PERIGLACIAL LANDFORMS AND ENVIRONMENTS

ON MOUNTAINS

IN THE NORTHERN HIGHLANDS OF SCOTLAND

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In accordance with the University of Edinburgh postgraduate study regulation 2.4.15, the following declaration is made:

This thesis was composed by me and is based on my own research.

January 1981.

ABSTRACT

This study seeks to establish the characteristics, age and genesis of periglacial landforms and deposits on mountains in the Northern Highlands of Scotland, and the nature of the environmental conditions under which these features were formed. The study concentrates on investigation of the periglacial features on three mountain massifs, those of An Teallach, the Fannich Mountains and Ben Wyvis.

A comprehensive classification of upland periglacial features is presented. This classification was employed in detailed mapping of the study areas at scales of 1:10560 and 1:10000. These maps allowed analysis of the distributional characteristics of each type of feature in terms of possible controlling variables. Different classes of landform were also surveyed in the field with a view to relating morphology to environment, and the structure of each type was investigated through trenching and sedimentological analysis of constituent material. Measurements were made of present geomorphological activity, and the nature of the present climatic environment was established through meteorological observations on An Teallach.

The results of the study indicate that almost all periglacial features were either formed during the Lateglacial period and have long been inactive, or formed during the Flandrian and are active at present. The severe conditions of the Lateglacial cold periods (the period of ice-sheet downwastage and the Loch Lomond Stadial) favoured large-scale frost weathering and the formation of a mantle of shattered detritus on plateaux and on slopes with gradients less than 40°, and the rapid accumulation of avalanche-modified talus at the foot of steeper slopes. Even the coarsest detritus was subject to downslope creep, forming sheets and lobes of boulders and debris. Lateglacial frost-sorting produced large-scale patterned ground, and nonsorted patterned ground may have survived in the form of hummocks and relief stripes. Small nivation benches formed in some areas.

The severity of the present "maritime" periglacial climate on these mountains is related to high precipitation and strong winds rather than extreme cold. Frost weathering is restricted to granular disintegration and flaking, although rockfall is not infrequent from glacially-steepened rockwalls. Debris-mantled slopes are subject to surficial frost creep and solifluction, the former combining with wind action to produce at least three different types of turf-banked terrace. Wind action is also responsible for the formation of deflation surfaces and wind stripes, and for the accumulation of niveo-aeolian sand deposits on lee slopes, the last mentioned being subject to active snowpatch erosion. Frost sorting under present conditions is capable of forming sorted circles and stripes that are typically 30-40 cm in width. Comparison of measured rates of mass-transport suggests that fluvial activity is presently the most effective form of denudation, followed by rockfall, and that the various forms of mass-transport active on high ground operate at rates comparable with (and sometimes greater than) those of similar forms of activity in more severe periglacial environments.

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CHAPTER 1 INTRODUCTION

1.1 Definition and scope

The term "periglacial" was introduced by Lozinski (1909, 1912) to describe the climate and the products of frost weathering in areas peripheral to the limits of Pleistocene glaciation and has probably proved more contentious than any other in the science of geomorphology. Many have urged its abandonment and replacement (e.g. Bryan, 1946; Capello, 1962; Shchukin, 1963; Linton, 1969) whilst others have used it in widely varying ways (Dylik, 1962, 1964 a,b; Dylikowa, 1962). Soviet authors in particular have been reluctant to employ the term (Jahn, 1975), but in Europe and North America its acceptance and scope have widened: Washburn (1973, p2), for example, considered "periglacial" to designate "...cold-climate, primarily terrestrial, nonglacial processes and features regardless of date or proximity to glaciers." It is in this extended sense that the term is employed here. "Periglacial environments", past or present, are considered to be those in which cold-climate nonglacial processes have operated to produce distinctive landforms and deposits, and the term "periglaciation" is employed to describe the operation of such processes.

Periglaciation has affected all of Great Britain, but its influence on the present landscape has been extremely variable, a major factor being that it has operated for much longer in some areas than in others. Paradoxically, it is in southern England (beyond the limits of Pleistocene glaciation) that periglacial processes have acted longest (hundreds of thousands of years). Elsewhere in Great Britain the effectiveness of periglaciation has been constrained (i) by the time that has elapsed since final deglaciation at any given location, and (ii) by the proportion of this time during which the climate was sufficiently cold to cause periglacial activity. The areas of Great Britain in which periglaciation has had least time to modify the landscape are therefore the final refuges of the Late Devensian ice sheet: the mountains of Highland Scotland. It is towards an understanding of the effects of periglaciation in these mountains that this dissertation is directed.

1.2 Aims and Methodology

Stated concisely, the aim of this study is to establish the

characteristics of past and present periglacial landforms, deposits, processes and environments on high ground in a sample area of the Scottish Highlands. In detail this incorporates the following.

(i) Establishment of the range of periglacial phenomena that occur on high ground in Scotland and construction of an appropriate classification of these phenomena.

(ii) Determination of the distribution of periglacial features and analysis of this distribution in an attempt to identify critical environmental controls.

(iii) Assessment of the age of periglacial landforms and deposits.

(iv) Establishment of the nature of past and present periglacial processes, the conditions under which such processes have operated, and rates of past and present periglacial activity.

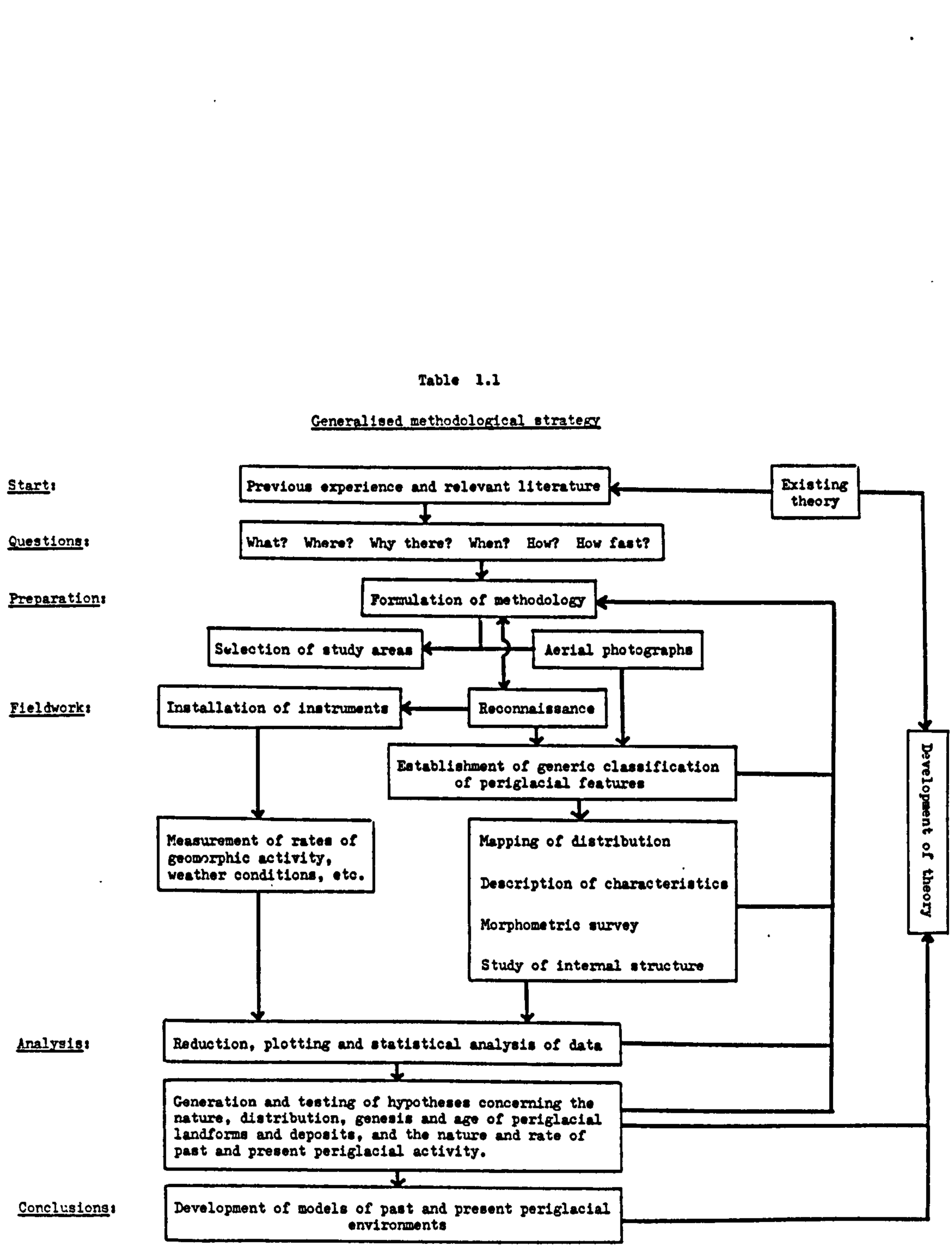
The above aims can be phrased as a series of questions (what? where? why there? when? how? how fast?) that dictated the a priori methodological strategy employed (table 1.1). Although the details of this strategy were modified as the study progressed, the framework depicted on table 1.1 remained basically unaltered.

The principal criterion employed in the selection of suitable study areas was that these should contain a wide variety of periglacial features within a manageable and reasonably accessible area. After consulting relevant literature and aerial photographs it was decided to concentrate the study on the mountain massifs of An Teallach ($57^{o}49'$ N., $5^{o}16'$ W.), Ben Wyvis ($57^{o}41'$ N., $4^{o}35'$ W.) and the intervening Fannich Mountains ($57^{o}42'$ N., $5^{o}02'$ W.) (figure 1.1; maps 1a, 1b and 1c). However, although these areas contain a very wide range of periglacial features, several types known to occur elsewhere in Scotland are absent, and it was necessary to extend the investigation to other areas (notably Rhum, the Cairngorms and Tinto Hill; figure 1.1) in order to construct a comprehensive classification of periglacial features for upland Scotland. Thus although this study concentrates on the investigation of periglacial phenomena on An Teallach, the Fannichs and Ben Wyvis, use is made of observations carried out in other parts of Scotland, and indeed elsewhere (mainly the Canadian Arctic Archipelago and the Jotunheimen Massif, Norway).

In order to assess present environmental conditions and rates of geomorphic activity in the study areas, various recording instruments and mass-movement measurement sites were installed early in the investigation (April-September, 1976). Although the measurement of

Table 1.1

Generalised methodological strategy



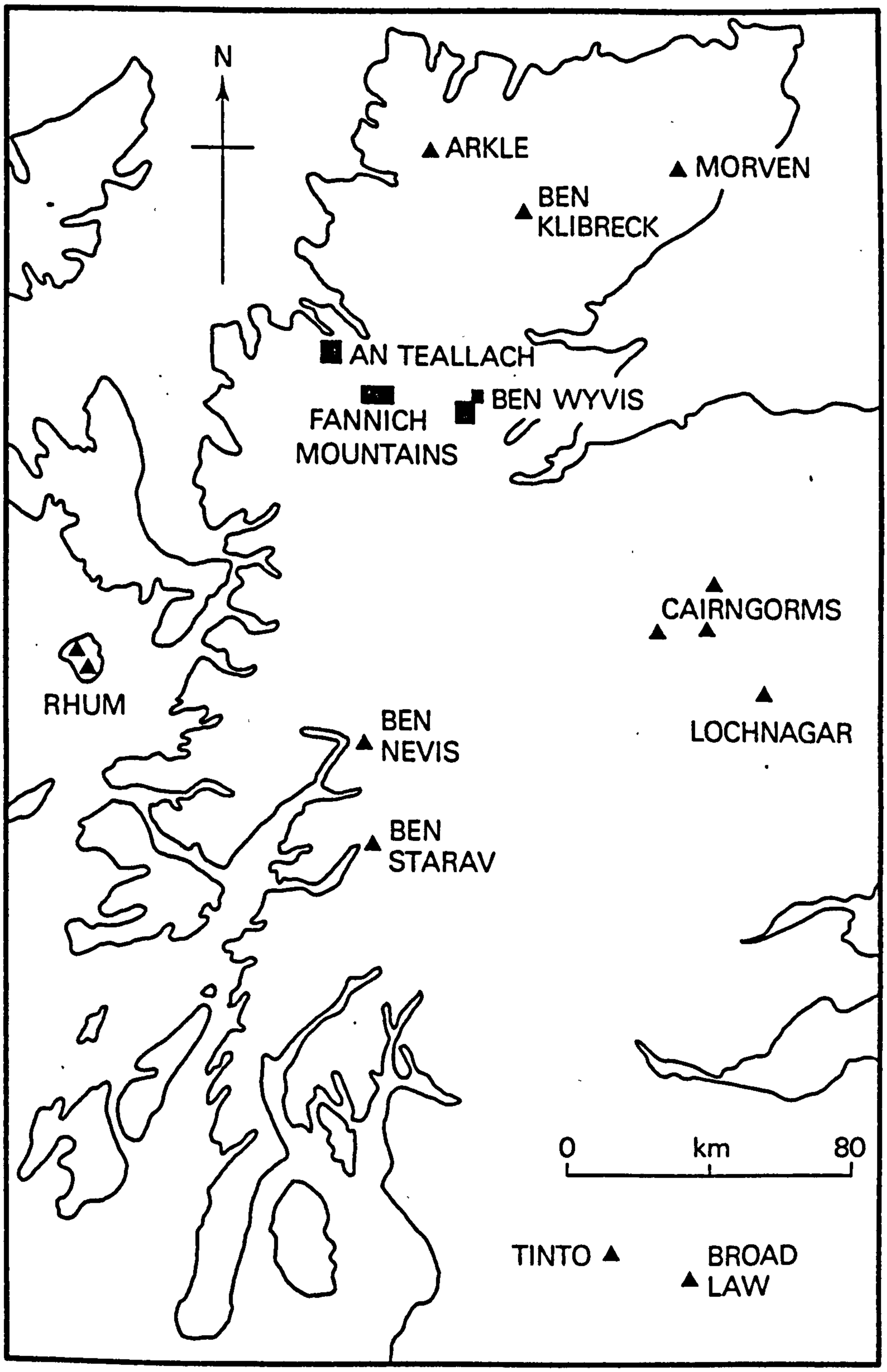


Figure 1.1: The three study areas (black rectangles) and some other mountain areas that receive frequent mention in the text.

mass-movement was carried out in all three areas, meteorological observations were restricted to An Teallach, which was visited at four to six week intervals from April 1976 to June 1979 in order to maintain the relevant instruments. Thus although most of the field-work was carried out during the summer months (May - September) observations on present conditions and activity were made at all times of the year.

