction.

The absence of exotic erratics in the Glen Nevis and Glen Lyon areas, and in the Ben Starav Range suggests that these areas were not overwhelmed by external ice. This implies that these high mountain areas, as during the Loch Lomond Stadial, maintained radial or subradial outflows of ice and formed ice-domes within the ice-sheet. In the Glen Lyon area supporting evidence exists in the form of roches moutonnées at high altitudes ($>$750m O.D.) and in the orientation of high-level breaches; both indicate that ice-flow took place toward Rannoch Moor in ice-sheet times.

Figure 13.2 depicts all known striae and friction cracks that have been mapped outside the Loch Lomond Advance limits. The area has been extended to include areas outside the study area in order to place the evidence in a wider setting.

The resultant pattern strongly correlates with the ice-flow directions inferred from the evidence of erratics. The dominance of the Rannoch Moor area as the centre of the ice-sheet would appear to be confirmed(Linton,1957; Sissons,1976; Price et al, 1977). However, it is suggested that modifications to this simple concept of radial outflow of ice from Rannoch Moor need to be made. Evidence cited previously in this chapter suggests that areas such as Glen Nevis(the area N of the Blackwater valley is taken to represent an extension of the Moor of Rannoch), Glen Lyon and the Starav Range maintained ice-flow into(and across) the basin even at ice-sheet maximum. Thus the position of the main ice-divide of the ice-sheet is likely to have been very similar to that depicted for the main ice mass of the Loch Lomond Advance(Figure 6.1). There is no evidence to suggest that the ice-divide migrated eastward as

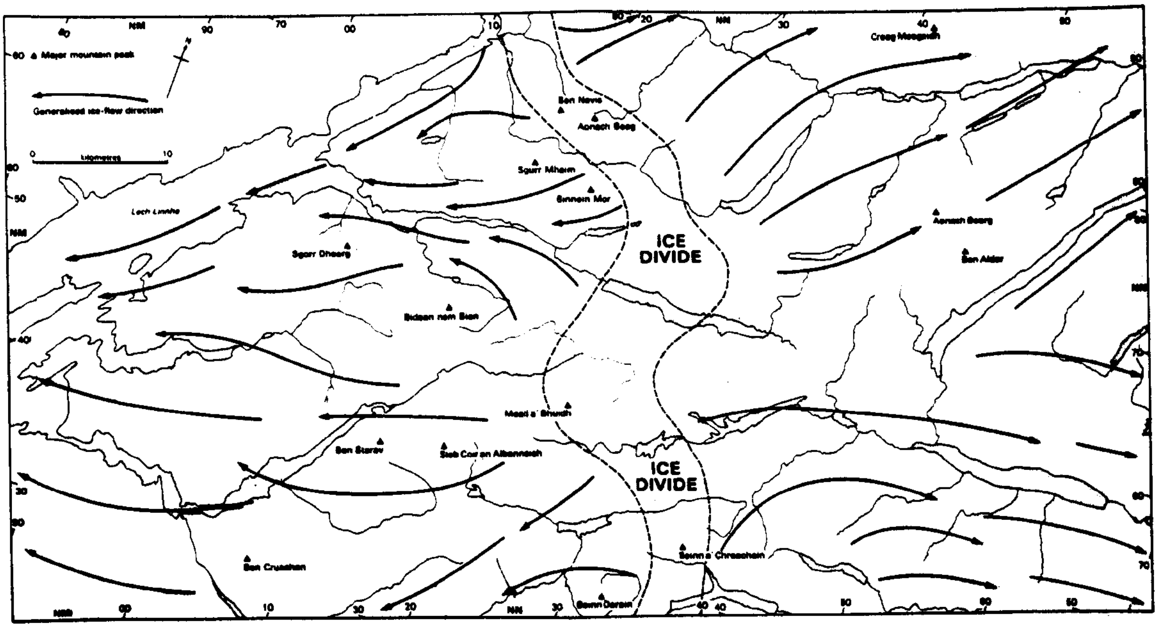


Figure 13.3 Generalised ice flow directions and the location of the ice-divide of the Late Devensian ice-sheet in the western Grampians.

hypothesised for the Scandinavian ice-sheet (Flint,1971). In addition recent geophysical soundings of the thickness of the East Antarctican ice-sheet reveal that the summits of ice domes can correlate with topographic rises as well as with topographic lows, and confirms earlier studies of the Greenland and former Laurentide ice-sheets(Sugden and John,1976).

It is suggested that once established the ice-domes over the western mountain blocks would have maintained their higher ice levels, particularly in view of possible summit altitudes > _ca_ 1500m O.D.(Sissons,1976; Boulton _et al_,1977) that would have emphasised the precipitation contrasts between the western and eastern sides of the ice-sheet. Snowfall from S or SW air-streams crossing the ice-domes would have decreased markedly on the E side once over the summit. Thus it is difficult to envisage an ice-shed farther E than the W side of Rannoch Moor(Figure 13.3).

13.3 Comparisons between the pattern of build-up of the Loch Lomond Advance glaciers and the Devensian ice-sheet

A strong similarity between the ice-flow movements and growth patterns of the Loch Lomond Advance ice mass and the Devensian ice-sheet is supported by several lines of evidence. The carry of regional erratics such as the Ben Nevis,Etive and Rannoch Moor granites away from the main glacier source areas demonstrates that the main directions of ice-flow were basically the same(_cf_ Figures 6.1 and 13.3). Similarly the spatial patterns of striae and friction cracks for the Loch Lomond Advance glaciers and the Devensian ice-sheet(Figures 8.1 and 13.2) demonstrate a _broad_ correlation in terms of ice-flow directions, if ice-flow movements within corries are excluded(see chapter 8). In addition in a number of areas, such as at the western end of the Mamore Forest Range and in the Glen Orchy area, the ice-flow directions , indicated by roches moutonnées and ridges streamlined by ice outside the limits of the Loch Lomond Advance,show strong correlations with directions indicated by similar features within the limits.

Such relationships,described above, imply that the climatic parameters of the Loch Lomond Stadial and the Devensian were broadly similar in both periods. This would seem to suggest that the ice mass depicted in Figure 6.1 represents an early stage in the build-up of an ice-sheet that was terminated by climatic amelioration(cf Sissons,1979b, p.41). In many areas the patterns of erosion on a macroscale become explicable if this conclusion is accepted, as will now be exemplified.

a) Corries

The large number of corries in the Western Mountain zone indicates that this zone in the study area, as during the Loch Lomomd Stadial, nourished numerous corrie glaciers during previous cold periods. It is inferred that this zone formed the primary source area for the early development of a large ice mass in western Scotland, during each glacial period. Away from this zone many of the corries show increasing modification by ice-sheet erosion, except where local ice may have been dominant in the early stages as on the highest mountains such as the Ben Alder and Creag Meagaidh massifs. In some areas several corries have undergone intense erosion by the ice-sheet in the form of foreshortened side walls that have been smoothed and rounded by ice-sheet erosion, as in the area W of Glen Creran and along the Creran-Etive range. An explanation for such erosion lies in their nearness to major source areas of ice and early over-riding by external ice.

b) V-shaped valleys

A number of tributary valleys facing directions between S and SW occur on the sides of mountains in the study area. These lack a typical glaciated shape and instead they often display a deep, V-shaped cross-profie(e.g. on the S side of the Mamore Forest Range at NN143653 and on the N side of Glen Etive at NN160523). Since these valleys failed to nourish glaciers(thick accumulations of frost-riven debris mantle their slopes) during the Loch Lomond Advance and as it has been demonstrated(p.178) that ice advancing up a reverse slope will tend to deposits large quantities of till, their V-shaped profile becomes explicable if similar conditions are extrapolated for previous glacial periods.

c) Breaches and cols

Numerous low-level and high-level breaches and ice-smoothed cols in the area testify to the intense erosion that has taken place and illustrate how ice utilised different breaches at different stages in the build-up of an ice-sheet. The orientation of many of the breaches often approximates to the ice-flow directions indicated by striae and friction cracks in the immediate area. Such examples related to ice-flow in ice-sheet times occur on the W side of Glen Creran at NN014485 and NN029509 and along the Creran-Etive Range at NN045440 and NN020405. However, such a relationship should not be regarded as consistent since some glacial breaches at low-levels show a relationship with ice that was strongly controlled by topography as shown by many of the glaciers of the Loch Lomond Advance.