Apart from the installation of instruments and mass-movement sites, early fieldwork was devoted to the formulation of a comprehensive classification of periglacial features, without which mapping and further analysis would have been impossible. The classification that was devised was of necessity generic rather than genetic, being based on morphology and other visible characteristics rather than structure and underlying processes, as these were largely unknown at the start of the study. Once a workable classification had been devised, detailed mapping of all three areas was carried out at scales of 1:10000 or 1:10560; the mountains of Rhum were also subsequently mapped at the same scale, as these contain features that are unrepresented in the original three study areas. Different classes of landform were then surveyed in the field with a view to relating morphology to environment, and the structure of each type was investigated through the digging of pits and sedimentological analysis of constituent material. Throughout the period of investigation hypotheses were generated concerning the age and origin of different features, and the original strategy was continually modified (at least in detail) to allow testing of such hypotheses.

1.3 Structure of the thesis

The text of this thesis has been ordered in such a fashion that information prerequisite to the discussion in any chapter has been introduced in previous sections. Interrelationships abound, however, and a certain amount of cross-referencing and restatement has proved necessary. The text falls into four broad sections:

(i) The background of the study (chapters 2 and 3). This section, based largely on published literature, sets the scene for subsequent discussion by outlining the nature of changes in environmental conditions on British mountains from the time when the last ice sheet reached its maximal extent to the present (chapter 2), and briefly

summarizes the history and principal findings of previous studies of periglacial phenomena on the mountains of Scotland, northern England and Wales, hereafter referred to as "upland Britain" (chapter 3). An extended review of the literature on periglaciation is omitted, as previous work pertinent to individual topics is reviewed in the appropriate chapters.

(ii) Environmental conditions and periglacial weathering in the study areas (chapters 4 - 6). These chapters are designed to provide a detailed background to the ensuing discussion of periglacial landforms and deposits. The relevant physical characteristics of the study areas(relief, geology, glacial geomorphology, soils and vegetation) are described in chapter 4. Chapter 5 presents the results of detailed meteorological observations made on An Teallach between 1976 and 1979. In chapter 6 the characteristics of periglacial weathering in the study areasare discussed.

(iii) Periglacial landforms and deposits (chapters 7 - 12). This section constitutes the core of the thesis. In chapter 7 the classification of periglacial features is introduced, the terminology employed is defined and the geomorphological maps are presented. Chapter 8 is devoted to a discussion of the characteristics, age and origin of the weathered detritus that mantles slopes and plateaux, and the following three chapters deal with groups of features that have developed on this detritus cover: mass-movement features (chapter 9), other small-scale features, mainly patterned ground (chapter 10) and landforms and deposits produced by wind action (chapter 11). The final chapter (12) in this section discusses the geomorphological role of snow and running water in the study area.

(iv) Synthesis and conclusions (chapters 13 and 14). This final section seeks to draw together the findings presented in previous chapters. Chapter 13 is devoted to an assessment of the relative effectiveness of different types of geomorphological activity operating at present on the mountains of Scotland, and the comparison of current rates of activity with those measured in other periglacial environments. Chapter 14 synthesizes earlier conclusions in the form of two models of periglaciation (Lateglacial and present-day) and provides a discussion of some of the principal themes that emerged from the study.

CHAPTER 2 PAST AND PRESENT ENVIRONMENTAL CONDITIONS IN UPLAND BRITAIN

2.1 Introduction

A recent reconstruction of the maximal thickness and extent of the last (Late Devensian) ice-sheet in the British Isles (Boulton et al., 1977) indicated that all ground over 600 m in altitude (with the exception of mountains in S.W. Ireland) was ice-covered. This is borne out for Scotland by reports of striae and erratics on high ground (Sissons, 1967a, 1976a). The final deglacierization of most areas above 600 m occurred during the downwastage of this ice sheet, and since then such areas have been exposed to subaerial weathering, erosion, transportation and deposition under periglacial conditions of very variable intensity. The timescale for the periglaciation of upland Britain is therefore (in geological terms) very short, as the Late Devensian ice-sheet is believed to have reached its maximal extent sometime after 18,500 B.P. (Penny et al., 1969). In upland areas that were covered by plateau ice caps during the subsequent Loch Lomond Advance (e.g. the Gaick Plateau of the central Grampians and parts of the western and S.E. Grampians) the timescale is even shorter, probably not exceeding 10,500 years.

The last 18,000 years, however, have witnessed drastic climatic changes. This chapter outlines the pattern and chronology of these changes and their environmental ramifications with reference to upland Britain in general and the Scottish Highlands in particular, thus providing a framework within which the age and significance of periglacial phenomena may be discussed.

By international consensus (Mitchell et al., 1973; Mangerud et al., 1974) the Devensian (Weichselian)/Flandrian boundary has been placed at 10,000 B.P. (radiocarbon years). The period under consideration therefore incorporates the last 8,000 years of the Devensian (18,000-10,000 B.P.) and the entire Flandrian (10,000 B.P. - present). In the discussion that follows, these periods have, for convenience, been divided into smaller units (figure 2.1). The Flandrian has been subdivided according to

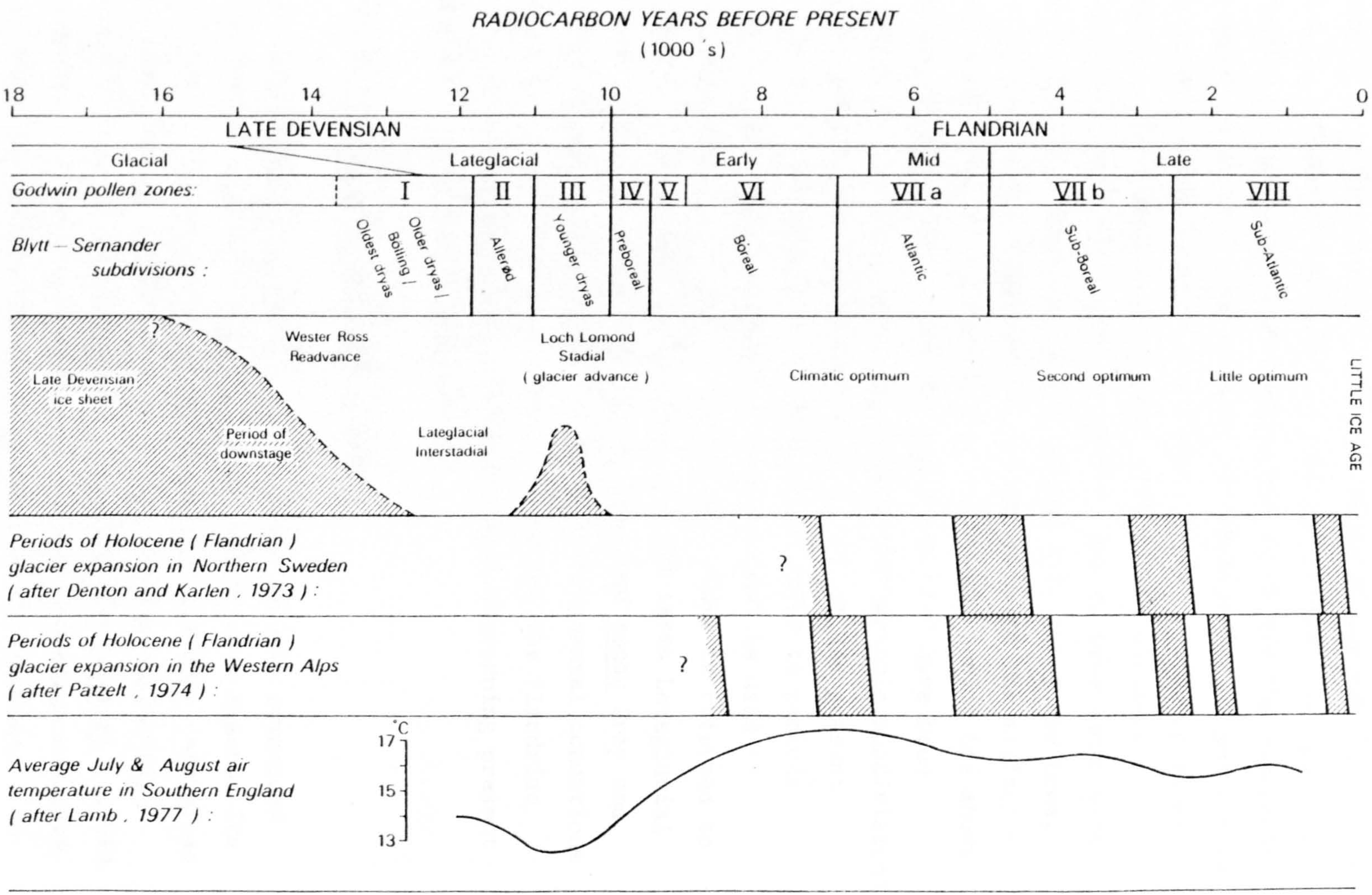


Figure 2.1: The last 18,000 years: chronological subdivisions and environmental change.

Blytt and Sernander's system, for although this offers only a broad correspondence with palynological evidence in Great Britain (Pennington, 1974) it is still widely used in this country to provide a framework for the discussion of environmental change. For present purposes the final 8,000 years of the Devensian may be subdivided into four periods: (i) a period of unknown duration when the Late Devensian ice-sheet covered all or most high ground; (ii) a period of slow ice-sheet downwastage which saw the emergence of high ground from under the ice and the commencement of periglacial activity under severe climatic conditions; (iii) the Lateglacial Interstadial, considered to have begun with marked climatic amelioration between 13,000 and 14,000 B.P. and to have ended with a return to glacial conditions around 11,000 B.P. (Gray and Lowe, 1977); (iv) the Loch Lomond Stadial, which occupied the millenium before the beginning of the Flandrian. Obviously period (i) above is of little concern here. Periods (iii) and (iv) have been recognised (Gray and Lowe, 1977) as climatostratigraphic subdivisions of a longer period, the Lateglacial period, but in the present context the term "Lateglacial" is employed to refer to periods (ii) to (iv) above, and "Lateglacial cold periods" is used to refer to periods (ii) and (iv). Period (ii) alone is referred to as the period of ice-sheet downwastage, and the terms Lateglacial Interstadial and Loch Lomond Stadial are employed sensu Gray and Lowe. The following discussion considers environmental conditions and changes in terms of each of these periods and the Flandrian period, and concludes with some generalizations concerning present environmental conditions on high ground.

2.2 The period of ice-sheet downwastage

The Late Devensian glaciation is believed to have commenced around 26,000 B.P. (Mitchell et al., 1973) and to have reached its maximal extent a few centuries after 18,500 B.P., a date determined by radiocarbon-dating of mosses in silts underlying supposedly Late Devensian till at Dimlington in Yorkshire (Penny et al., 1969). It is probable that initial downwastage of this ice-sheet resulted from reduction in snowfall rather than temperature increase. On the evidence supplied by assemblages of coleoptera (Bishop and Coope,

1977), climatic amelioration was not marked in Scotland until around 13,000 B.P., yet radiocarbon dates obtained from sediments from sites in the heart of the Scottish Highlands suggest than many Highland glens were deglaciated by, or shortly after, 13,000 B.P. (Kirk and Godwin, 1963; Sissons and Walker, 1974; Pennington, 1975; Lowe and Walker, 1977; Vasari, 1977). It therefore seems likely that considerable downwastage took place before 13,000 B.P., thereby exposing high ground to periglacial activity under a severe climatic regime.