In some areas the alignments of the breaches or

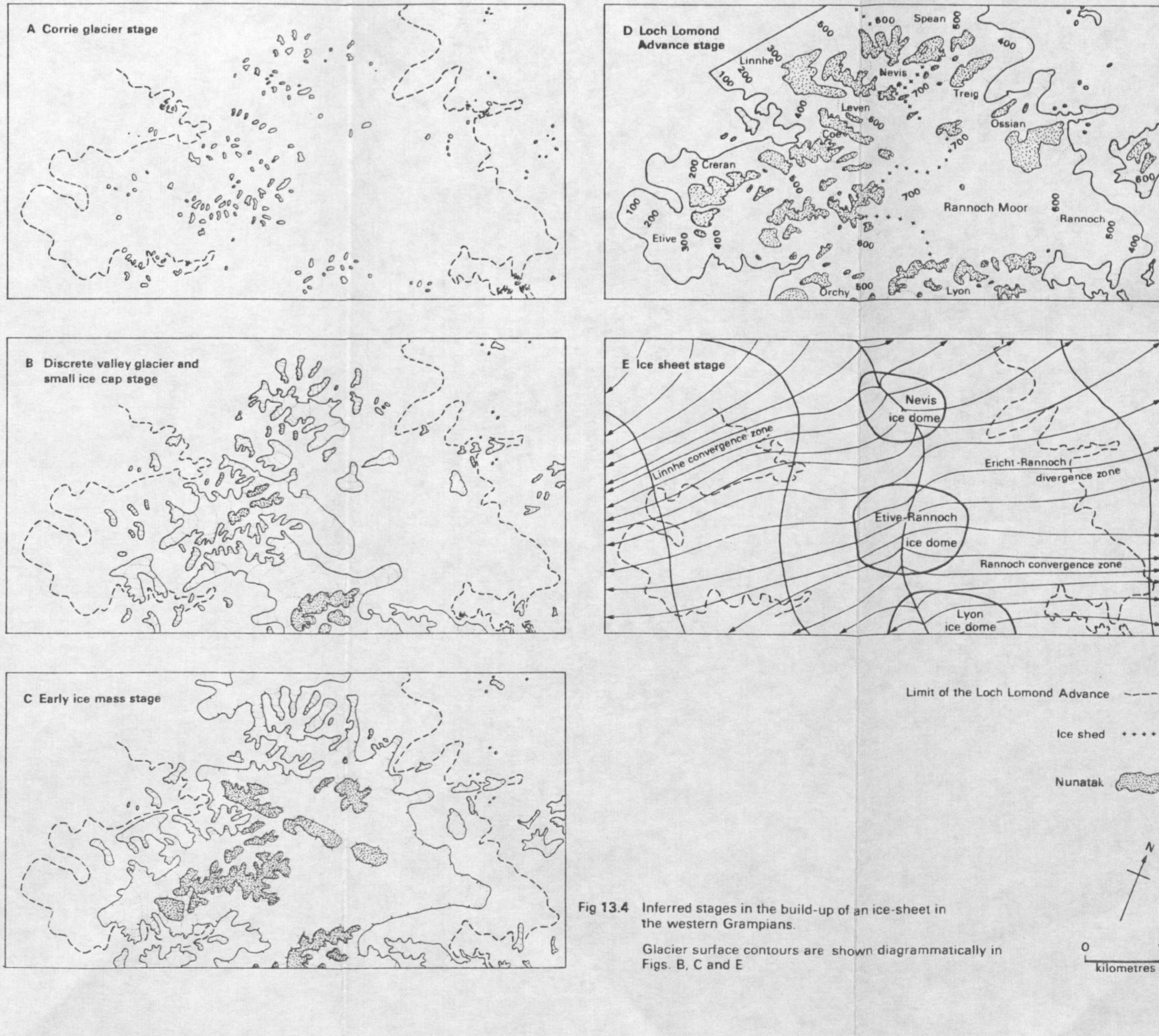


Fig 13.4 Inferred stages in the build-up of an ice-sheet in the western Grampians.

Glacier surface contours are shown diagrammatically in Figs. B, C and E cols are puzzling and difficult to relate to the local topography until they are related to ice-sheet conditions. One such example is the col(NN144728) NW of Ben Nevis and occupied by Lochan Meall an t - Suidhe. Its N - S orientation suggests that it was unlikely to have formed from local ice nourished in the corries on the N side of Ben Nevis. Although its orientation might suggest an origin related to ice flowing N from Glen Nevis during conditions of limited ice cover this can be discounted on the basis of evidence cited previously (p.103). Its orientation becomes explicable if it is seen as a result of ice from W of the Great Glen, under ice-sheet conditions, diverging S along the flanks of Ben Nevis to combine with ice flowing south-westward from the Glen Nevis area.

In view of the similarities of the main ice mass of Loch Lomond Stadial age with the last ice-sheet it is reasonable to assume that the development of each took place in similar stages. Such a possible sequence of stages is depicted in Figure 13.4.

A number of assumptions have been made in the reconstructions. These are :

i) that complete deglaciation of the preceding ice-sheet had taken place;

ii) that the climatic parameters were broadly similar to those that operated during the Loch Lomond Stadial (<u>i.e.</u> snowfall decreased in a SW to NE direction across the area);

iii) that only those corries known to be occupied by ice during the Loch Lomond Stadial contained glaciers; this may not have applied to the early stages of the last ice-sheet since the corries unoccupied by ice during the Loch Lomond Stadial must have nourished ice at some time in the past if indeed they have been correctly identifas erosional forms produced by corrie glaciation ; and

iv) that the build-up of ice was most rapid in areas of high mountains containing many corries and where snowfall is inferred to have been heaviest.

The following sequence of stages is inferred on the basis of these assumptions and the data presented in this thesis.

A) In Stage A the Western Mountain zone is inferred to have dominated as the primary source of ice. Small valley glaciers are suggested where several corrie glaciers were likely to have coalesced in the early stages. Glaciers in the E are few in number and largely restricted to the higher mountain blocks and to the plateau area in the vicinity of Loch Garry. The large area of Rannoch Moor and the Rannoch valley devoid of ice is clearly demonstrated. It is assumed that the basin of the Moor of Rannoch did not nourish independent glaciers.

B) By Stage B many corrie glaciers in the W are assumed to have coalesced to form large valley glaciers that are advancing down the main outlet valleys in the W. Tongues of ice are shown as advancing across the Moor from corries on the Ben Starav and Orchy-Lyon ranges; these ice lobes would have been favoured in their growth by their aspect(facing directions between E and N) and by the higher altitude of the Moor in comparison with the outlet glaciers to the W descending steeper gradients down to sea level.

C) In Stage C glaciers from the main sources in the Western Mountain zone are shown as having coalesced while Rannoch Moor has filled with ice and a large glacier is flowing E down the Rannoch valley. Many low-level breaches are shown as being utilised by over-spilling ice, including the major breach between glens Creran and Etive.

D) Stage D has been discussed at length in this thesis and requires no further elaboration.

E) The ice-sheet stage is based on the evidence presented in this chapter and depicts the main ice-shed as aligned in a NW - SE direction through the three inferred ice-domes of Nevis,Lyon and Etive-Rannoch(This may explain why no exotic erratics have yet been found on the upper slopes of Ben Nevis). The main flow of ice from Rannoch Moor was to the E and NE down the Rannoch and Spey valleys respectively and around the flanks of the Gaick Plateau and the Cairngorms farther E (Sissons,1976). The only major flows of ice from Rannoch Moor toward the W were down the Coe-Leven outlet and probably down the glens of Strae and Orchy. A very strong convergence of ice on the great structurally-aligned hollow of Loch Linnhe is suggested by the evidence of striae,friction cracks,ice-moulded bedforms and the distribution of erratics(Bailey et al,1960; mapping by the writer).

13.4 Conclusions

The field evidence described in this thesis supports Cornish's findings in the western Southern Uplands(1982) that a strong similarity existed between the former source areas of the ice-sheet and those of the Loch Lomond Advance glaciers and that, unlike the Scandinavian ice-sheet(Flint,1971), the main ice-divide did not migrate eastward during the build-up of the Late Devensian ice-sheet to its maximum extent. The main ice-divide extended in a NNW - SSE direction across the western Grampians,from the Ben Nevis Range in the N to the head of Glen Lyon in the S.

Denton and Hughes(1981) have stressed that ice-divides are the least dynamic components of an ice-sheet and that the ice-divide will consist of alternating ice-domes and ice-saddles. In addition recent reconstructions of the Laurentide ice-sheet(Shilts, 1981; Andrews,1982) particularly emphasise its multi-domed morphology.

In the western Grampians the Late Devensian ice-sheet is believed to have consisted of six alternating ice-domes and ice-saddles. The three saddles were located in the vicinities of the head of Loch Linnhe, the western part of the Blackwater valley and the basin occupied by Loch Tulla. Strong convergent flows of ice with ice-streaming took place to the W or SW of the three saddles. Such powerful ice-streams are supported by the hills,ridges and rock knobs streamlined by ice in a W to SW direction(Figure 13.1) to the W of the three saddles, although it is likely that much of the erosion took place prior to the build-up of the ice-sheet to its maximum extent and during earlier glaciations. In contrast the three ice-domes centred over Glen Nevis, western Rannoch Moor and upper Glen Lyon are inferred to have been characterised by far less powerful diverging flows of ice (Denton and Hughes,1981).

The position of the main ice-divide also accords with the ellipsoid pattern of maximum glacio-isostatic uplift aligned in an approximate N - S direction over the western Grampians that is to be expected on the basis of deformation of the Main Lateglacial Shoreline(Gray and Lowe,1977; Gray, 1978). However, recent shoreline evidence suggests that the centre of uplift shifted ca 30km to the E in the early Flandrian(Gray, oral comm.). Whether this relates to different loading of the crust during the Loch Lomond Advance, by for example, ca 411×10^9 tonnes of ice in the western Grampians (Table 12.1), compared with the loading of the crust in the Late Devensian is uncertain at the present time.

The vertical and horizontal dimensions of the Late Devensian ice-sheet at its maximum extent are far less well known(Sissons,1981b) and no new evidence to elucidate such dimensions was found during the present study.

CHAPTER 14

CONCLUSIONS AND THEMES FOR FUTURE RESEARCH

14.1 Conclusions

The methodology of reconstructing former glaciers based on the field mapping of a wide range of detailed morphological evidence(summarised in Sissons,1979d) is strongly supported by the results of this study. In addition the mapping of 'erosional' trimlines, particularly in the accumulation areas of the former glaciers, where other forms of ice-marginal evidence are absent or scanty, has been shown to be a valid field mapping technique.

Although trimlines have been referred to in a number of studies involving glacier reconstructions none of the workers mapped trimlines systematically over a wide area. Furthermore, many such studies have only identified 'erosional' trimlines on the basis of aerial photographs (Porter,1975; Porter and Orombelli, 1982; Meierding,1982). Attempts by the writer to apply this technique to the study area led to very large errors of $\pm$ 200m when checked by field mapping and attempts to identify 'erosional' trimlines on aerial photographs were abandoned at an early stage in the study.

The technique of mapping trimlines in the field in the western Grampians does, however, have a number of important limitations. These are:

i) A trimline cannot provide an exact limit in numerical terms.

ii) The method produces the most consistent results when applied to large valley glaciers or ice masses of the type that existed in the western Grampians during the Loch Lomond Stadial, especially if there were many nunataks within the ice mass. The technique produces poor results when applied to small discrete glaciers.

iii) As large a number of trimlines as possible needsto be identified to avoid erroneous reconstructions.

iv) Variations in lithology introduce complications into the interpretation of the evidence, above and below an inferred trimline, since different rock types vary greatly in their response to glacial and periglacial processes. For example, the high resistance to frost-riving of some massive rock types, such as the Starav granite, reduces the clarity of the inferred trimline and errors of interpretation are more likely to occur in the field.

v) Trimlines become less distinct below altitudes of <u>ca</u> 400 - 500m O.D. in the study area, mainly because the periglacial processes operating during the Loch Lomond Stadial were less effective at low altitudes and therefore serious errors are more likely, where the glaciers descended to low altitudes.

The usefulness of recording striae and friction cracks over a wide area is to be stressed since these reflect most accurately the former direction of ice-flow. Crescentic fractures, in particular,are the most valuable form of friction crack in the study area since they are usually the most abundant, they are least affected by minor rock structures and they occur on a wide variety of rock types, especially those rich in quartz that are resistant to weathering processes. Unfortunately, not all of the areas in the western Grampians contain abundant striae and friction cracks and this is particularly true for the eastern half of the study area where psammitic-type and granitic-type rocks predominate. Above altitudes of ca 500 - 600m O.D. the contrasts in the number of glacial markings,inside and outside the limits of the Loch Lomond Advance,are generally very pronounced with large numbers of glacial markings on ice-smoothed bedrock inside the limit contrasting with only isolated glacial markings outside the limit(such a comparison can only be made for the same rock type inside and outside the limit). At altitudes lower than ca 400 - 500m O.D. the contrasts are far less marked and considerable numbers of glacial markings related to the ice-sheet outside the limit can sometimes be preserved on suitable bedrock. Such contrasts clearly relate to the varying severity of the former periglacial processes with altitude during the Loch Lomond Stadial.

The results of this study confirms the view of Sissons(1976) that the study area was the major source area for both the ice mass of Loch Lomond Stadial age and for the Devensian ice-sheet. The zone of high mountains extending from Ben Cruachan northwards to Ben Nevis was of paramount importance in providing, initially, numerous corrie glaciers that coalesced to form large glaciers that flowed westwards toward Loch Linnhe and eastwards into Rannoch Moor. The development of a large ice-cap on Rannoch Moor was aided, partly by the lower temperatures(estimated at ca 3.5°C for May to September) due to the altitude of the Moor(ca 300 - 400m O.D.) and partly because its main source areas in the Ben Starav and Orchy-Lyon ranges lay to the W and S: glaciers derived from these source areas would have had aspects toward the E or N and values of direct insolation would have been low during their development.

The field evidence for reconstructing the upper limit of the Rannoch Moor ice-cap indicates that the maximum height of 700 - 750m O.D for the surface of the ice-cap was less than the 850 - 900m O.D. predicted by Thompson(1972) and Sissons (1980).

The maximum limits of the Creran,Etive,Treig, Ossian and Rannoch glaciers as proposed by Thompson(1972), Gray (1975a) and Sissons(1979b) are supported by the field evidence descrbed in this thesis. The belief, however, that the Leven and Linnhe glaciers remained as independent glaciers and terminated at the North Ballachulish outwash spread(McCann,1966) and in the vicinity of the Corran Ferry-Onich area(Peacock,1971a),respectiv-ely,is not upheld. The evidence provided by the four outwash fans near Kentallen(p.105) and the trimline and other ice-marginal evid-ence in the Loch Leven area demonstrates that the Leven and Linnhe glaciers became confluent and advanced ca 7km beyond the outwash spreads at Corran Ferry and North Balluchulish.

Firn line values calculated for the reconstructed glaciers range from a minimumof 356m in the SW to a maximum of 966m in the NE and imply that the main snow-bearing winds were from the SW with amounts of snow decreasing rapidly toward the NE. Trend surface analysis of the firn line data, of the altitudes of the floors of 303 corries and of the mean annual precipitation at 125 stations in western Scotland give added support to such a view and further demonstrate that the climatic parameters that operated in the past were broadly similar to those that operate at the present time. The results of such analyses strongly support the spatial pattern of firn line values,for the Loch Lomond Advance glaciers in the western Grampians,proposed by Sissons(1979c).

Broad similarities also existed between the pattern of build-up and ice-flow directions of the ice-sheet and the main ice mass of Loch Lomond Stadial age, as shown by similar movements of regional erratics and by similar ice-divide positions.

Topographical differences across the study area also become explicable when related to the palaeoclimatic conditions that prevailed during the stadial. The deeply dissected Western Mountain zone, replete with corries,deep glacial troughs,numerous glacial breaches and extensive areas of bare ice-scoured bedrock can be related to high-activity glaciers with high velocities and high discharge rates; farther E decreasing snowfall led to glaciers with lower-activity rates and glacial erosion was less severe than in the W. Such contrasting patterns of glacier activity are likely to have been repeated many times during the Quaternary Period.

14.2 Themes for future research

A number of glacial limits in the western part of the study area, especially where the former glaciers descended to altitudes lower than ca 500m O.D., should be regarded as only crude approximations since they are based on only minimal morphological evidence. Such considerations apply particularly to the ice-limits SE of Connel Bridge, in Glen Duror and along parts of the Creran-Etive Range(Figures 4.2 and 4.3). A more detailed investigation of the stratigraphic and morphological evidence in these areas,together with the use of pollen stratigraphy and radiocarbon dating,might help to resolve the problems of interpretation that have arisen in these areas.

The floors of the western sea lochs have yet to be investigated using either seismic(Boulton et al, 1981) or sediment sampling methods. Such investigations might be usefully applied to the areas in the vicinity of the inferred ice-limits of the Linnhe and Etive glaciers and to areas within the ice-limits where stand-stills may have occurred during ice retreat.

Detailed sedimentological work on the till sequences in the western Grampians has not yet been done and a good deal of scope exists for work to be carried out on till provenances,fabrics and geneses.

At least 15 major landslips(Figures 5.2 - 5.5) occur in the study area, but only preliminary investigations have been conducted to date(G.Holmes,pers.comm; present writer) to determine their mode or time of formation. Detailed investigations could provide valuable information on the significance of landslipping to the retreat of the mountain slopes during and following periods of glaciation.

Detailed palaeoclimatic inferences based on glacier reconstructions, using the techniques developed by Sissons and Sutherland(1976), have yet to be derived from the glaciological data described in this thesis. When more detailed palaeoclimatic data becomes available this should provide greater insights into the relationships between such data and former glacier régimes and between glacier régimes and glacial bedforms within the study area.

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EXPLANATORY NOTES ON APPENDIX A

Appendix A contains the essential data relating to the periglacial and glacial evidence observed on 277 cols and spurs and used to calculate ice-limits.

a. Column one contains the O.S. map six figure co-ordinates. All are prefixed by the grid letters NN.

b. Column two lists the reference number of each col or spur as shown on Figure 4.4. Cols are prefixed by the letter C and spurs by the letter S.

c. Column three provides the minimum altitude in metres of the periglacial evidence mapped along each spur.

d. The maximum altitude in metres of the glacial evidence on the spur is shown in column four.

e. The figures in column five are derived by summing the values in columns three and four and finding the mean value(spurs only). On some spurs only a lower limit to the periglacial evidence or an upper limit to the glacial evidence could be discerned. In these cases the altitude shown in column five represents either a maximum or a minimum altitude only. The values for the cols either represent a maximum altitude or a minimum altitude that can be imposed on the former glacier surface(see section 4.6).

f. The different forms of periglacial and glacial evidence are represented in column six by letters according to the classification shown below.

Periglacial evidence above the inferred ice-limit

f = frost-riven bedrock with angular edges and frost-widened joints.

sc = thick scree that is largely inactive.

s = debris-mantled slope.

l = solifluction lobes and/or terraces and/or sheets.

a = angular bedrock, but weakly frost-riven.

Glacial evidence below the inferred ice-limit

i = strongly ice-moulded bedrock.
fr = numerous friction cracks.
t = thick till.
b = many perched boulders.
r = roches moutonnées.
m = moundy drift.
h = hummocky moraine.

g. In column seven the type of evidence located on each spur or col is classified into thirteen categories according to the predominating periglacial/glacial evidence as described below.

A. The evidence consists primarily of bedrock surfaces that display contrasting periglacial and glacial features.

B. Solifluction lobes and/or sheets and/or terraces and debris-mantled slopes above the inferred limit contrast with ice-smoothed bedrock below the limit.

C. Frost-riven bedrock and/or thick relict scree above the limit contrast with slopes mantled with thick till below the limit.

D. The evidence is similar to that described in category C except that the till is in the form of mounds and/or ridges or is in the form of low hummocky moraine.