Several readvances of the downwasting ice-sheet have been postulated. In Scotland, most of these have been challenged and successfully refuted (Sissons, 1974a, 1976a; Gray and Lowe, 1977) and the evidence for postulated readvances in northern England (e.g. Greswell, 1967; King, 1976) is questionable. However, it is difficult to avoid the conclusion that a marked drop in the marine limit in the Forth and Tay valleys and elsewhere (Sissons, 1976a) represents at least a halt in the retreat of the ice sheet, and possibly a minor readvance, although evidence for the latter is equivocal (Gray and Lowe, 1977). Positive evidence for a readvance, the Wester Ross Readvance, has been found in the N.W. Highlands, where end and lateral moraines delimit the former presence of a lobe of ice 25 km in width across Lochs Gairloch and Ewe and the surrounding low ground (Robinson and Ballantyne, 1979). In the area N.W. of Glen Torridon and west of An Teallach moraines marking the limit of this readvance show that high ground was ice-free at a time when glaciers in the surrounding valleys were advancing. Information about the wider extent, timing or climatic significance of this readvance is lacking, however.

2.3 The Lateglacial Interstadial

Although various authors (e.g. Peacock, 1970; Sugden, 1970a) have considered that glacier ice may have survived the Lateglacial Interstadial in Scotland, the evidence of the interstadial dates mentioned above, particularly those at Loch Etteridge (13,151 $\pm$ 390 B.P.; Sissons and Walker, 1974) and Loch Droma (12,810 $\pm$ 155 B.P.;

Kirk and Godwin, 1963) suggests that all Scotland was deglaciated early in the interstadial; Sissons (1976a) considered total deglaciation by 12,500 B.P. "a conservative suggestion". This certainly seems likely in view of the conclusion of Bishop and Coope (1977) (from the study of assemblages of coleoptera) that mean July temperature in S.W. Scotland was about 15°C (similar to that at present) around 13,000 B.P. From the composition of later assemblages at the same site they concluded that this represented the warmest part of the interstadial, being followed by a marked decline of mean July temperatures (to about 12°C) slightly before 12,000 B.P. with continued slow deterioration until around 11,000 B.P., when a further rapid fall initiated the glacial conditions of the Loch Lomond Stadial. If this pattern is representative of events throughout Scotland, then the highest mountain summits (such as that of Ben Nevis (1343 m), which currently experiences a mean annual temperature only fractionally above 0°C) must have experienced mean annual temperatures 2-3°C below freezing and hence may have supported permafrost and been subject to fairly intense periglaciation throughout the second half of the interstadial. Other workers, however, particularly palynologists (e.g. Birks, 1973; Walker and Lowe, 1977), have found that vegetation succession and soil development apparently progressed without marked interruption during the Lateglacial Interstadial until the onset of stadial conditions around 11,000 B.P. These differences may reflect the slower response of the flora to changing conditions (Coope, 1975, 1977), in which case the temperature curve of Bishop and Coope is a more accurate reflection of conditions during the interstadial. Palynological evidence has also indicated the existence of a short-lived climatic recession between 12,000 and 11,800 B.P. (the equivalent of the Older Dryas chronozone of Scandinavia) at a number of sites in Highland Scotland and northern England (Oldfield, 1960; Walker, 1966; Vasari and Vasari, 1968; Pennington et al., 1972; Pennington, 1975, 1977; Vasari, 1977). No such oscillation has been detected in studies of beetle assemblages (Coope, 1970, 1975, 1977; Coope and Brophy, 1972; Bishop and Coope, 1977) and indeed at other interstadial pollen sites (Birks, 1973; Lowe and Walker, 1977; Vasari, 1977),

so the question of an Older Dryas recession in upland Britain remains unresolved.

Palynological studies have yielded considerable information on interstadial vegetation. In contrast to the open herbaceous tundra vegetation of the preceding and ensuing periods, that of the latter half of the interstadial consisted of a closed cover of dwarf shrub heath and herbaceous grassland with birch copses in some areas. In the northern Highlands the dominant association was one of ericaceous (Empetrum) heath with variable representation of juniper (Pennington et al., 1972; Robinson, 1977). The nature of vegetation cover on high ground at this time has not been investigated.

2.4 The Loch Lomond Stadial

The deterioration in climate that marked the end of the Lateglacial Interstadial c. 11,000 B.P. resulted in the regeneration of glaciers in mountain areas throughout northern Great Britain, an event now generally termed the Loch Lomond Advance (rather than readvance, as Britain was apparently totally deglaciated during the preceding interstadial). The glaciers reached their maximal extent sometime after 11,800 - 11,300 B.P., this being the range of radiocarbon dates obtained from shells contained in marine sediments that were incorporated in or buried by Loch Lomond Advance end moraines (Sissons, 1967b; Peacock, 1971; Gray and Brooks, 1972). Dates obtained from organic sediments marking the return of warmer conditions show a wide range (Gray and Lowe, 1977, p. 177) but most indicate deglaciation by 10,500-10,000 B.P. Together, these groups of dates indicate that glaciers began to form before 11,000 B.P. and that deglaciation may have been complete well before 10,000 B.P. The former certainly seems likely if the pattern of thermal deterioration indicated by assemblages of coleoptera for the second half of the interstadial is representative (Coope et al., 1971; Bishop and Coope, 1977).

The extent of the Loch Lomond glaciers over much of Great Britain has been mapped in detail (e.g. Unwin, 1975; Sissons,

1976a, 1979a). In many areas, including the northern Highlands, Hebrides, Cairngorms, S.E. Grampians, Southern Uplands, Lake District and Snowdonia, glaciers were largely or entirely confined to corries and valleys, and most high ground was above the level of the ice and exposed to the operation of periglacial processes. Most summits and ridges rose above the large ice cap that occupied the western Grampians, and hence were similarly exposed; only in limited areas, such as the Gaick Plateau (Sissons, 1974b) was high ground almost entirely buried by ice.

In Scotland, northern England and Wales, ice wedge casts are found in Late Devensian glacial deposits that lay beyond the limit of the Loch Lomond Advance (Worsley, 1966; Sissons, 1974a; West, 1977). These imply that permafrost must have existed in such areas during the decay of the last ice sheet and/or during the Loch Lomond Stadial, thereby indicating that at these times mean annual temperature was at most -1°C near sea level, and possibly several degrees less (Péwé, 1966; Williams, 1975). The existence of ice-wedge casts immediately within the Loch Lomond Advance limit in Strath na Sealga (3 km south of An Teallach; Sissons, 1977a) and on Mull (J.S. Bibby, cited in Sissons, 1976a) proves that permafrost existed on low ground during the stadial, and indeed persisted for some time after the glaciers had started to retreat. Solifluction and slopewash were also very active on low ground at this time (partly because of the open nature of the grassland and heath vegetation) giving rise to rapid minerogenic sedimentation in depressions and hence a characteristic tripartite stratigraphy (organic/minerogenic/organic, corresponding approximately to interstadial/stadial/Flandrian deposition) in kettle holes and other enclosed sites. Peat overlain by solifluced till near Glasgow has been dated at 11,650-11,150 B.P. (Dickson et al., 1976), and Bishop and Coope (1977) dated the solifluction of gravels in S.W. Scotland to between 11,500 and 9,500 B.P.

Estimates of mean July temperature at sea level during the Loch Lomond Stadial show remarkable concordance. From coleoptera assemblages Coope et al. (1971) estimated 10°C for "lowland Britain" and Bishop and Coope (1977) put the figure at 8-9°C for

S.W. Scotland. From the analysis of firn line altitudes of former glaciers, mean July sea-level temperatures have been inferred for several mountain areas: 6°C for the S.E. Grampians (Sissons and Sutherland, 1976), 5°C for Mull (Gray and Lowe, 1977), 7.5°C for the Lake District (Manley, 1959; Sissons, 1980) and 7.0°C for the western Grampians (Sissons, 1979a). As the existence of permafrost at this time implies mean annual temperatures no greater than -1.0°C, mean January temperatures in the western Grampians are unlikely to have been higher than -9°C under full-stadial conditions (Sissons, 1979a).

Using these data, it is possible to reconstruct idealized curves of mean monthly temperature under full-stadial conditions for sea level and, assuming a drop in temperature of 0.6°C/100 m, for high ground (figure 2.2). The solid line on plots 1-3 in figure 2.2 indicates the maximum possible mean monthly temperatures (being based on a mean annual sea level temperature of -1.0°C); the dashed line indicates inferred temperatures based on a mean annual sea level temperature of -6°C, which Péwé (1966) considered maximal for ice-wedge formation. Curves 4-6 are based on data given in Lamb (1972). That for Ben Nevis relates to the period 1884-1903, when the summit observatory was operational.

A number of points emerge from examination of these curves. Comparison of (1) and (4) indicates that the range of mean monthly temperature during the Loch Lomond Stadial was much greater than at present (Sissons, 1979a). This is attributable to the oceanic polar front being much farther south than at present (Ruddiman et al., 1977) so that the moderating influence of the North Atlantic Drift was absent, and also (to a lesser extent) to probable freezing over of the North Sea in winter (Sissons and Sutherland, 1976). In fact the curve for mean monthly temperatures at sea level resembles that for present-day "alpine" environments (5) in which diurnal freeze-thaw cycles are of marked intensity during spring and autumn but the annual freezing period is neither intense nor of long duration. In this context much depends on the "true" values of stadial winter temperatures. On high ground (curves 2 and 3), however, the present-day thermal analogue is

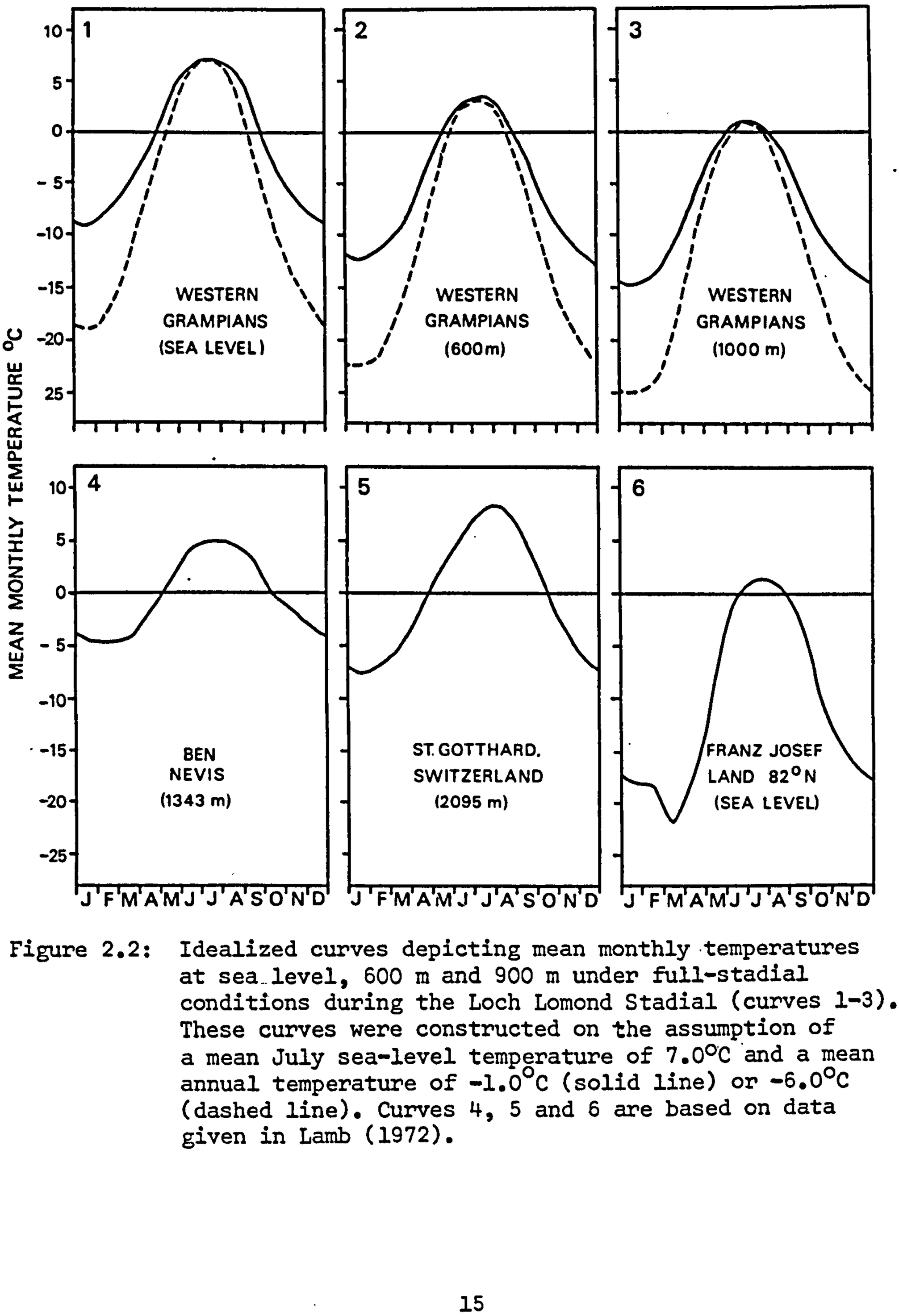


Figure 2.2: Idealized curves depicting mean monthly temperatures at sea level, 600 m and 900 m under full-stadial conditions during the Loch Lomond Stadial (curves 1-3). These curves were constructed on the assumption of a mean July sea-level temperature of 7.0°C and a mean annual temperature of -1.0°C (solid line) or -6.0°C (dashed line). Curves 4, 5 and 6 are based on data given in Lamb (1972).

high arctic (6) with an intense annual freezing period of long duration. Diurnal freeze-thaw cycles in present arctic environments are restricted to short periods during spring thaw and autumn freezeback (Fraser, 1959; Cook and Raiche, 1962a). Given the much higher insolation received in Highland Scotland, however, diurnal temperature variations are likely to have been large, and short-term freeze-thaw cycles probably operated for up to 6 months of the year at 600 m, although for shorter periods at higher altitudes.