E & F. The type of periglacial evidence described in category B contrasts with slopes mantled with thick till below the limit or with the glacial evidence described in category D respectively.

G. This category is used for spurs where the field evidence is insufficient to place the ice-limit within a vertical zone of less than 60 metres. Hence the derived ice-limit value shown in column five is at best only a very crude approximation or at worst it may be in error by as much as 100m or more in some cases.

H & I. These categories contain the spurs on which only a clear lower limit to periglacial evidence or a clear upper limit to glacial evidence could be discerned respectively.

J. The altitude of either the periglacial or glacial evidence on these spurs is clearly at variance with the ice-limits derived from spurs and cols in the immediate vicinity and, therefore, is regarded as anomalous(see section 4.6).

K. Cols that contain periglacial evidence in the form of frost-riven bedrock and/or thick relict scree and/or solifluction lobes, terraces or sheets impose an absolute maximum altitude that could have been attained by the former glacier surface in the vicinity.

L. Cols that contain strongly ice-moulded bedrock and/or thick till and/or hummocky moraine are inferred to provide an absolute minimum altitude reached by the former glacier surface.

M. Cols that do not contain either positive periglacial or glacial evidence or display contradictory evidence are placed into this category. Their relationships with the Loch Lomond Advance limits have been determined from the evidence mapped in the vicinity.

APPENDIX

Spurs and cols

169748	S1	650	610	630	s,l/i	B
144725	C2			565	?	M
142735	S3	590	570	580	f/i	A
138734	S4		500	500	/i,b	I
153697	S5	640	600	620	f,s/i	A
160701	S6	880	850	865	f,s/i,r	A
173703	S7	830	800	815	f,s/i	A
179700	C8			565	/i,r	L
188722	C9			830	?	M
199701	S10	830	810	820	f/i	A
207700	S11	850	780	815	f,s,l/i,fr	G
213707	C12			733	/i,r	L
220707	S13	800		800	f,l/	H
222680	S14	790	730	760	f,s/i,fr,r	A
239669	S15	830	790	810	f,s/i	A
233654	S16	840	800	820	f,s/i	A
221656	S17	860	830	845	f,s/i	A
225655	C18			738	/l,b	L
221674	S19	840	790	815	f,s/i	A
219673	C20			744	/i,b	L
194658	C21			783	/i	L
203641	S22	720	700	710	f,s/i	A
182638	S23	670	650	660	f,s/i,fr	A
169652	C24			870	f,s/	K
179655	C25			831	f,s/	K
182659	C26			868	f,s/	K
187663	C27			857	?	M
189675	S28	820	730	775	f,s/i	G
174679	S29	790	750	770	f,s/i	A
155675	S30	800	650	725	f/i	G
145671	S31	580	550	565	f,s/t,i	E

159654	C32			761	f,s/	K
140649	S33		550	550	/t	I
132673	S34	600	450	525	f,s/i,t	G
129690	S35	700	700	700	f,s/i	J
145629	S36	550	510	530	f,l/i,fr,r	A
130634	S37	540	500	520	f,s/i,fr	A
111644	S38	500	460	480	f,s,l/i,r,b	A
100624	C39			465	f,l/	K
091631	S40	480	450	465	a,s/i,fr	A
090632	C41			405	/i,fr	L
093652	S42	400	300	350	s,a/i,t	G
089670	S43	350	300	325	f,s/i,t	E
084684	S44	400	150	275	s/m,b	G
061670	S45	450		450	f,s/	H
072606	S46	370	330	350	f/i,r	A
032590	S47	400	250	325	f/i,t	G
059584	S48	330	290	310	f,s/i,t	C
070579	S49	360	340	350	f,s/i,m	A
043518	C50			636	f,s/	K
058521	S51	400	350	375	f/i,t,b	C
069542	C52			395	f,s/	K
093542	C53			375	?	M
092541	C54			325	/i,b	L
096567	S55	450		450	s/	H
110566	S56	480	430	450	f/i,m	A
120547	S57	525	475	500	f/i,t	A
130555	S58	500	450	475	f,s/i,t	A
134575	S59	520	460	490	f,s/i	A
120597	S60	440	400	420	f,s/i,r,fr	A
139599	S61	490	450	470	f,s/i	A
148598	S62	510	490	500	f,s/i	A
161602	S63	570	530	550	f,s/i,t	A
172606	S64	630	590	610	f,s,l/i	A
183607	S65	660	660	660	f,sc,l/i,fr	A
168592	S66	640	620	630	f,sc,l/i,r,m	A
192588	S67	690	650	670	f,sc/i	A

208585	S68	680	660	670	f,sc/i	A
211575	S69	690	650	670	f,l/i	A
253634	S70	610	570	590	f,sc/i	J
248576	S71	720	680	700	f,sc/i	A
183571	S72	700	600	650	f,sc/i	G
173573	S73	520	480	500	f/i	A
139556	S74	550	500	525	f,sc/i	A
176558	S75	580	550	565	f/i,r	A
236566	S76	680	660	670	f,sc,l/i	B
251564	S77	730	700	715	f,sc/i	A
216548	S78	700	650	675	f,sc/i,r	A
221535	S79	660	620	640	f,sc/i,r	A
211531	S80	670	640	655	f,sc/i	A
204526	S81	670	640	655	f,sc/i	A
182521	S82	640	600	620	f,sc/i	A
203537	S83	690	650	670	f,sc/i	A
201554	S84	670	640	655	f,sc/i	A
188553	S85	660	620	640	f,sc/i	A
188547	C86			748	f,sc/	K
181543	S87	690	650	670	f,sc/i	A
179539	S88	690	650	670	f,sc/i	A
169535	C89			489	/i,t	L
166529	S90	660	620	640	f,sc/i	A
183529	C91			489	/i,t	L
163520	S92	620	580	600	l,sc/i,fr	B
143509	S93	575	450	512	f,sc/b	G
130511	S94	550	500	525	f,sc/i,r	A
131524	C95			579	?	M
153538	C96			710	f,l/	K
118513	C97			627	?	M
127498	S98	550	450	500	f/i	G
106489	S99	520	500	510	f,sc/i	A
111505	C100			450	/i,b,m,r	L
096508	S101	600	550	575	f,sc/i,r	A
104523	C102			525	/i,r	L
059491	S103	550	400	475	s/t	G

057520	S104	420	370	395	f/t,m,fr	C
014486	C105			391	s/	K
073468	S106	500	450	475	f/i	A
045441	C107			560	s,f/	K
018414	C108			570	f,sc,s/	K
020405	C109			530	f,sc,s/	K
024396	C110			465	/h,t	L
031406	S111	510	470	490	f/i,fr,r	A
039424	S112	510	460	485	f,sc/i	A
068429	S113		+586	+586	/i	J
079420	S114		+550	+550	/i	J
080451	S115	520	480	500	f/i,fr	A
102458	S116	480	450	465	f,sc/i	A
109403	S117	500	440	470	f/i	A
109404	S118	500	440	470	f/i,m	A
112419	S119	480	450	465	f,sc/i	A
132445	S120	550	500	525	f,sc,l/i	A
139425	C121			765	f,sc/	K
141439	S122	620	600	610	f,l/i	A
154433	S123	660	640	650	f,sc/i	A
162433	C124			738	sc,l/	K
155468	S125		+800	+800	/i	J
178454	S126	620	590	605	f/i	A
175445	C127			754	f,sc/	K
189460	C128			632	/i	L
175477	S129	600	550	575	f/i	A
159483	S130	560	500	530	f/i	A
168493	S131	600	550	575	f/i	A
191472	S132	610	570	590	f/i	A
189504	S133	580	550	565	f,s/i	A
198474	S134	670	650	660	s,l/i	B
195484	S135	600	550	575	f/i	A
204492	S136	580	530	555	f/i	A
209474	S137	650	630	640	f/i	A
210503	S138	650	610	630	f/i	A
208504	S139	640	600	620	f,sc,l/i	A

224511	S140	670	630	650	f/i	A
240523	S141	680	620	650	f,sc/i	A
258523	S142	650	620	635	f,sc/i	J
255516	C143			720	f,sc,l/	K
268511	S144	760	650	705	f,l/i	G
267492	S145	700	680	690	f,sc/i	A
264495	C146			685	s,l/	K
253490	S147	730	680	705	f/i	A
230485	C148			694	f,l/	L
230471	S149	730	690	710	f/i	A
249468	S150	700	660	680	f/i,fr	A
244459	C151			670	/i,fr	L
240459	S152	740	710	725	f/i	A
268458	S153	730	670	700	f/i,fr	A
257445	S154	670	630	650	f,sc/i,b	A
247451	S155	700	670	685	f,sc/i	A
239442	S156	685	655	670	f,sc/i	A
223448	S157	690	650	670	f,sc/i	A
216444	C158			590	/t,b	L
181430	S159	690	650	670	f/i,fr	A
171430	S160	700	670	685	f,l/i	A
143416	C161			630	/i,fr	L
132409	S162	680	620	650	f/i,r	A
119383	S163	540	480	510	f/i	A
116373	S164	560	460	510	f/i	G
161398	S165	600	560	580	f/i,fr	A
171405	S166	610	570	590	f/i,fr	A
026372	C167			470	f,s/	K
030362	C168			455	f,s/	K
023352	C169			305	/i	L
028349	S170	500	400	450	f/i,b	G
051331	S171	420	380	400	f,s/i,m	F
074316	S172	650	600	625	f/i	A
104319	C173			564	f,l/	K
080342	S174	490	450	470	f/i	A
111349	S175	520	460	490	f/i	A
125340	C176			480	/i,r	L