From an analysis of firn line altitudes and glacier dimensions throughout the Scottish Highlands and the Lake District, Sissons (1979a) concluded that snowfall during the stadial was mainly associated with southerly airstreams preceding warm and occluded fronts, and that precipitation was high in the western Grampians and the area west of the Great Glen but low in the N.W. Highlands and very low (c. 600 mm y^{-1}) in the Cairngorms. This pattern he attributed to depressions following tracks farther south than those of to-day and to the freezing over of the North Sea in winter, the former being related to the southerly position of the oceanic polar front (Ruddiman et al., 1977). He also considered that snow blown from plateaux was important in nourishing glaciers in lee locations, which implies that a thick protective snow blanket is unlikely to have formed on exposed high ground.

All the above evidence strongly suggests that during the Loch Lomond Stadial high ground in Britain experienced a climate that was highly conducive to the operation of periglacial activity in all its manifestations: high winds stripped slopes and plateaux clear of protective snowcover, thereby exposing such areas to intense winter freezing; during the summer months freeze-thaw cycles were frequent; and the precipitation was sufficiently high to cause snowmelt floods and saturation of the ground during thaw. Periglacial activity would have been unmitigated by protective vegetation, as even on low ground there was only an open cover of grasses and moss heath (Gray and Lowe, 1977). The mountain environment of northern Britain during the Loch Lomond Stadial was clearly a periglacial environment par excellence.

2.5 The Flandrian Period

The change to warmer conditions at the end of the Loch Lomond Stadial was rapid. Minerogenic sedimentation in lakes and depressions was abruptly replaced by organic accumulation (e.g. Lowe and Walker, 1977) and by about 9,500 B.P. summer temperatures in S.W. Scotland were at least as warm as those at present (Bishop and Coope, 1977; Coope, 1977). Vegetation responded rapidly to climatic amelioration and by 9,800-9,700 B.P. juniper scrub had replaced the open tundra vegetation of the preceding stadial throughout most of the Scottish Highlands (Birks, 1977; Birks and Mathewes, 1978).

Temperatures continued to rise throughout the Boreal period, reaching a maximum during the climatic optimum of the early Atlantic period when mean annual temperature in England and Wales was 1.3-1.6^{o}C higher (and precipitation was 10-15% greater) than at present (Lamb et al., 1966). Subsequent climatic changes have included several periods of deterioration and amelioration (figure 2.1), but at no time in the Flandrian did the amount of change approach that of the Lateglacial period.

During the final 500 years of the Atlantic period (5,500-5,000 B.P.) there was a recession to cooler conditions followed by partial recovery and a second optimum between 4,500 and 3,500 B.P. The final Flandrian chronozone, the Sub-Atlantic (2,500/3,000 B.P. to the present), began with renewed thermal deterioration and by 2,500 B.P. mean annual temperatures in England and Wales were about 2^{o}C below those of the climatic optimum. In Europe north of the Alps there was a marked increase in wetness and storminess, and a climate of relatively mild winters but cool summers prevailed at this time (Lamb, 1977). The final two millenia of the Sub-Atlantic period witnessed a series of relatively short-term fluctuations in climate, although variations in mean annual temperature in England and Wales probably did not exceed 1.0^{o}C in amplitude. Temperatures reached a peak around 400 A.D. and during the "little optimum" of 1,100-1,300 A.D. then declined to a minimum during the 17th and 18th centuries, a period widely known as the Little Ice Age (even though there is no

evidence for the formation of glaciers in Great Britain at this time, or indeed during any part of the Flandrian).

During the Little Ice Age, average temperature in N.W. Europe were apparently lower than at any time since the early Flandrian, and many glaciers in the Alps and Norway advanced to limits beyond any achieved during the previous eight millenia (Lamb, 1977; Griffey and Matthews, 1978). Mean annual temperatures in England and Wales were 0.6°C lower than at present and in the Jotenheimen Massif (south-central Norway) the years 1700-1725 experienced average summer temperatures 1.6°C lower than now (Matthews, 1977). Colder water than at present lay off the north coast of Scotland, so it is likely that the decline in temperature was slightly greater in northern Scotland than in England; a mean annual temperature roughly 1.0°C lower than at present seems likely. Under such a regime mean annual temperatures on ground above 1,300 m would have been below 0°C.

One consequence of this depression in temperature was the establishment of perennial snowbeds on high ground. These were reported by contemporary travellers on Ben Nevis (1,343 m) and the Cairngorms (1,200-1,300 m) (Manley, 1949, 1971a; Sugden, 1971). Even Ben Wyvis (1,046 m) retained perennial snowpatches. In 1770 Thomas Pennant wrote of it that " ... snow lies in the form of a glaciere throughout the year", and in a gazetteer compiled a century later (Wilson, 1873) it is documented that Ben Wyvis had not been snow-free within living memory, apart from September 1826 (after an exceptionally warm summer). The survival of snow on Ben Wyvis during the second half of the nineteenth century is symptomatic of increased snowfall that accompanied thermal improvement after the coldest years of the Little Ice Age. These lay in the 17th and early 18th centuries, when precipitation was rather less (7% less in England and Wales) than at present (Lamb et al., 1966; Thom and Ledger, 1976), but reported snowfalls in lowland Scotland were most frequent towards the end of the 18th century and later (Pearson, 1976). Lamb (1977) considered that increased snowfall after the coldest conditions had passed was responsible for the relatively late readvances of some European glaciers.

Throughout the Flandrian a succession of woodland types developed in Scotland and northern England (Birks, 1977). The Scottish forests reached their maximum altitudinal development at the time of the Atlantic climatic optimum. The present potential tree-line in Scotland declines from about 750 m in the central Highlands to about 400 m in the N.W. Highlands (Birks, 1977). The highest Flandrian tree line was probably not much higher: Birks (1973) considered that much ground above 600 m on Skye had been treeless throughout the Flandrian; the highest tree stumps in blanket peat in the Cairngorms found by Pears (1975a) were at 793 m; and the highest tree remains found by the present author on An Teallach were below 500 m. The decline of some pine forests may have been initiated as early as 7,000 B.P. (Pears, 1975a; Birks, 1977) with deterioration in the base status of soils under the wet conditions of the climatic optimum, a change that also marked the initial formation of acid oligotrophic blanket peat in the western Highlands. Pine recolonized some peat bogs under the drier conditions of the Sub-Boreal (Steven and Carlisle, 1959), but further climatic deterioration at the beginning of the Sub-Atlantic period brought renewed bog formation. In general, the Mid- to Late-Flandrian decline in pine forests and concomitant peat formation was due to climatic and edaphic changes, rather than to burning and felling (Steven and Carlisle, 1959; O'Sullivan, 1977).

Peat growth rates at high altitudes in northern England and Scotland were highly variable. Around 600 m in the Pennines blanket peat presently reaches a depth of 4-5 m, the result of relatively rapid growth since the beginning of Atlantic times, but in the Cairngorms Pears (1975b) found peat depths of only 60-152 cm at altitudes between 610 m and 793 m, implying rather slow growth rates (1.4-3.0 cm/100y).

Above the limits of forest colonization and peat growth vegetation development followed a very different course (Pennington, 1974). The climax vegetation in this higher zone was high montane grassland, which Pearsall (1968) considered a natural development from Rhacomitrium heath. This grassland consisted of species associated with Lateglacial pollen spectra in Highland Britain,

such as Festuca ovina, Deschampsia flexuosa, Agrostis tenuis, Empetrum spp., Vaccinium spp. and Salix herbacea, suggesting an upwards migration of the lower limit of distribution of such plants in the early Flandrian, when climatic amelioration allowed their replacement by more successful competitors on low ground. In general it was only on the very highest ground (above 900 m in the central Highlands) that true arctic-alpine communities survived postglacial climatic amelioration. The Flandrian period therefore saw the development of the present three vegetational zones on high ground: a lowermost zone of peat, often containing the remains of vanished forests; a middle zone of montane grassland; and an upper zone of open vegetation including arctic-alpine species.

2.6 Present environmental conditions in upland Britain

From the description given earlier of climatic change in Great Britain during the last 10,000 years it is apparent that, whilst present mean annual temperature and precipitation fall slightly below average for the Flandrian period, they lie well within the extremes represented by the Atlantic optimum (warm and wet) and the Little Ice Age (cold and dry). Mean annual temperatures in England and Wales ranged from 1.3-1.6°C above present values to 0.6°C below present values; the equivalent figures for mean annual precipitation were + 10-15% to -7% (Lamb et al., 1966). During much of the Early- and Mid-Flandrian mean annual temperatures in upland Britain were higher than at present, and even during the Little Ice Age they did not sink to below 1°C lower than now. This pattern suggests that periglacial activity in upland Britain is likely to have been less pronounced than at present during much of the Flandrian, and that even during the coldest years of the Little Ice Age conditions were not markedly different from those now. During the Little Ice Age the altitudinal limit of certain types of periglacial activity, such as small-scale frost shattering, frost sorting and solifluction, was probably depressed by 100-150 m; rockfall, as in Norway (Grove, 1972), was probably more frequent; snowbeds more persistant; and annual depth of freezing on high ground must have been slightly

greater. On the other hand, although the very highest summits probably experienced mean annual temperatures below 0°C, it is unlikely (in view of the protection offered by snowcover in winter) that permafrost returned to upland Britain, or that there was more than a slight increase in the frequency, amplitude or intensity of freeze-thaw cycles. Periglaciation in the Flandrian apparently never approached the intensity of periglaciation during the Lateglacial cold periods.

Given the rather small differences in mean annual temperature between the Flandrian extremes and the present day, it is possible to regard the present periglacial environment of upland Britain as typical of that of the entire Flandrian period from the early Pre-Boreal on (whilst bearing in mind that periglacial activity was probably less during much of the Early- and Mid-Flandrian, and accelerated during cold phases within the Sub-Atlantic period). Two major environmental groupings therefore emerge: Lateglacial, typified by the severe conditions of the Loch Lomond Stadial; and Flandrian, typified by conditions at the present-day.

There is an unfortunate lack of data on the present upland climate of Great Britain (Taylor, 1976), and virtually none for the northern Highlands. The author's observations on An Teallach are reported in chapter 5 and the present discussion is restricted to generalizations concerning the principal features of the environment of upland Britain.

Regular meteorological observations were made from the observatory at the summit of Ben Nevis between 1884 and 1903 (Buchan et al., 1905-1910) and this remains the longest series available for any station above 600 m. At this, the highest point in Great Britain, the absence of extreme cold is remarkable; the lowest temperature recorded was -17.3°C. The lack of extreme cold is reflected in the shallow curve of mean monthly temperatures shown in figure 2.2. This is attributable to three factors: the moderating influence of the North Atlantic Drift, the reduction in lapse rate at low temperatures, and the flow of cold air off summit areas during calm (usually anticyclonic) conditions, giving marked temperature inversions. The present mean annual temperature

on Ben Nevis is fractionally above 0°C (Manley, 1971b), implying positive mean annual temperatures over all of upland Britain, with the possible exception of the Cairngorm summits. During the period of operation of the Ben Nevis observatory, when mean annual temperature was slightly (c. 0.3°C) lower than now, mean daily temperatures were below freezing for about 215 days each year (Omand, 1910) but the winter freezing period was liable to interruption in any month with the intrusion of tropical maritime air. Judged in terms of temperature alone, therefore, upland Britain is less "periglacial" than Southern Ontario or Moscow.

The severity of the climate at high altitudes in Great Britain is related not to extreme cold, but to moderate cold combined with high exposure and wetness. Westerly airstreams and frontal structures are forced to rise sharply on meeting the mountain barriers of Wales, the Lake District and the Western Highlands. Airflow is concentrated and accelerated, giving high wind speeds. Birse and Robertson (1970) have shown that mean wind velocity on mountains throughout Scotland generally exceeds 6 ms^{-1} at altitudes below 800 m, and 8 ms^{-1} at higher altitudes. Manley (1952) estimated the mean wind speed to be 10 ms^{-1} at the summit of Great Dun Fell (695 m) in the Pennines and 15 ms^{-1} at the summit of Ben Nevis where, during the period of observations, there was an average of 261 gales exceeding 22.7 ms^{-1} (50 m.p.h.) per year (Pearsall, 1968). Such averages conceal the occurrence of gusts of great strength. Even lee slopes may be affected by strong winds through the formation of strong frictional eddies under conditions of great turbulence.

The uplift of fronts over the western mountains also results in increased cloud cover and precipitation. The lower parts of cold fronts are deformed, and occluding depressions undergo marked structural changes resulting in heavy precipitation (Taylor, 1976). When potentially unstable Polar Maritime air rises over high ground instability is triggered by cooling, deep clouds form and again heavy precipitation results; with the passage of Tropical Maritime air the cloudbase lowers and there is lighter though more prolonged precipitation. Lee slopes experience less rain as such airmasses

descend, stability is regained and cloudbase rises. Polar Continental air crossing the North Sea brings low cloud, mist and rain (and heavy snow in winter) to the Cairngorms, south-east Grampians and Pennines, but the dominant pattern is one of declining precipitation from west to east. The highest precipitation (4,000+ mm y^{-1}) falls on mountains inland from Loch Nevis and in the Scafell area of the Lake District (Meteorological Office, 1977). Skye, Rhum and Mull average 3,200-3,400 mm y^{-1} and the Torridon Hills, Ben Nevis and Ben Cruachan all experience 3,600+ mm y^{-1}, but farther east precipitation is much less: around 2,000 mm y^{-1} for the Cairngorms, Monadhliath Mountains, Tweedsmuir Hills and Pennines. The driest massifs are those of Ben Wyvis and Lochnagar, with around 1,600-1,800 mm y^{-1}.

Related to this pattern is the pattern of cloudiness, which greatly reduces insolation over upland Britain. In the Western Highlands, for example, the annual value of received bright insolation is only about 20% of the potential value (Taylor, 1976). Snowcover, however, is related to altitude and latitude rather than longitude. Length of average snow-lie at any locality shows a fairly regular (linear) increase with altitude. For several mountain areas in Great Britain Manley (1971a) reported 50-79 days of snowcover at 450 m, 70-116 days at 600 m, and 105-180 days at 900 m, and he estimated a southward rise in the minimum altitude at which there is 100 days of snowcover from 600 m in the eastern Highlands to 750 m in the Pennines and 900 m in Snowdonia. Snowpatches in favoured locations may, however, survive for several years. Ben Nevis became snow-free for the first time for centuries in 1933 (Manley, 1971b) and snow "survivals" in the Cairngorms are common at present (Baird, 1957; Spink, 1970, 1975, 1976).

The severe climate of mountain areas in the British Isles exercises a rigorous control on the nature of upland vegetation. Trees and shrubs of normal stature cannot grow in exposed areas, and the "natural" or potential tree-line has been estimated to lie at only 610-685 m (Pears, 1967) or 750 m (Birks, 1977) in the Cairngorms and to decline with increasing exposure northwards and westwards. Plants that have colonised exposed summits and ridges in the Scottish Highlands are adjusted to survival under conditions

of high winds, heavy snowfall, long snow-lie and a growing season of 3-4 months or less. Tansley (1968) estimated that above 900 m 50% of the cover consists of hemicryptophytes (half-buried plants with persistent buds) and a further 25% consists of chamoiphytes (dwarf woody plants) such as Vaccinium or Empetrum species.

The nature of the vegetation cover (open or closed) and of component species is important in determining the distribution of different types of periglacial activity and landforms. It is convenient to consider the influence of vegetation on periglacial phenomena in terms of the three upland types identified earlier: the blanket bog peat formation, the montane grassland/heath formation, and the arctic-alpine formation (Eyre, 1963; Pennington, 1974).

The blanket peat formation is found up to altitudes of 900 m. In the Pennines even the summit plateaux are covered, but on the more exposed and steeper mountains of the Scottish Highlands it is often absent above 700 m. This formation is generally dominated by Sphagnum moss with heath plants such as Calluna vulgaris, Erica tetralix, Molinea caerula, Arctostaphylos alpinus and Myrica gale. There is a lack of periglacial features in such areas, largely because the peat cover inhibits current activity, except possibly solifluction (Galloway, 1958, 1961a) and gulley formation (Bower, 1960, 1961, 1962). Relict features are also rarely evident, as they have been buried by the peat. Pennington (1974) reported relict sorted polygons under peat in the Pennines, and Shaw (1977) believed Lateglacial boulder lobes to be buried under peat in the Lochnagar area.

The nature of the vegetation in the montane and arctic-alpine formations differs from area to area, depending amongst other factors on the base status of the soil, itself inversely related to the acidity of the parent rock. This accounts for the relative richness of species on the basic Dalradian rocks of the southern Highlands (Tansley, 1968). Here Festuca ovina and Alchemilla alpina are common on gentle slopes, and an association of Nardus stricta, xerophilous mosses and lichens occupies exposed cols and plateaux. Amongst the higher plants Silene acaulis, Loiseleuria procumbens and Carex bigelowii are common, together with shrubs

such as Empetrum nigrum and Vaccinium spp., and rare but typically arctic-alpine species such as Dryas octopetala and various species of Salix and Saxifraga. On base-poor soils in the Highlands, however, Calluna and Vaccinium communities and various grasses (Deschampsia flexuosa, Juncus trifidus, Nardus stricta) develop on slopes covered by thin blanket peat, and Rhacomitrium heath (Rhacomitrium lanuginosum and Carex bigelowii) covers wide tracts of plateau.

The vegetation types described above influence periglacial activity in several ways. First, cover is of prime importance. On exposed slopes or plateaux where vegetation cover is incomplete or sparse the ground surface is laid bare to the operation of deflation, wash, frost sorting, frost shattering and frost creep. Secondly, the tendency of vegetation to colonise the sheltered risers of mass-movement features, immobilising some parts of the slope but not others, leads to the formation of turf-banked terraces (chapter 9). Where vegetation cover is complete, the binding of the underlying regolith by plant roots (particularly those of woody shrubs) may inhibit mass-movement. Finally, vegetation cover affects surface microclimate, protecting the ground surface from wind, rain, insolation and frost, and aiding the retention of soil moisture and snowcover.

The role of vegetation in the present periglacial environment is therefore a negative one; in general vegetation cover tends to diminish periglacial activity. Were it not for the strong winds that inhibit vegetation growth and colonization, evidence for current periglaciation in upland Britain would be slight. The characteristic feature of the present periglacial environment is therefore not extreme cold, but extreme exposure, allied to high precipitation and cloudiness. No provision for an environment of this type has been made in classifications of periglacial types (e.g. Tricart, 1969; Washburn, 1979), and to describe it as "arctic" or "alpine" would be misleading in view of the mildness of winter temperatures. The dominant feature of the climate of upland Britain is the passage of maritime airmasses, so it is apposite that the present environment of mountain areas be characterised by the term "maritime periglacial".

CHAPTER 3 PREVIOUS RESEARCH

3.1 Introduction

The purpose of this chapter is to trace the development of research on the periglacial landforms and deposits of upland Britain. As the results of previous research are described in detail in later chapters, attention is here focussed on methodological developments and major findings with a view to establishing the current " state of the art" and thereby identifying major gaps in knowledge. The chapter is divided into three parts: section 3.2 traces the history of studies in this field; section 3.3 describes major developments in the study of major groups of features; and the concluding section suggests ways in which the understanding of upland periglacial phenomena may be furthered.

3.2 History

3.2.1 Early work

The earliest observations on upland periglacial phenomena in Great Britain were those made by officers of the Geological Survey in the early years of this century. Such observations, although brief, are often striking in conciseness of description and accuracy of inference. The best examples describe features in the N.W. Highlands and Caithness. In the former area, for example, Peach et al., (1912, 1913a, b) observed plateau frost debris " ... consisting of flat slabs and subangular fragments ... derived from disintegration of the rock ..." and mass-movement features in the form of " ... narrow subparallel terraces, separated from one another by small nearly-vertical scarps, often from one to three feet in height" (1912, p. 159-160). Crampton (1911) and Crampton and Carruthers (1914) made observations on the blockfields of the quartzite hills of Caithness, attributing this "thick mantle of frost debris" to Lateglacial frost shattering and noting terrace formation "due to irregularity in the downward creep of debris". Scree (talus) slopes were also envisaged to have accumulated mainly

in Lateglacial times (Harker, 1901) and occasional observations were made on other features produced by rapid mass-movement. Bailey and Maufe (1916), for example, described a striking landslide deposit near Glencoe, and Barrow and Cunningham Craig (1912, p. 119) provided an early description of debris flow following heavy rain.

The few accounts of upland periglacial phenomena published in the period 1915-1955 were principally concerned with small and apparently active patterned ground features. Stone polygons were reported by Gregory (1930) on Ben Lawers and the Merrick and by Simpson (1932) on Ben Iadain, Argyll, and active stone stripes were described by Hay (1936, 1943) in the Lake District and by Miller et al., (1954) on Tinto Hill in Lanarkshire. By far the most comprehensive account of upland periglacial phenomena to appear during this period was that published by Hollingworth (1934) describing several types of apparently active frost action and mass-movement features in the Lake District. These included small sorted polygons and stripes, turf-banked and vegetation-covered lobes, "islands" of upthrust stones in vegetated areas, small debris flows, stone streams and turf-banked terraces. Hollingworth compared these features with forms reported from arctic areas and concluded that the surface debris cover reflected widespread upfreezing of clasts and that downslope movement resulted principally from frost creep. "The stony debris", he wrote, "is a sea of moving material sweeping around islands of more or less stationary turf" (p. 177).

The final years of this period also witnessed the awakening of interest in upland tors. Fine examples of tors on the Cairngorms had been described by Hinxman (1896, 1915), and Linton (1949, 1955) noted further examples in the Ochils, Caithness and elsewhere. He proposed an ingenious "two cycle" explanation of such features, suggesting that they represented buried corestones, formed under conditions of deep chemical weathering, that had subsequently been exhumed, possibly under periglacial conditions.

3.2.2 The work of R.W. Galloway

When Fitzpatrick reviewed the existing literature on periglacial features in the British Isles in 1956 little was known about upland periglacial landforms. His 1958 paper "An introduction to the periglacial geomorphology of Scotland" provided only a brief description of upland features (solifluction sheets, mass-movement terraces, frost weathering, tors and small-scale patterned ground) as did the contemporary papers by Godard (1958, 1959). In contrast, a Ph.D. thesis entitled "Periglacial phenomena in Scotland" by R.W. Galloway (1958) represented a major contribution to the study of upland periglacial features. Apart possibly from Hollingworth's 1934 paper, this was the first study to consider the entire range of such features. This thesis and Galloway's subsequent papers (1961a, b) may be regarded as providing the foundation for subsequent studies of upland periglaciation.

Although detailed investigation of some of the sites examined by Galloway (such as Lochnagar and Ben Wyvis) has shown that his descriptions were sometimes inaccurate and his conclusions incorrect (Shaw, 1977; McMillan, 1978; chapters 9 and 10 below), there can be little doubt that his work both stimulated and strongly influenced later research. His map of periglacial features on Ben Wyvis, though misleading, has been reproduced in several texts (e.g. Sissons, 1964; Embleton and King, 1975; Curtis et al., 1976) and his papers have been widely cited in the international literature on periglaciation.

Galloway's principal contribution lay in documenting the range of upland periglacial features in Scotland and in emphasizing the widespread nature of such phenomena. He also recognised that many features were formed under severe climatic conditions in Lateglacial times. Into this category he placed frost-shattered detritus (blockfields), large fossil polygons and stripes, possible nivation and altiplanation features, periglacially-modified slopes and stone-banked lobes and terraces. The active features he described included amorphous cryoturbation (frost heave), patterned ground, ploughing boulders, solifluction lobes and terraces and stone streams. His work, however, suffers from drawbacks: his

essentially regional, ideographic approach precluded systematic classification of different types of feature and his explanations of underlying processes relied heavily on existing theory and subjective judgement rather than detailed measurement and hypothesis testing. Many of his generalizations must therefore be treated with caution. Even so, it would be wrong to belittle his achievement; his study was made at a time when periglacial theory was poorly developed, when knowledge of environmental change in Scotland was in its infancy and when there was little Scottish literature on the topic to guide his investigations.

3.2.3 Subsequent methodological developments

In Galloway's thesis most types of periglacial phenomena now known to occupy high ground in Great Britain were described and discussed. Later developments have therefore been of a largely methodological nature, with researchers applying increasingly sophisticated techniques to the study of upland periglacial phenomena.

Detailed description of individual periglacial phenomena and logical inference based on their characteristics constituted the most frequently-employed weapons in the methodological armoury of Galloway and earlier workers. Both are still widely and successfully used, although standards of both description and inference have become, in general, more rigorous. They have been employed, for example, in studies of tors (Palmer, 1956; Palmer and Radley, 1961; Palmer and Neilson, 1962), ploughing boulders (Tufnell, 1972), turf hummocks (Tufnell, 1975) and fossil rock glaciers (Sissons, 1975; Dawson, 1977). Other studies have made widespread use of these techniques within the context of areas of differing size, usually describing and discussing all periglacial phenomena within the study areas but concentrating attention on those that are best represented. Into this category fall several theses or dissertations, including those of King (1968; Cairngorms), Whyte, 0.970; Mamores) and Shaw (1977; S.E. Grampians) or chapters in studies concerned with wider aspects of the areas investigated

(e.g. Godard, 1965 (northern Highlands); Ryder, 1968 (Rhum); Birks, 1973 (Skye); Robinson, 1977 (Applecross)). Individual papers have described periglacial assemblages in generally smaller areas, such as the Rhinog Mountains in north Wales (Ball and Goodier, 1968; Goodier and Ball, 1969), Rhum (Ryder and McCann, 1971), Ronas Hill in Shetland (Ball and Goodier, 1974) and Ward Hill in Orkney (Goodier and Ball, 1975). Other accounts have attempted to synthesize published literature and individual observations to provide generalizations concerning the age, distribution and climatic or palaeoclimatic significance of periglacial features over wide areas (e.g. Tufnell, 1969; Sissons, 1976a).