152350	S177	550	420	485	f,sc/i,t	G
163358	S178	550	520	535	f,sc/i,t,h	A
170364	S179	550	530	540	f/i,t,h	D
169373	S180	560	520	540	f/i,t	A
183380	S181	530	480	505	f/i	A
185365	C182			483	?	M
195375	C183			396	/i,t	L
204378	S184	550	520	535	f.l/i,fr	A
209372	S185	520	480	500	f/i,r	A
205393	C186			585	?	M
208403	S187	610	580	595	f/i,h	A
225375	S188	520	500	510	f,sc/t,i	F
247376	C189			520	/h	L
255384	S190	560	520	540	f/i,t	A
259377	C191			500	/h,t	L
250371	S192	540	520	530	f,sc/t,h	D
279394	S193	560	520	540	f,sc,l/i	A
318398	S194	610	550	580	f,sc/t,h	D
298363	S195	650	400	525	f,l/t,h	G
280348	S196	520	480	500	f,sc,l/i,h	D
224369	C197			370	/t,h	L
225363	S198	500	460	480	f,l/i,h	A
214341	S199	450	420	435	s,l/t,h	E
193334	S200	390	370	380	s,sc/t,h	E
178353	C201			515	s,l/	K
181348	S202	440	400	420	f,s/t,i	E
175343	S203	450	420	435	f,sc/i,fr	A
158327	S204	360	340	350	f/i	A
161340	C205			612	f,s,l/	K
230710	S206	840		840	f,sc,l/	H
243719	S207	840	800	820	f,l,sc/i,fr,t	E
261715	S208	770	750	760	l,s/i	B
259713	C209			750	/i	L
267729	C210			803	/i,fr,r	L
266731	S211	880	840	860	f,sc,l,s/i,fr,r	B
273739	S212	750	730	740	f,sc,l/i,fr	A

277749	S213	700	650	675	f/i	A
282757	S214	640	610	625	f,sc/i	A
284754	C215			586	/i	L
289747	S216	680	650	665	f,sc/i	A
288742	C217			540	/i	L
298725	S218	730	700	715	f,sc/i	A
279705	S219	700	680	690	f,l/i	A
301709	S220	680	650	665	f,sc/i	A
313708	S221	630	610	620	f,sc/i	A
350734	S222	650	610	630	f,sc,l/i	B
343696	S223	650	630	640	f,sc/i	A
262648	S224	690	650	670	f,sc/i	A
311627	S225	730	700	715	f,sc,l/i	B
330630	S226	750	680	715	f,sc/i	G
323642	C227			796	f,sc,l/	K
341650	S228	740	720	730	f,sc/i,b	A
333663	S229	670	610	640	f,/t,i	C
360687	S230	660	640	650	f,sc,l/i,b	A
387695	S231	700	620	660	f/b,h	F
385722	S232	700	550	625	f,l/i,t	G
376725	S233	660	640	650	f,s/i,h	A
390740	S234	650	600	625	f/t	C
393716	C235			589	/i,r	L
402710	S236		550	550	/i,r,b	I
422721	S237	670	620	645	f,s/t,h	F
436716	S238	680	620	650	f/t	C
468732	S239	760	720	740	f/t,h	C
481732	C240			722	/h,b	L
465717	C241			653	/t,h	L
430692	S242	660	620	640	f/t,h	C
406677	S243	680	650	665	f/t	C
358571	S244	700	670	685	f/i	A
445618	S245	660	600	630	f/i,b	A
459627	S246	650	600	625	f,s/t,i	A
461641	S247	650	610	630	f/i	A
445654	C248			675	i,r	M
460689	S249	700	670	685	f,s/t,m	D

502688	S250	600	580	590	f,sc/i,b	A
513653	S251	620	570	595	f,sc/i	A
540650	S252	600	580	590	s,l/t	E
531529	S253	640	600	620	f/i,fr	A
500513	S254	630	600	615	f/i,fr	A
521488	C255			695	/m	L
468482	C256			705	/m	L
454473	C257			695	/m	L
435468	C258			650	/m	L
388461	S259	650	630	640	f,l,s/i,t	E
350440	S260	680	640	660	s,l/t	E
353434	C261			813	f,l,s/	K
323420	S262	680	640	660	f,sc,l/i,t	A
341405	C263			760	f,sc,l/	K
325399	C264			746	f/	K
343391	S265	720	680	700	s/i,b,t	E
386412	S266	730	690	710	f/i,r	A
416432	S267	750	650	700	f/t,r	G
419451	C268			745	s,f/	K
435440	C269			707	/t,r,b	L
444435	S270	750	710	730	f/t	C
458446	S271	720	670	695	f/h	D
465468	S272	740	650	695	f/t	C
487491	C273			744	f,s/	K
506480	S274	730	710	720	f/i	A
505454	S275	740	720	730	f/i	A
529478	S276	730	710	715	f/i,r	A
549474	S277	700	600	650	f,s/t,h	G

Summary statistics for spurs and cols

Type	Category	Number	Totals
Spur	A	132	
Spur	B	8	
Spur	C	11	
Spur	D	6	
Spur	E	10	
Spur	F	4	
Spur	G	20	
Spur	H	4	
Spur	I	3	
Spur	J	6	204
Col	K	29	
Col	L	35	
Col	M	9	73
			277

EXPLANATORY NOTES ON APPENDIX B

Appendix B contains the data derived from the measurement of selected parameters of 271 corries in the western Grampians.

a. Column one lists the reference number of each corrie as shown on Figure 6.1.
b. O.S. map six figure grid references are given in column two. All are prefixed by the grid letters NN.
c. Column three provides the azimuth of the corrie measured in degrees east of grid north.
d. Column four gives the altitude of the corrie floor in metres a.m.s.l. as measured at the base of the backwall or from the surface altitude of a lochan where present.
e. The width of the corrie in metres is shown in column five.
f. The length of the corrie in metres is shown in column six.
g. Column seven provides the vertical height of the corrie backwall in metres from the foot to the crest of the backwall.
h. Column eight gives the mean angle of slope of the corrie backwall in degrees.
i. Column nine provides length/width ratios calculated by dividing length by width.
j. Column ten provides length/depth ratios calculated by dividing length by depth.
k. Corrie volumes are shown in column eleven in km^3 and were calculated by multiplying width, length and depth and dividing by two.

APPENDIX B

Corrie measurement data

Ben Nevis Range

1	162718	45	915	580	580	410	42	1.000	1.415	0.069
2	173713	42	945	640	580	373	52	0.906	1.555	0.069
3	158704	90	853	580	630	160	35	1.086	3.938	0.011
4	173706	98	850	530	365	277	45	0.689	1.318	0.027
5	153708	290	853	530	470	183	37	0.887	2.568	0.023
6	173734	53	914	480	370	168	39	0.771	2.202	0.015
7	177730	64	1006	575	650	172	43	1.130	3.779	0.032
8	178726	68	1036	800	630	187	46	0.788	3,369	0.047
9	180719	136	970	580	420	164	45	0.724	2.561	0.020
10	194743	70	1000	600	750	190	41	1.25	3.947	0.043
11	197739	78	838	670	740	374	43	1.104	1.979	0.093
12	198735	40	900	700	830	310	27	1.186	2.677	0.090
13	197721	52	885	740	530	331	46	0.716	1.601	0.065
14	209739	63	790	900	750	125	23	0.83	6.000	0.042
15	221712	354	793	630	450	60	45	0.714	7.500	0.002
16	206703	90	800	620	620	165	45	1.000	3.758	0.032
17	239730	20	862	1000	740	205	28	0.74	3.610	0.076
18	235725	160	945	840	610	135	37	0.726	4.519	0.035
19	244726	110	930	680	430	185	31	0.632	2.324	0.027
20	250732	5	976	630	540	128	31	0.857	4.219	0.022
21	254733	70	968	690	310	112	33	0.449	2.768	0.012
22	258739	330	953	690	540	205	30	0.783	3.553	0.038
23	263749	40	825	740	620	287	31	0.838	2.160	0.066
24	268742	140	825	1000	880	352	33	0.880	2.500	0.155

Mamore Forest Range

25	117659	350	760	720	710	140	32	0.986	5.071	0.036
26	123670	40	650	1000	750	260	39	0.750	2.885	0.098
27	118654	133	720	530	500	190	21	0.943	2.632	0.025
28	126663	74	730	800	480	200	32	0.600	2.400	0.038
29	134656	360	750	850	450	160	33	0.529	2.813	0.031

30	144660	310	750	700	740	160	35	1.057	4.625	0.046
31	148659	24	750	400	300	162	39	0.75	1.852	0.010
32	151655	5	716	530	320	283	41	0.604	1.131	0.024
33	162662	270	730	500	650	260	46	1.300	2.500	0.042
34	161654	310	770	530	380	190	35	0.717	2.000	0.019
35	164669	20	922	425	730	177	32	1.718	4.124	0.027
36	170674	40	823	600	430	140	39	0.717	3.071	0.018
37	168669	130	885	290	270	214	47	0.931	1.262	0.008
38	165664	140	890	380	340	100	53	0.895	3.400	0.006
39	165658	43	855	575	500	146	44	0.870	3.425	0.021
40	170653	23	790	900	380	211	31	0.422	1.801	0.036
41	168648	134	680	700	800	235	40	1.143	3.404	0.066
42	188663	68	793	640	520	208	37	0.813	2.500	0.035
43	184658	130	730	480	300	179	50	0.625	1.676	0.013
44	180654	120	680	630	460	352	40	0.730	1.307	0.051
45	180648	35	800	250	300	100	38	1.200	3.000	0.004
46	205655	305	825	1320	1000	230	30	0.758	4.348	0.200
47	205646	80	800	650	600	255	40	0.923	2.353	0.050
48	214669	30	825	800	620	303	33	0.775	2.046	0.075
49	215665	60	824	320	620	304	31	1.938	2.039	0.030
50	214657	55	855	1100	900	195	31	0.818	4.615	0.097
51	233658	90	808	620	450	202	34	0.726	2.228	0.028