The principal advantage of "area" studies is that they provide an opportunity for the researcher to establish relationships between each type of landform and both the "passive" elements of the environment (lithology, slope, aspect, altitude, regolith and vegetation cover) and other landforms, and to evolve a classification of local types. Such opportunities have rarely been fully exploited, however, and several of the above-noted studies tend to lapse into descriptive lists.

The lack of attempts to evolve an explicit classification scheme is surprising, as the existence of such a scheme would have helped to unify terminology (often highly idiosyncratic) and facilitate inter-area comparison of different types of feature. An outstanding exception is the scheme proposed by Ball and Goodier (1970) for frost-action features in Snowdonia, but this is insufficiently comprehensive to be of widespread application; the same authors employed completely different classification schemes in their accounts of features in the Northern Isles (Ball and Goodier, 1974; Goodier and Ball, 1975).

Equally surprising is the failure of researchers to make adequate use of the distributional characteristics of periglacial features in their work. There has been an almost total lack of accurate mapping of upland periglacial features. Such maps as have been published have been small in scale and grossly over-generalized (e.g. Galloway, 1961a; Kelletat, 1970a; Sugden,

1970b; Ryder and McCann, 1971) and even the larger-scale maps drawn by, for example, King (1968) and Ryder (1968, 1975) are rather inaccurate as well as cartographically inept. In part this deficiency may be attributed to the absence of an explicit classification scheme for upland periglacial features and to the lack of accurate large-scale (1:10,000) contour maps when many of these studies were carried out. The failure of most researchers to carry out detailed mapping has resulted in under-use of analysis of distribution as a methodological tool for relating periglacial forms to environment and for identifying associations of features. The potential value of detailed distribution studies has been illustrated in a series of papers (e.g. Sissons and Grant, 1972; Sissons et al., 1973; Sissons, 1977a, 1979b; Ballantyne and Wain-Hobson, 1980) that have indicated the absence of certain large-scale periglacial features (particularly boulder lobes) inside the Loch Lomond Advance limit, thereby indicating that such features have not been active since Lateglacial times. The value of attempts to identify the lower limits of present periglaciation across wide areas (Galloway, 1958, 1961b; Kelletat, 1970) is less certain, as such attempts have been based on very limited samples and, as Galloway observed, local factors such as slope, rock type, and vegetation removal may be of paramount importance in determining the lowest altitude at which features such as patterned ground or solifluction lobes are active.

Apart from surveys of talus slope profiles (Andrews, 1961; Statham, 1973, 1976a), morphometric techniques have rarely been successfully employed in the study of upland periglacial features. King (1968, 1971a, 1972) carried out width measurements on samples of lobes and patterned ground features in the Cairngorms in order to compare different types, but whilst such measurements add to the accuracy of King's descriptions, his statistical analyses of these figures are largely invalid. The studies of Shaw (1977) on boulder lobes and ploughing boulders in the S.E. Grampians rested heavily on morphometric analysis. Shaw tried to relate the dimensions of such features to variables such as altitude, aspect and slope. Although based on large samples, this attempt met with only limited success, probably because some critical variables

(such as depth of regolith, and depth to which ploughing blocks were embedded) remained undetermined, but also because the forms of statistical analysis employed were often inappropriate and not based on reasoned hypotheses.

Since frost action and mass-movement create deposits of distinctive composition and structure, it is not surprising that sedimentological studies of upland periglacial phenomena have enjoyed more success than most other approaches. Detailed investigations of slope deposits in the Southern Uplands (Tivy, 1962; Ragg and Bibby, 1966) indicated that the regolith had been produced by frost shattering of the underlying rocks and had been subject to vertical sorting and solifluction under severe periglacial conditions, and silt droplets discovered in upland soils (Romans et al., 1966; Romans and Robertson, 1974) have been interpreted as indicative of former permafrost conditions. Studies of sorting and particle orientation on talus slopes (Statham, 1973; 1976a) have indicated that these are now largely unaffected by snow or debris avalanching. Some periglacial deposits, moreover, exhibit stratigraphic sequences interpretable in terms of environmental change: Watson (1969), for example, related a solifluction/slopewash/solifluction sequence in Wales to changing conditions during the Lateglacial period.

Rates of current geomorphic activity on the mountains of Great Britain have been assessed in several ways. Miller et al., (1954) disturbed an area of sorted stripes on Tinto Hill and found perfect re-formation over two winters. Sequential observations by Tallis and Kershaw (1959) on small sorted polygons in Wales showed that such features could form in a single winter yet be destroyed by high winds and heavy rain during the following summer. Patterned ground destroyed by King (1968, 1971a) in the Cairngorms did not re-form, however, although individual stones showed measurable movement over the ground surface. The rate of downslope movement of clasts on striped slopes was measured over a single year by Caine (1962, 1963a) and shown to be relatively rapid (12-50 cm y^{-1}) and to increase with slope angle. This study was particularly valuable as Caine combined these observations with continuous measurements of ground surface temperature and daily precipitation

readings and thereby demonstrated that rapid movement was associated with abundant freeze-thaw cycles and periods of rain or snowmelt. Direct measurements of downslope movement have been made on ploughing blocks by Tufnell (1972, 1976) in northern England and Shaw (1977) on Lochnagar. The former found that individual blocks tend to display fairly consistent rates of movement of 0-8 cm y^{-1}, with almost negligible summer displacement; the latter reported much lower rates (generally < 0.5 cm y^{-1}).

Measurements of sediment transport have been few. The sediment yield resulting from mudflow and debris flow events has been measured directly by Statham (1976b) and indirectly by Prior and Douglas (1971), but the only assessments of overall "denudation rate" that have been obtained for upland Britain (11.8 and 12.8 m^3 km^{-2} y^{-1}) were measured for moorland areas at 300-500 m (Ledger et al., 1974) and no figures are available for higher altitudes. These figures were obtained by measuring sediment accumulation in reservoirs in S.E. Scotland over a known time period. Similar techniques were employed by Dawson (1977) and Sissons (1976b), who extrapolated rockwall retreat rates during the Loch Lomond Stadial from measurements of the volumes of a rock glacier in Jura and a protalus rampart in Wester Ross respectively.

Equally little attention has been devoted to dating the periglacial features of upland Britain. King (1968, 1971a, 1972) attributed formation of certain relict lobes and patterned ground features to the Little Ice Age on the basis of lichenometry. His sampling methods were not rigorous, however, and as he based his "dates" on lichen growth curves from Norway and failed to consider that maximum lichen size may simply reflect increased snowcover during the Little Ice Age (chapter 2) these findings must be treated with caution. The validity of the conclusions of Goodier and Ball (1969) regarding the Little Ice Age activity of certain features in the Rhinog Mountains in North Wales is also questionable, as the supposedly soliflucted "wall" upon which their conclusions rest closely resembles the riser of a detritus sheet of a type thought by others (e.g. Shaw, 1977) to be of Lateglacial age. The results of Shaw's interesting attempt to date the movement of ploughing blocks on Lochnagar using a form of dendrochronology (he

measured the number of growth rings on specimens of Calluna and Vaccinium at measured distances in the furrows upslope of four boulders) must also be dismissed as the distance/time curves calculated on the basis of this technique are highly improbable.

Radiocarbon dating has yielded more acceptable results. Dates of 4880 ± 135 B.P. and 2680 ± 120 B.P. were obtained from "vegetation layers" under "solifluction lobes" in the Cairngorms by O'Brien (Sugden, 1970b, 1971), but some of the value of these dates is lost through Sugden's failure to specify the location of the sites or to describe the type of lobes involved. In contrast, the "solifluction terrace" on Ben Arkle (Sutherland) under which Mottershead (1978) obtained dates of 3984 ± 50 B.P., 4734 ± 45 B.P., 5195 ± 55 B.P. and 5441 ± 55 B.P. was excellently described in an earlier paper (White and Mottershead, 1972). This work demonstrated that solifluction sensu stricto has been active at least intermittently during the second half of the Flandrian.

3.3 Summary of progress

3.3.1 Weathering, mountain-top detritus and tors

Most early accounts of mountain-top detritus in Great Britain described it as the product of frost-shattering in Late-glacial times (Crampton, 1911; Peach et al., 1912, 1913a, b; Crampton and Carruthers, 1914; Bailey and Maufe, 1916; Marr, 1916) and Galloway (1958, 1961b) reached a similar conclusion. Others saw frost action as operating on certain rocks at present (Thompson, 1950; Fitzpatrick, 1958; Godard, 1958, 1959, 1965; King, 1968; Hills, 1969; Ryder, 1968). No evidence has been presented to substantiate either view, although Sissons (1967, 1977b) pointed out a marked contrast in degree of frost weathering between adjacent areas inside and outside the Loch Lomond Advance limit, which suggests that most frost shattering occurred in Lateglacial times. Ragg and Bibby (1966) concluded that both coarse and fine material in periglacial deposits in southern Scotland had been produced by frost shattering under conditions

more severe than those at present, and through studies of boulder roundness Shaw (1977) demonstrated that small-scale granular disintegration had taken place during the Loch Lomond Stadial. Apart from the experimental studies of Potts (1970) no attempt has been made to assess the effect of periglacial weathering on different lithologies. Similarly, apart from the work of Ragg and Bibby (1966) the sedimentological characteristics of mountain-top detritus have remained largely unexplored, although Whyte (1970) has described the influence of rock type on the nature of surficial detritus in the Mamores.

Linton's (1949, 1955) two-cycle theory of tor formation, although accepted by some authors (e.g. Waters, 1964) has been attacked by others. Palmer and Radley (1961) found no evidence of deep-weathered rock around Pennine tors and concluded that these had been formed through periglacial denudation in Lateglacial times. Palmer and Neilson (1962) concluded that the weathered granite on Dartmoor had formed through frost shattering rather than deep weathering and that the tors had been exposed by removal of weathered material by solifluction, views that found support in later studies by Eden and Green (1971) and Doornkamp (1974). The Cairngorm tors were studied in great detail by King (1968) who was, however, unable to reach a definite conclusion as to their origins. Sugden (1965) found evidence of ice-moulding on some Cairngorm tors and erratics in the vicinity of others, thereby demonstrating that Linton had been incorrect in interpreting tors as symptomatic of areas that had escaped glaciation. Sissons (1967, 1976a) suggested that the preservation of large tors in the eastern Cairngorms reflected their having been relatively sheltered from glacier ice moving from the south-west.

3.3.2 Detritus slopes

Many talus slopes in upland Britain are partly or entirely vegetation-covered, and several authors have concluded that such slopes were largely formed during the Lateglacial period (Galloway, 1958; Andrews, 1961; Ball, 1966; Ryder, 1968; Tufnell, 1969; Ryder and McCann, 1971). The formation of rock

glaciers and protalus ramparts during the Loch Lomond Stadial (see below) certainly suggests considerable rockfall activity at this time. Well-developed talus slopes have, however, been described for areas within the Loch Lomond Advance limits (Whyte, 1970; Statham, 1976a) and these must have formed since the disappearance of the last glaciers. Most authors have considered present talus accumulation to be slow, however (Thomson, 1950; Galloway, 1958; Godard, 1958; Ryder, 1968; Whyte, 1970; Ryder and McCann, 1971; Mathieson, 1977), and rockfall data collected by Shaw (1977) appears to support this view. Morphological and sedimentological studies of unvegetated talus (Davison, 1888a; Andrews, 1961; Statham, 1973, 1976a; Shaw, 1977) have shown that such slopes typically possess an upper straight section and basal concavity, maximum slope angles of around 36°, a degree of fall-sorting and downslope orientation of surface clasts, all features typical of unmodified rockfall talus. Kirkby and Statham (1975; Statham, 1976a) have attempted to explain these characteristics using a mathematical model based on the energy possessed by falling particles.

Little detailed work has been carried out on blockslopes and other detritus-mantled slopes. Several authors (Crampton, 1911; Crampton and Carruthers, 1914; Clark, 1962; Whyte, 1970) have suggested that blockslopes are undergoing downslope movement at present, but have supplied no supporting evidence. Information on detritus slope structure is largely restricted to that provided by Ragg and Bibby (1966) who interpreted the detritus cover on Broad Law (Southern Uplands) as the product of frost weathering associated with an annual freeze-thaw cycle under permafrost conditions and subject to solifluction. The stratigraphy of slope deposits in mid Wales has been interpreted by Watson (1969) as demonstrating the occurrence of solifluction during the period of ice-sheet retreat and during the Loch Lomond Stadial, with slopewash dominant in the intervening period. The alternating dominance of solifluction and slopewash during the Lateglacial period is also reflected in stratified screes in the Lake District (Boardman, 1977, 1978).