<u>Aonach Eagach Range</u>

52	135585	342	700	630	640	210	40	1.016	3.048	0.042
53	141589	6	700	360	480	170	40	1.333	2.824	0.015
54	144584	11	810	695	500	157	32	0.719	3.185	0.027
55	153584	44	720	760	600	220	39	0.789	2.727	0.050
56	166584	54	731	755	620	209	35	0.821	2.967	0.049
57	184580	14	640	900	750	220	36	0.833	3.409	0.074
58	195581	46	648	960	850	232	36	0.885	3.664	0.095
59	174602	25	730	600	380	130	25	0.633	2.923	0.015
68	135581	128	810	370	250	100	30	0.676	2.500	0.005
69	174581	166	693	800	570	217	38	0.713	2.627	0.049

Creran-Duror Range

61	026565	68	610	500	400	290	49	0.800	1.379	0.029
62	033557	5	620	1100	500	204	34	0.455	2.451	0.056
63	043558	348	690	500	550	311	44	1.100	1.768	0.043
64	049557	7	640	1100	650	290	40	0.591	2.241	0.104
65	060561	360	830	670	620	194	31	0.925	3.196	0.040
66	067567	66	530	550	550	320	39	1.000	1.719	0.048
67	101561	356	560	450	490	70	13	1.089	7.000	0.008
70	019523	355	610	500	450	100	27	0.900	4.500	0.011
71	030519	347	690	520	500	189	37	0.962	2.646	0.017
72	042518	348	580	1200	500	280	29	0.417	1.786	0.008
73	000500	45	510	330	300	130	41	0.909	2.308	0.006
74	025496	90	400	760	900	150	27	1.184	6.000	0.051
75	026504	96	500	380	260	87	35	0.684	2.990	0.004
76	033514	152	600	630	500	279	33	0.794	1.792	0.010
77	054524	60	550	450	300	110	26	0.667	2.727	0.007

Bidean nam Bian Range

78	096531	360	660	380	460	90	31	1.211	5.111	0.008
79	099531	345	680	380	410	60	39	1.079	6.833	0.005
80	112532	317	650	880	400	344	39	0.455	1.163	0.060
81	122525	47	710	420	550	258	33	1.310	2.132	0.030
82	120529	62	690	400	390	160	33	0.975	2.438	0.012
83	108515	171	740	880	580	170	40	0.660	3.412	0.043
84	115518	108	790	680	400	160	33	0.588	2.500	0.022
85	125512	30	560	450	300	160	47	0.667	1.875	0.011
86	088499	280	730	400	340	140	29	0.850	2.429	0.010
87	093501	7	730	250	450	200	34	1.800	2.250	0.011
88	098501	8	670	450	420	280	45	0.933	1.500	0.026
89	104500	355	620	630	400	280	41	0.635	1.429	0.035
90	110499	90	650	390	260	170	40	0.667	1.529	0.009
91	136547	6	890	300	460	160	52	1.533	2.875	0.011
92	144545	335	900	750	490	241	36	0.653	2.033	0.044
93	149553	70	860	900	570	255	36	0.633	2.235	0.065
94	147544	102	920	630	540	230	40	0.857	2.348	0.039
95	157537	84	823	740	400	227	37	0.541	1.762	0.034

96	180538	352	731	500	390	227	39	0.780	1.718	0.022
97	193527	146	823	400	300	133	34	0.750	2.256	0.008
98	203532	338	716	740	460	214	31	0.622	2.149	0.036
99	217543	24	731	580	610	161	26	1.052	3.789	0.028
100	204529	143	747	600	330	74	26	0.550	4.460	0.007
101	214538	147	747	1000	700	264	28	0.700	2.652	0.092
60	248571	123	640	1060	800	206	30	0.755	3.883	0.087

Creran-Etive Range

107	013412	345	550	670	400	130	27	0.597	3.077	0.017
108	027421	79	650	830	880	140	29	1.060	6.286	0.051
109	021434	2	530	390	400	100	39	1.026	4.000	0.008
110	042452	357	710	380	480	140	27	1.263	3.429	0.013
111	048457	46	690	620	520	140	27	0.839	3.714	0.023
112	058464	88	750	350	300	187	34	0.857	1.604	0.010
113	083444	350	610	800	890	190	22	1.113	4.684	0.068
114	091445	340	670	620	620	97	38	1.000	6.392	0.019
130	029371	63	490	620	850	180	33	1.371	4.722	0.047

Ben Starav Range

102	243520	40	701	560	630	295	36	1.125	2.136	0.032
103	239515	86	899	630	440	198	28	0.698	2.222	0.027
104	242503	26	853	660	520	215	32	0.788	2.419	0.037
105	244496	92	780	870	1250	318	35	1.437	3.931	0.173
106	233494	274	853	800	880	206	26	1.100	4.272	0.073
115	169483	260	660	500	420	70	19	0.840	6.000	0.007
116	170489	10	700	420	300	160	39	0.714	1.875	0.010
117	173492	300	700	500	310	145	49	0.620	2.138	0.011
118	180494	9	680	400	480	154	31	1.200	3.117	0.015
119	176486	82	640	430	300	117	43	0.698	2.564	0.008
120	224484	360	640	800	830	206	27	1.038	4.029	0.068
121	196490	360	710	880	570	180	33	0.648	3.167	0.045
122	207467	36	579	830	550	320	33	0.663	1.719	0.073
123	212461	325	790	830	500	140	35	0.602	3.571	0.029
124	223464	336	760	1000	800	230	33	0.800	3.478	0.092
125	228476	55	670	630	580	120	20	0.921	4.833	0.022

126	224453	142	853	800	600	123	24	0.750	4.878	0.030
127	235459	69	760	770	1100	327	33	1.429	3.364	0.138
128	238450	118	610	680	620	395	28	0.912	1.570	0.083
129	264461	35	730	750	750	202	32	1.000	3.713	0.057
131	119412	345	700	630	280	122	39	0.445	2.295	0.011
132	124413	290	760	410	260	140	39	0.634	1.857	0.007
133	127418	250	750	650	480	160	33	0.738	3.000	0.025
134	121428	318	750	450	300	250	48	0.667	1.200	0.017
135	123432	322	700	550	750	370	38	1.364	2.027	0.076
136	128443	12	600	490	400	190	37	0.816	2.105	0.019
137	130435	80	800	500	320	70	43	0.640	4.571	0.006
138	129428	40	920	500	430	148	48	0.860	2.905	0.016
139	133428	50	770	500	440	250	45	0.880	1.760	0.028
140	141427	330	680	630	500	200	39	0.794	2.500	0.032
141	147430	340	630	890	890	280	33	1.000	3.179	0.111
142	157432	11	770	820	430	227	37	0.524	1.894	0.040
143	164431	142	630	820	630	260	43	0.768	2.423	0.067
144	175437	170	760	830	1000	284	28	1.205	3.521	0.118
145	167440	180	920	520	700	80	18	1.346	8.750	0.015
146	163454	40	650	700	770	300	37	1.100	2.567	0.081
147	171448	30	750	900	570	294	30	0.633	1.939	0.039
148	182456	349	620	500	580	210	23	1.160	2.762	0.030
149	190455	355	680	630	630	240	28	1.000	2.625	0.048
150	196447	122	760	500	500	140	35	1.000	3.571	0.018
151	128404	270	530	620	820	124	35	1.325	6.613	0.032
152	129411	100	810	400	280	108	23	0.700	2.593	0.006
153	138422	162	720	840	860	180	22	1.024	4.778	0.055
154	131424	130	930	500	500	138	38	1.000	3.623	0.017
155	153411	360	610	1150	630	330	31	0.548	1.909	0.120
156	170402	180	500	1000	570	102	20	0.570	5.588	0.029
157	160408	178	670	500	500	80	28	1.000	6.250	0.010
158	159404	110	660	440	620	90	12	1.409	6.889	0.012
159	157402	120	720	500	600	130	25	1.200	4.615	0.020
160	156397	140	594	480	600	100	27	1.250	6.000	0.014
188	210405	310	610	500	500	53	23	1.000	9.434	0.007