3.3.3 Mass-movement features

Lobes and terraces produced by slow downslope movement of detritus rank amongst the most widespread products of periglaciation in upland Britain. Galloway (1958, 1961a) described large lobes of Lateglacial age as occurring above 540-600 m in all parts of Scotland and considered that smaller active features are equally widespread though restricted to higher altitudes in the southern and eastern Highlands. Galloway also recognised that lobes represented the crenulate fronts of sheets of debris ("irregular terraces") and identified two types: large "stone-banked" lobes and smaller "turf-banked" (actually vegetation-covered) lobes. He ascribed formation of the former to the Lateglacial period but considered the latter type to be active at present. King (1968, 1972) argued that even the largest lobes in the Cairngorms had been formed in postglacial times, but this view was necessitated by his acceptance of Sugden's (1970a) interpretation of deglaciation in this area, an interpretation that was subsequently demonstrated incorrect (Sissons, 1979b). Detailed mapping of the limits of the Loch Lomond Advance glaciers in several parts of Scotland and in the Lake District (Sissons, 1976a, 1979a, 1980) showed that large (usually bouldery) lobes and terraces often occur immediately outside these limits but are absent in the areas formerly occupied by glacier ice. The Flandrian dates obtained from under "solifluction lobes" in the Cairngorms (Sugden, 1970b, 1971) and a small "solifluction terrace" on Ben Arkle (Mottershead, 1978) indicate that lobe formation has taken place in postglacial time, however, and this evidence supports Galloway's scheme of large Lateglacial lobes and smaller active lobes.

The mechanism responsible for the formation of such lobes is not understood. The Lateglacial lobes were considered by Galloway to be solifluction features immobilised through washing out of fines, but Shaw (1977) advanced the theory that they formed through "rock glacier creep". Flandrian lobes have been widely attributed to "solifluction", though the mechanics of this process have not been specified.

The term "terrace" has been used to refer to accumulations of solifluction detritus on valley floors (e.g. Crampton and Taylor, 1967), sheets of detritus with a straight rather than crenulate frontal scarp (e.g. Shaw, 1977) and small-scale, generally turf-banked features with treads that lie either parallel or oblique to contours. Features of the last type were described by early workers in the northern Highlands (Crampton, 1911; Peach et al., 1912, 1913a, b; Crampton and Carruthers, 1914) who considered them active and attributed their formation to some type of "earth creep". Hollingworth (1934) proposed that such features were formed through the retardation of downslope-creeping detritus behind vegetation bands, a conclusion also reached by Mottershead and White (1969). Galloway (1958, 1961a) attributed features of this type to active solifluction, however, and other researchers have attempted to relate terrace distribution to dominant wind direction, but have not specified the nature and cause of this relationship (King, 1971b; Ball and Goodier, 1974; Goodier and Ball, 1975).

Fossil rock glaciers and incipient rock glaciers have recently been reported from various localities. Sissons (1975) described a mass of detritus 1.2 km long at the foot of Beinn Alligin (Wester Ross) and attributed its disposition to the reactivation of glacier ice following its burial by boulders supplied by a massive rockfall. A similar though smaller example of this type of feature has been described by Robinson (1977). Talus foot rock glaciers in Jura, the Cairngorms and the Lake District have also been reported (Dawson, 1977; Sissons, 1979b, 1980). All of these features are thought to have formed during the Loch Lomond Stadial.

Like lobes and terraces, ploughing boulders are of widespread distribution (Hay, 1937, 1942; Galloway, 1958; Tivy, 1962; King, 1968; Goodier and Ball, 1969; Sugden, 1970b; Kelletat, 1970a). Tufnell (1972) studied 500 ploughing boulders in northern England and provided a comprehensive classification of their characteristics. Average rates of movement have been shown by Shaw (1977) and Tufnell (1976) to range from a few millimetres to a few centimetres per year. The causes of movement are still unknown, although possible mechanismswere discussed by Tufnell (1972).

Landforms produced by rapid mass-movement, such as slump features and debris chutes (debris flows or mudflows) are also common in Upland Britain. Observations in the Cairngorms (Baird and Lewis, 1957) and at lower altitudes (e.g. Prior and Douglas, 1971; Beven et al., 1978) have illustrated recent localised slope failures, in all cases following prolonged intense rain. Prior and Douglas (1971) and Statham (1976b) have assessed the amount of sediment transport associated with such activity.

3.3.4 Patterned ground

Both active and fossil patterned ground have been found at many upland localities in Great Britain. Small sorted polygons, circles and stripes have been described by many authors (Gregory, 1930; Simpson, 1932; Hollingworth, 1934; Fitzpatrick, 1958; King, 1968; Tufnell, 1969; Ball and Goodier, 1970; Kelletat, 1970a; Ryder and McCann, 1971; Birks, 1973) and shown to be capable of forming at present (Miller et al., 1954; Tallis and Kershaw, 1959). Hay (1936, 1943) and King (1971a) attributed patterned ground formation to the action of needle ice, but Caine (1962, 1963b) has shown that differential frost heave, small-scale mass-movement and rill wash are important in stripe genesis, and measured present downslope movement of up to 50 cm y^{-1} on stripes in the Lake District. Galloway (1958) stressed the importance of vegetation-free terrain in patterned ground development. In no instance have documented active patterned ground features exceeded a few decimetres in width, and the maximum width of sorted clasts has generally been reported as 10-15 cm.

Fossil sorted patterned ground has been less frequently reported. Fossil features display a wide size range, from less than 2 m width to over 15 m (Ryder, 1968; Tufnell, 1969). The sorted material is generally much coarser, and inactivity is indicated by vegetation cover in the centre of polygons and circles and on "fine" stripes. Features of this size have generally been attributed to the Lateglacial period, (e.g. Ryder and McCann, 1971) and have been considered symptomatic of former permafrost conditions (Williams, 1975; Ballantyne and Wain-Hobson,

1980). King's (1971a) conclusion that large-scale patterned ground in the Cairngorms formed during the Little Ice Age or is active at present was again conditioned by acceptance of Sugden's (1970a) incorrect interpretation of deglaciation in that area. Godard (1965) reported both active and Lateglacial patterned ground in the northern Highlands, but also described inactive polygons with widths of 0.4-1.0 m; these may be smaller Lateglacial features or may have formed during the Little Ice Age.

Nonsorted ("ridge and furrow") stripes and circles (turf hummocks or thufurs) on Ben Wyvis were considered by Galloway (1958, 1961a) to reflect an underlying sorted pattern. Similar stripes in north Wales were considered fossil by Goodier and Ball (1969) but Tufnell (1975) presented evidence that indicates that turf hummocks are capable of forming under present conditions in the northern Pennines. Birks (1973) described turf hummocks on Skye and attributed their formation to differential rates of ground freezing induced by "microtopographical" variations in soil type and vegetation cover.

3.3.5 The action of wind, snow and running water

In addition to playing a vital role in the formation of some of the landforms described above, wind, snow and running water have left their own characteristic traces on the periglacial landscape of upland Britain. King (1968, 1971b) described wind-patterned ground in the Cairngorms in the form of scars in the vegetation cover, typically 1 m wide and 2-4 m long, that he termed denuded surfaces. Kelletat (1970a) showed that such "deflations-formen" are widespread on the mountains of Scotland, taking the form of stripes eroded in the vegetation cover and patterns of vegetation such as clumps, bands and garlands. Wind-patterned ground covers the summit plateaux of hills in Orkney and Shetland, where Ball and Goodier (1974; Goodier and Ball, 1975) classified various types according to shape. The general explanation for these features, first proposed by King, is that they form through "Rasenabschalung' or turf exfoliation, a combination of needle ice erosion and stripping by wind, often leaving a lag surface of granules. Aeolian deposits, termed (wrongly) hill dunes have

also been described on the hills of the Northern Isles by Ball and Goodier, but the most impressive accumulations of aeolian sand are associated with lee slopes on mountains of Torridon Sandstone (Peach et al., 1913a; Godard, 1965; Sissons, 1976a; Ballantyne, 1977).

Nivation has been considered responsible for the formation of supposed cryoplanation features and nivation cirques in various mountain areas (Galloway, 1958; Watson, 1966; King, 1968) but these may be essentially structural or glacial features. The former existence of perennial snowpatches during the Loch Lomond Stadial is, however, marked by the survival of protalus ramparts, sometimes of remarkable size (Sissons, 1976b, 1979b, 1980). The occurrence of protalus ramparts of more recent (Little Ice Age?) origin on Ben Nevis was reported by Gatty (1906), and Tufnell (1971) has described present erosional activity associated with snowpatches in the northern Pennines.

The present role of running water in upland Britain has received little study. Various authors have, however, expressed the view that running water constituted a highly effective geomorphic agent during the Lateglacial cold periods and have attributed the formation of large alluvial fans along the western margin of the Pennines and the southern margin of the Ochils to accelerated fluvial activity at these times (Galloway, 1958; Sissons, 1976a; Worsley, 1977).

3.4 Conclusions

From the preceding summary it is possible to identify major gaps in present knowledge. This section outlines some of the more promising approaches that may be employed to close these gaps. Many of these approaches were employed in the present study.

Given the size of the literature on upland periglacial landforms in Great Britain, it seems unlikely that many undiscovered types of feature await description. However, students of periglaciation reading the literature on, for example, mass-movement features in upland Britain are likely to be confused by the profusion and

ambiguity of the terms employed. Some of these (e.g. "solifluction terrace") are used to describe features with markedly different characteristics. In other instances features of common origin have been denoted by several different terms. Systematic classification of features and standardization of terminology are therefore long overdue.

Classification is also a necessary step if meaningful and comparable large-scale maps of periglacial phenomena are to be drawn. This is in turn prerequisite to full utilization of distribution studies as a methodological tool for identifying the relationships between landforms and local environment (lithology, aspect, slope, altitude and glacial limits), threshold conditions of occurrence and associations of forms and deposits.

Sedimentological techniques have already been successfully applied to the study of periglacial landforms and deposits, but the potential of sedimentological analysis has not been realised. Studies of mountain-top detritus, mass-movement features and patterned ground would profit from detailed sedimentological investigations employing the methodological tools (e.g. quantitative particle size and shape analyses, clay mineral analysis, fabric analysis and measurement of cohesive strength and void space) that have been developed by sedimentologists and engineering geologists. Similarly, the quantitative analysis of morphometry may yet yield useful results in relating form to environment (and thereby, indirectly, to process). However, the sledgehammer approach of measuring every dimension of features and attempting to relate these measurements to site conditions may profitably be replaced by the measurement of selected variables in order to test reasoned hypotheses. In this context it is also desirable that researchers adopt a more rigorous approach to the use of inferential statistics.

There appear to be three main ways in which the age of periglacial phenomena in upland Britain may be assessed. First, comparison of the distribution of different types of features with the limits of the Loch Lomond Advance may be used to indicate which landforms and types of activity have not been active since the end of the Lateglacial period. Secondly, monitoring of present activity may be employed to indicate the scope of current

periglaciation. Thirdly, radiocarbon-dating of buried organic material may aid direct identification of periods of activity, although this method is likely to be restricted in application to the dating of slow mass-movement features and possibly aeolian accumulation, and is subject to certain difficulties of interpretation (e.g. Worsley and Harris, 1974).

Techniques for the measurement of rates of present mass-movement and sediment transport have reached a considerable degree of sophistication (e.g. Williams, 1957, 1962; Washburn, 1960; Anderson and Finlayson, 1975), yet apart from a few studies of surficial mass-movement such techniques have not been employed in upland Britain. Only Caine (1962, 1963a) has attempted to relate rates of activity to direct meteorological observations.

In summary, the range of periglacial phenomena in upland Britain has now been described in detail, and inferences (sometimes conflicting) have been made concerning the distribution, morphology, genesis and age of individual features. There is, however, considerable terminological confusion, and existing classifications are only local in scope. Explanation of process has often been based on surface appearance, qualitative description, subjective inference and analogy with findings elsewhere; the potential of more powerful techniques, such as rigorous analysis of distribution, sedimentology or morphology has rarely been realized. Data on rates of activity are limited. The formation of periglacial features not considered active at present has generally been assigned to the Late-glacial period, but the inactivity of certain supposedly Lateglacial features has been questioned and several studies have invoked formation of fossil features during the Little Ice Age.

Perhaps the greatest weakness of previous research is that it has concentrated on explaining the characteristics of individual features. As indicated above, there is much scope for extending knowledge in this way; indeed this constitutes the primary aim of this thesis. A second aim, no less important, is to attempt to demonstrate that many features hitherto considered in isolation may be related, and represent differing responses of the terrain to Lateglacial and present-day environmental conditions. The ultimate if unattainable

aim is perfect knowledge of the characteristics of these periglacial "systems" (Chapter 14). Explanation of the whole (albeit imperfect) is more meaningful than the summed understanding of its parts.

CHAPTER 4 THE STUDY AREAS

4.1 Introduction

Periglacial landforms and deposits are produced through the response of terrain to the operation of exogenous agents such as wind, rain, frost and temperature fluctuations under predominantly cold-climate conditions. Changes in the nature and intensity of such exogenous agents since the disappearance of the last ice-sheet were described in chapter 2, and the present climate of An Teallach will be discussed in chapter 5. This chapter briefly outlines other relevant characteristics of the environment of the study areas: relief, geology, glacial history, soils and vegetation. As will become apparent in subsequent chapters, these characteristics play an important role in determining the mode and effectiveness of periglacial activity and hence the nature of resulting landforms and deposits.