Ben Cruachan Range

161	052321	270	610	320	400	230	28	1.250	1.739	0.015
162	058322	20	560	640	700	300	35	1.094	2.333	0.067
163	061314	37	620	550	850	286	36	1.600	3.077	0.069
164	068309	5	800	900	650	304	34	0.722	2.138	0.089
165	063298	225	700	850	1220	350	27	1.435	3.486	0.181
166	073299	115	670	840	900	456	33	1.071	1.974	0.172
167	089301	195	630	1000	850	340	31	0.850	2.500	0.145
168	101289	80	540	1000	1200	356	35	1.200	3.371	0.214
169	099308	105	670	760	500	310	32	0.658	1.613	0.059
170	078308	15	690	1000	1200	390	32	1.200	3.077	0.234
171	087309	50	760	780	1000	249	42	1.282	4.016	0.097
172	095313	325	700	500	570	290	40	1.140	1.966	0.041
173	101336	20	570	650	520	190	35	0.800	2.737	0.032
174	109332	5	750	430	350	230	40	0.814	1.522	0.017
175	116330	15	740	900	500	200	34	0.556	2.500	0.045
176	125330	350	590	1000	820	290	36	0.820	2.828	0.119
177	110325	170	790	380	420	160	33	1.105	2.625	0.013
178	121324	140	670	500	640	200	32	1.280	3.200	0.032
179	128328	191	670	750	650	290	25	0.867	2.241	0.071
180	143328	90	550	820	1200	439	32	1.463	2.733	0.216
181	134333	328	660	900	870	310	36	0.970	2.806	0.121
182	143338	360	750	490	500	110	41	1.020	4.545	0.013
183	148343	350	570	550	580	200	34	1.055	2.900	0.032
184	150336	100	620	560	680	260	31	1.214	2.615	0.050
185	157345	30	500	520	750	300	34	1.442	2.500	0.059
186	174346	98	520	390	490	80	33	1.256	6.125	0.008
187	216356	345	520	760	550	230	35	0.724	2.391	0.048

Creag Meagaidh Range

189	385865	20	850	1500	1000	150	31	0.666	6.600	0.113
190	388859	140	850	500	370	190	35	0.740	1.947	0.018
191	385855	70	840	800	400	210	35	0.500	1.905	0.034
192	388848	50	910	460	400	110	29	0.870	3.636	0.010
193	389844	85	860	500	480	180	31	0.960	2.600	0.022
194	418893	340	670	1250	500	180	39	0.400	2.700	0.056

195	421881	300	910	800	650	190	35	0.813	3.421	0.049
196	419870	180	730	1000	1250	370	45	1.250	3.378	0.231
197	427859	165	820	1000	500	160	39	0.500	3.125	0.040
198	432868	90	810	1000	700	210	40	0.700	3.300	0.074
199	445878	325	830	500	500	215	28	1.000	2.326	0.027
200	435880	62	620	1360	1250	430	36	0.919	2.907	0.366
201	431885	92	800	750	650	270	40	0.860	2.407	0.066
202	442894	150	840	1050	460	190	21	0.438	2.421	0.046
203	436899	356	820	890	400	126	35	0.449	3.175	0.022
204	488908	140	750	750	900	210	35	1.200	4.286	0.071
205	500917	110	750	800	650	160	22	0.813	4.063	0.042
206	507925	2	660	750	300	220	29	0.400	1.364	0.025

Ben Alder Range

207	309736	40	823	1000	1000	293	36	1.000	3.413	0.147
208	309723	120	740	620	480	160	35	0.744	3.000	0.024
209	321736	128	823	620	500	277	43	0.806	1.805	0.043
210	325740	98	884	630	470	122	35	0.746	3.852	0.018
211	328748	90	790	560	300	185	37	0.536	1.622	0.016
212	362742	50	740	1250	1120	235	32	0.896	4.766	0.165
213	376734	160	762	760	750	206	42	0.987	3.641	0.059
214	381738	110	853	580	620	183	31	1.069	3.388	0.033
215	381698	72	792	600	500	122	21	0.833	4.098	0.018
218	422731	325	670	650	650	80	22	1.000	8.125	0.017
219	430738	330	800	750	650	100	27	0.860	6.500	0.024
220	440742	355	825	550	750	150	21	1.364	5.000	0.031
221	450738	352	890	900	750	210	40	0.830	3.571	0.071
222	460745	345	915	1200	500	200,	34	0.417	2.500	0.060
223	439731	160	885	1000	1000	145	24	1.000	6.897	0.073
224	456737	160	945	1000	900	91	37	0.900	9.890	0.041
225	479756	30	805	1150	1250	305	29	1.087	4.098	0.319
226	488749	70	700	1250	1400	400	28	1.120	3.500	0.350
227	473785	330	825	650	1050	245	39	1.615	4.286	0.084
228	482784	98	885	700	650	115	25	0.929	5.652	0.026
229	493798	118	670	1250	1500	240	31	1.200	6.250	0.225

230	509816	30	900	740	500	100	22	0.676	5.000	0.019
231	522773	87	640	650	300	180	36	0.462	1.600	0.018
238	482722	315	945	650	700	100	29	1.077	7.000	0.023
239	502731	60	855	650	550	205	39	0.846	2.683	0.037
240	503726	120	800.	750	600	280	43	0.800	2.143	0.063
241	499714	68	830	1100	750	318	39	0.682	2.358	0.131
242	520721	60	890	800	550	110	30	0.688	5.000	0.024
243	514707	118	740	900	750	250	38	0.833	3.000	0.084
244	585748	2	850	740	340	120	31	0.459	2.833	0.015
245	599733	27	880	650	400	110	40	0.615	3.636	0.014

Rannoch Forest Range

216	316635	150	625	1210	1100	198	28	0.909	5.556	0.132
217	331650	25	670	1260	1250	220	30	0.992	5.682	0.273
232	416668	10	700	1150	1000	241	26	0.870	4.149	0.139
233	458697	85	710	1100	1000	115	19	0.909	8.696	0.063
234	449682	90	650	1400	900	280	26	0.643	3.214	0.176
235	448668	90	670	1250	350	148	26	0.280	2.365	0.032
236	455645	40	580	1500	500	250	35	0.333	2.000	0.094
237	454636	120	710	500	350	128	27	0.700	2.734	0.011

Orchy-Lyon Range

246	322399	270	610	710	580	240	44	0.817	2.417	0.049
247	324398	10	710	200	400	125	21	2.000	3,200	0.005
248	328385	350	620	810	810	60	26	1.306	13.500	0.020
249	332392	10	790	620	630	202	32	1.016	3.119	0.039
250	331380	130	760	420	480	280	37	1.143	1.714	0.028
251	338388	88	685	760	820	285	30	1.079	2.877	0.089
252	326413	27	671	1000	750	329	45	0.750	2.280	0.123
253	338409	60	760	740	880	242	33	1.189	1.189	0.079
254	345417	115	884	630	360	38	17	0.571	9.474	0.004
255	345425	91	831	1220	800	172	30	0.656	4.651	0.084
256	352438	22	671	760	880	366	36	1.158	2.404	0.122
257	364436	105	884	630	260	76	45	0.412	3.421	0.006
258	378439	90	869	350	360	198	31	1.029	1.818	0.012

259	377444	50	945	380	430	136	47	1.132	3.162	0.011
260	367446	10	700	1100	680	240	47	0.618	2.833	0.090
261	381454	90	732	760	400	175	41	0.526	2.286	0.027
262	411439	95	680	900	950	170	34	1.050	5.588	0.073

Rannoch-Lyon Range

263	485431	145	650	1000	750	214	28	0.750	3.505	0.080
264	476449	358	650	800	550	160	39	0.688	3.438	0.035
265	487448	3	700	780	890	260	48	1.141	3.423	0.090
266	498447	86	750	920	1000	150	37	1.087	6.600	0.069
267	501494	70	790	900	750	142	30	0.830	5.282	0.048
268	630508	6	750	1250	1250	220	20	1.000	5.682	0.172
269	638505	75	800	850	750	220	27	0.824	3.182	0.065
270	651519	20	820	810	1000	160	27	1.236	6.250	0.065
271	673513	160	860	1120	750	130	31	0.670	5.769	0.055

APPENDIX C

Statistical analysis of glacierized and non-glacierized corries

The aim was to find if the altitude (defined as the height of the corrie floor(A) at the base of the backwall in metres above mean sea level) and the shape of a corrie (defined by the parameters of width(W), length(L), depth(D) and backwall gradient(G) in metres or degrees) determined whether or not a corrie glacier developed during the Loch Lomond Stadial.

A total of 145 corries was selected and these were subdivided into two categories on the basis of field evidence according to whether or not they contained a corrie glacier during the stadial. The 124 glacierized corries and the 21 non-glacierized corries were further subdivided into four categories according to aspect(i.e. NW, NE, SE and SW) in order to eliminate the influence of this variable.

Analysis of the glacierized and non-glacierized corries was undertaken by two methods:

i) Discriminant analysis.

ii) Minimum distance cluster analysis.

These perform the same function, although the first method assumes a particular distribution of the data and requires the calculation of parameters.

First, however, it was necessary to represent W,L,D,A and G with a single variable. This was done using principal factor analysis. The purpose of this was to make it easier to classify the corrie with only one variable instead of five.

A formula for a new variable representing W,L, D and A was found for the glacierized and non-glacierized data in the NW quadrant only:

Glacierized:

$Y = 0.267W + 0.239L + 0.263D + 0.232A$ explaining 67% of the variation in W,L,D and A.

Non-glacierized:

$Y_1 = 0.75W + 0.60L + 0.22D + 0.19A$ explaining 40.5%

$Y_2 = -0.02W + 0.19L + 0.70D + 0.69A$ explaining 37.1%

The regression of the new variable Y on the gradient and the correlation coefficient of Y and G were found.

Glacierized:

$G = 15.09 \log_{10} Y - 11.10$ Correlation coefficient $\log_{10}$ 0.177 (not significant)

The regression line of G against Y_1 was found:

$G = -0.005304\ Y_1 + 40.29$ Correlation coefficient -0.246 (not significant)

The regression line for G and Y_2 was not calculated, since the correlation for G and Y_2 was seen to be small from a graph.

Since the correlation in each case was too small to be significant the use of the backwall gradient to represent the parameters of W,L,D and A could not be used to estimate accurately the value of Y.

Discriminant analysis classification method

The aim is to find a formula that will indicate from the values of W,L,D,A and G whether a corrie is glacierized or non-glacierized. The method was carried out on all 21 non-glacierized observations and on a randon sample of 21 of the 124 glacierized observations. This method assumes that the observations have a normal distribution.

The result was to classify a corrie as 'glacierized' if: $459W + 324L + 988D - 170A + 6370G > 708664$ otherwise it would be classified as 'non-glacierized'.

The results given by the formula of the 21 glacierized and non-glacierized corries were as follows:

Glacierized				Non-glacierized			
q	n	f	c	q	n	f	c
NW	1	814590	G	NW	1	670936	NG
NW	2	704290	NG	NW	2	773460	G
NW	3	876455	G	NW	3	442080	NG
NW	4	769940	G	NW	4	644060	NG
NW	5	836500	G	NW	5	492710	NG
NW	6	888830	G	NW	6	499370	NG
NW	7	776740	G	NW	7	756452	G
NE	8	561786	NG	NW	8	497650	NG
NE	9	705502	NG	NE	9	622144	NG
NE	10	562110	NG	NE	10	544346	NG
NE	11	661560	NG	NE	11	653571	NG
NE	12	1113916	G	NE	12	905140	G
NE	13	774810	G	NE	13	357970	NG
NE	14	1004632	G	NE	14	933120	G
NE	15	1020794	G	NE	15	586848	NG
SE	16	791620	G	NE	16	484550	NG
SE	17	998002	G	SE	17	673590	NG
SE	18	690994	NG	SE	18	544490	NG
SE	19	589193	NG	SE	19	695330	NG
SE	20	727956	G	SE	20	625410	NG
SW	21	927510	G	SW	21	578380	NG

q = quadrant   n = number   f = formula   c = classification

Misclassifications

quadrant	glacierized	non-glacierized	total	percentage
NW	1 out of 7	2 out of 8	3 out of 15	20
NE	4 out of 8	2 out of 8	6 out of 16	37.5
SE	2 out of 5	0 out of 4	2 out of 9	22
SW	0 out of 1	0 out of 1	0. out of 2	0
Total	7 out of 21	4 out of 21	11 out of 42	
Percentage	33	19	26	

Minimum distance classification method

This method unlike the previous one is non-parametric. Since parameters do not have to be calculated and no distribution is assumed this method could be considered to be preferable since it does not impose a structure on the data. The same sample of glacierized corries was used as in the previous method.

The method is to find the 'distance' between each possible pair of corries. The distance is defined by:

$$d = (W_1 - W_2)^2 + (L_1 - L_2)^2 + (D_1 - D_2)^2 = (A_1 - A_2)^2 + (G_1 - G_2)^2$$

where W_1; L_1, D_1 etc are parameters of the first corrie and W_2, L_2, D_2 etc are parameters of the second corrie.

A table showing the distance between each pair of corries was compiled. From the table a dendrogram was drawn to link those corries that were closest together as this should indicate natural groupings of corries. The purpose was to see if these groupings coincided with the glacierized and non-glacierized observations.

However, the method was not successful as the diagram did not show any clear groups of glacierized and non-glacierized corries.

EXPLANATORY NOTES ON APPENDIX D

Appendix D lists the precipitation data for 124 stations in western Scotland provided by the meteorological Office, Bracknell. The data relates to the period 1941 to 1970. The data for station 39 were derived from Hann(1912) and relates to records kept for the summit of Ben Nevis(1344m) for the years 1883-1903.

a. Column one lists the reference number of each station. Only the reference numbers of the stations within the study area are shown on Figure 8.6.
b. Map co-ordinates are listed in column two. These were calculated in units equivalent to 4 kilometres east and north of O.S. map reference NM780140.
c. The mean annual precipitation in millimetres is shown in column three.
d. The altitude of the station in metres O.D. is listed in column four.
e. Column five shows the distance in kilometres of the station east of Northing NM 780.
f. Column six shows the distance in kilometres of the station north of Easting NM 140.

APPENDIX D

Precipitation data for western Scotland

1	003136	2216	15	1.2	54.4
2	014098	2618	12	5.6	39.2
3	038005	2286	288	15.2	2.0
4	038055	1493	3	15.2	22.0
5	031059	1537	5	12.4	23.6
6	039067	1596	6	15.6	26.8
7	034081	1699	8	13.6	32.4
8	040114	2564	8	16.0	45.6
9	042004	2191	98	16.8	1.6
10	045069	1698	15	18.0	27.6
11	049103	1676	12	19.6	41.2
12	055009	2089	46	22.0	3.6
13	056027	2188	201	22.4	10.8
14	059123	2072	4	23.6	49.2
15	054124	2464	18	21.6	49.6
16	068037	2197	61	27.2	14.7
17	068101	2850	296	27.2	40.4
18	063124	2029	15	25.2	49.6
19	061124	1975	22	24.4	49.6
20	062138	2358	15	24.8	55.2
21	069160	1802	30	27.6	64.0
22	075035	2452	38	30.0	14.0
23	076039	3095	427	30.4	15.6
24	071051	2116	30	28.4	20.4
25	070123	2121	128	28.0	49.2
26	074127	2231	146	29.6	50.8
27	078147	1920	11	31.2	58.8
28	079149	1930	66	31.6	59.6
29	089002	2122	9	35.6	0.8
30	083044	3714	503	33.2	17.6
31	080113	2321	26	32.0	45.2
32	088152	1702	23	35.2	60.8

33	096003	2537	415	38.4	1.2
34	095015	2437	351	38.0	6.0
35	096017	2442	366	38.4	6.8
36	098035	1978	61	39.2	14.0
37	097094	3071	73	38.8	37.6
38	091136	2256	52	36.4	54.4
39	096143	4081	1344	38.4	57.2
40	092151	2320	305	36.8	60.4
41	091171	1977	46	36.4	68.4
42	099176	1869	30	39.6	70.4
43	109123	2666	344	43.6	49.2
44	102154	2400	564	40.8	61.6
45	101174	1767	24	40.4	69.6
46	110003	2452	311	44.0	1.2
47	118008	3176	328	47.2	3.2
48	114018	2930	351	45.6	7.2
49	118038	2509	200	47.2	15.2
50	119051	2347	111	47.6	20.4
51	116115	2502	335	46.4	46.0
52	118128	2727	343	47.2	51.2
53	110154	2558	320	44.0	61.6
54	119162	1642	198	47.6	64.8
55	110170	1594	61	44.0	68.0
56	129105	2152	361	51.6	42.0
57	125116	2256	326	50.0	46.4
58	128123	2189	366	51.2	49.2
59	124153	2486	503	49.6	61.2
60	131064	2413	186	52.4	25.6
61	136123	1980	326	54.4	49.2
62	132138	1881	251	52.8	55.2
63	138159	1967	320	55.2	63.6
64	143131	2005	411	57.2	52.4
65	143132	2036	411	57.2	52.8
66	142161	1751	259	56.8	64.4
67	156148	1875	360	62.4	59.2
68	156169	1484	262	62.4	67.6

69	161109	1665	307	64.4	43.6
70	171108	1382	242	68.4	43.2
71	187111	1124	206	74.8	44.4
72	183123	1428	365	73.2	49.2
73	189160	1519	433	75.6	64.0
74	181162	1402	251	72.4	64.8
75	203099	1527	390	81.2	39.6
76	203106	1134	232	81.2	42.4
77	215147	1563	416	86.0	58.8
78	238083	1268	130	95.2	32.2
79	236126	1197	262	94.4	50.4
80	236133	1238	434	94.4	53.2
81	231143	1280	317	92.4	57.2
82	233163	1686	497	93.2	65.2
83	214177	1206	358	85.6	70.8
84	051224	3056	216	20.4	89.6
85	061233	2906	232	24.4	93.2
86	061223	2874	213	24.4	89.2
87	062222	2814	173	24.8	88.8
88	071211	2518	143	28.4	84.4
89	080218	2225	146	32.0	87.2
90	098219	1756	107	39.2	87.6
91	103240	2089	183	41.2	96.0
92	106234	1975	229	42.4	93.6
93	102189	1939	314	40.8	75.6
94	100184	1944	39	40.0	73.6
95	124221	1431	90	49.6	88.4
96	129218	1413	55	51.6	83.2
97	129209	1934	46	51.6	83.6
98	141223	1304	55	56.4	89.2
99	139232	1346	160	55.6	92.8
100	147238	1242	41	58.8	95.2
101	150054	1287	30	60.0	21.6
102	161199	1685	355	64.4	79.6
103	171205	1470	346	68.4	82.0
104	189232	1462	358	75.6	92.8
105	196196	1159	268	78.4	78.4

106	217201	994	253	86.8	80.4
107	231186	1106	408	92.4	74.4
108	131036	2583	279	52.4	14.4
109	137042	2392	241	54.8	16.8
110	168034	2152	168	67.2	13.6
111	176041	2333	427	70.4	16.4
112	194035	1874	287	77.6	14.0
113	199048	1467	116	79.6	19.2
114	154048	3067	457	61.6	19.2
115	158049	2783	274	63.2	19.6
116	171052	2416	523	68.8	20.8
117	167063	2501	524	66.8	25.2
118	172056	2035	216	68.8	22.4
119	189051	1874	229	75.6	20.4
120	192052	1721	116	76.8	20.8
121	202039	1878	430	80.8	15.6
122	200052	1531	183	80.0	20.8
123	209044	1830	427	83.6	17.6
124	204063	1973	515	81.6	25.2
125	207061	1924	457	82.8	24.4