4.2 Relief

The three study areas are located on a curved transect that stretches from Little Loch Broom on the west coast to the Cromarty Firth in the east (figure 4.1). A substantial part of each area lies above 600 m, and the highest parts of each massif are over 1000 m in altitude (maps 1a, 1b, 1c). All three areas contain a great range of slopes, from flat or gently undulating plateau surfaces to corrie headwalls at angles of 35-70°. The northern part of An Teallach (map 1a) consists of gently-rolling plateau between 730 m and 860 m, but farther south steep rocky arêtes rise above spectacular corries and culminate in the pyramidal summits of Bidean a'Ghlas Thuill (1062 m) and Sgùrr Fiòna (1059 m). In contrast, an extensive undulating plateau constitutes the highest ground of Ben Wyvis, reaching an altitude of 1046 m at the summit (Glas Leathad Mór) and terminating eastwards in steep corrie headwalls (map 1c). Plateau surfaces are rarer in the Fannich Mountains, which consist of a central ridge running W.N.W.-E.S.E. with many offshoots. This ridge is interrupted by a col at 550 m (NH 174706; map 1b) that divides the massif into the Western Fannichs (A'Chailleach, 999 m; Sgùrr Breac, 1000 m) and the Eastern Fannichs, which culminate in

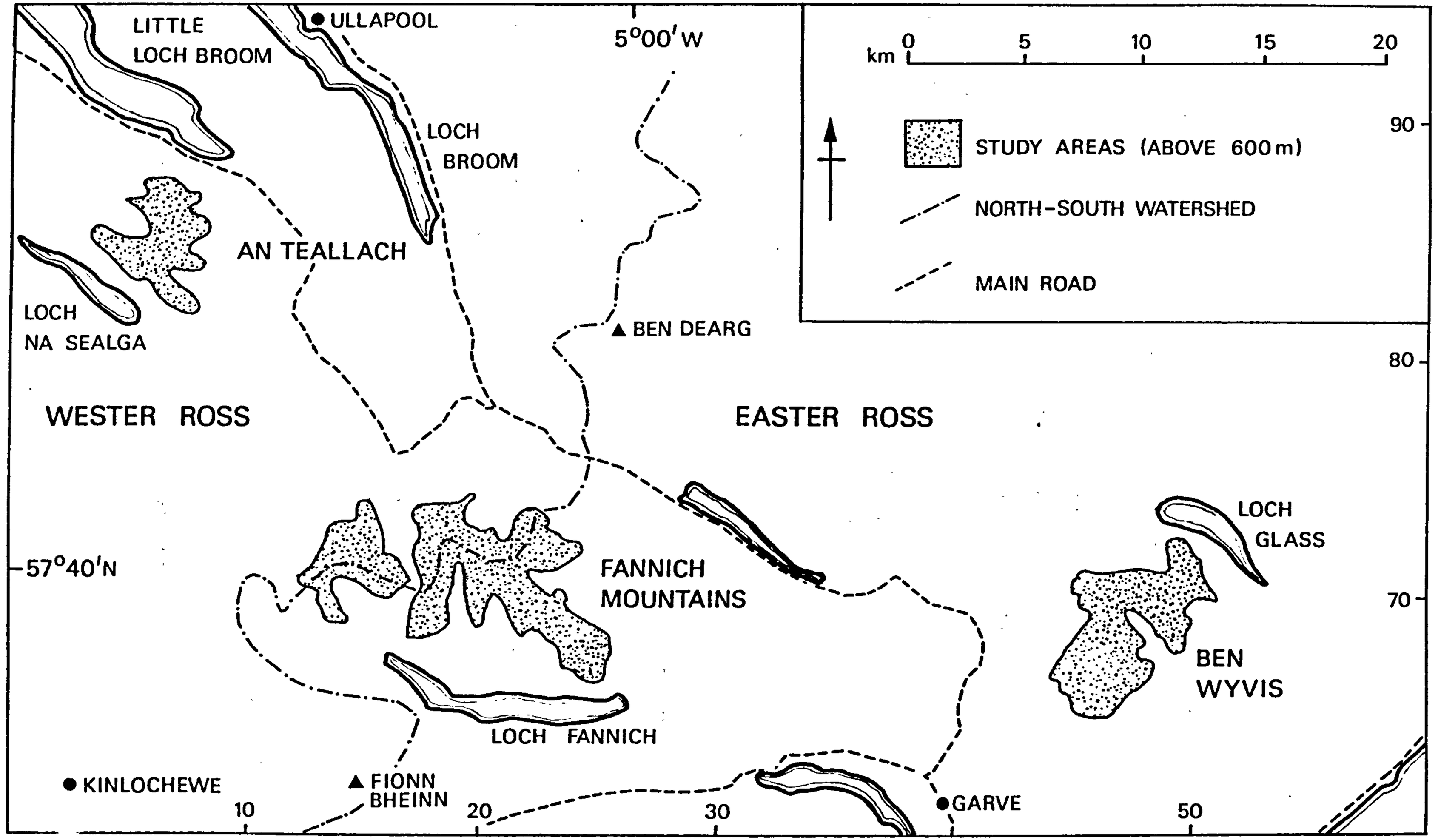


Figure 4.1: Location of the three study areas.

Sgùrr nan Clach Geala (1093 m) and the highest peak in the area, Sgùrr Mór (1110 m). The slopes of the main ridges are steep, but precipitous only above north- and east-facing corries.

4.3 Geology

The rocks of the study areas are members of four ancient systems. The oldest rocks belong to the Lewisian Gneiss Formation, which constitutes a fundamental basement underlying the other groups but within the study areas crops out only in the Fannich Mountains. Resting uncomformably on the Lewisian Gneiss are sedimentary rocks of the Torridonian system and metamorphics (predominantly metasediments) of the Moine Series. The Torridonian and Moinian rocks are sharply divided along the line of the Moine Thrust, which lies about three kilometres east of An Teallach (figure 4.2), placing this massif in the Torridonian zone and the other study areas in the Moinian zone. The youngest rocks in the area are of Cambrian age and crop out immediately west of the Moine Thrust, where they rest uncomformably on Torridon Sandstone.

4.3.1 An Teallach (figure 4.2)

The An Teallach massif is composed of rocks belonging to the Applecross formation of the Torridonian system (Peach et al., 1913a), except for unconformable outliers of Cambrian Quartzite on each of the three eastern spurs of the massif. The Torridonian rocks, referred to hereafter as Torridon Sandstones, consist of chocolate and red arkosic grits with occasional conglomeratic layers. The Torridonian Sandstones were deposited fluvially in conditions that permitted oxidation (Stewart, 1962) during the late Pre-Cambrian period, c. 1000-800 m.y. (Johnson, 1964). These rocks have suffered remarkably little subsequent alteration, being practically unfolded and dipping southeastwards at angles less than 15°. Current bedding is conspicuous, and the strata retain their original structure (Phemister, 1960).

East of the massif is an escarpment of Cambrian Quartzite that dips eastwards at about 17° to pass under the Moine Thrust. The outliers on the eastern spurs of An Teallach consist mainly of

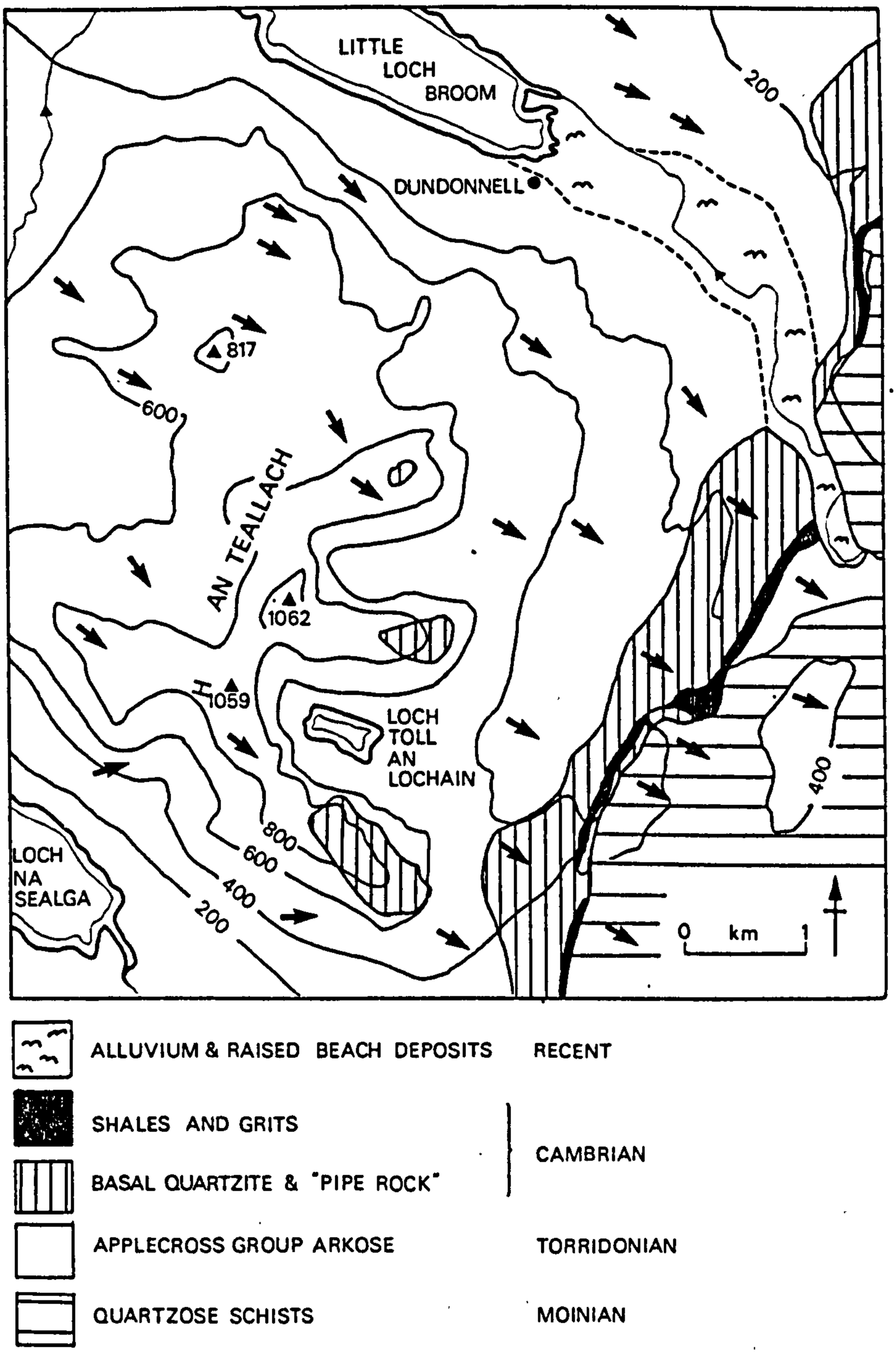


Figure 4.2: The geology of the An Teallach massif (based on Geological Survey 1:63360 map). Arrows indicate direction of dip of strata or cleavage planes.

"basal quartzite" (sensu Walton, 1964) consisting of a basal pebble conglomerate overlain by false-bedded grits and quartzites. The two largest outliers also contain patches of the overlying "pipe-rock", a fine-grained quartzite riddled with worm casts.

4.3.2 The Fannich Mountains (figure 4.3)

Apart from a ring of hornblende-rich acid orthogneiss and schist that surrounds the Western Fannichs, the rocks of the Fannich Mountains all belong to the Moine Series. The ring of acid orthogneiss was first interpreted as a stratigraphic inlier of Lewisian rocks (Flett, 1906; Peach et al., 1913a), and although the pre-Moine age of these rocks has been disputed (Read, 1934, 1956) they are now generally regarded as Lewisian thrust slices or the Lewisian core of a recumbent fold (Johnson, 1964). In general they comprise a grey, rudely-foliated gneiss consisting of quartz and feldspar with variable quantities of pyroxene, hornblende and mica (Phemister, 1960).

The Moinian rocks that occupy the rest of the area form part of a metamorphosed sedimentary succession that was deposited either before or at the same time as the unaltered Torridonian rocks west of the Moine Thrust. The nature of the sediments, which occasionally retain their original structures (Sutton and Watson, 1954) indicates shallow water deposition. There is debate as to the main period of metamorphism, though most evidence (Phemister, 1960) favours an early Caledonian (Siluro-Devonian) age. The Moinian rocks, hereafter referred to as Moine Schists, consist of alternating psammitic and pelitic groups. The psammitic types range from quartzose schists to massive flaggy quartz-feldspar granulites. The pelitic schists are those in which felts of muscovite and biotite form the bulk of the rock, giving marked schistosity. There is every gradation between the two groups, intermediate rocks being generally described as semipelitic. Peach et al. (1913a) recognized four varieties of Moine Schist in the Fannichs:

(i) a massive biotite gneiss in the Western Fannichs and the area immediately east of the Lewisian outcrop; (ii) flaggy, fine-grained granulitic quartzose schist or gneiss (Mheall a'Chrasgaidh Rock) in a band east of (i); (iii) pelitic biotite-schist running north-south through Sgùrr Mór and incorporating a band of quartzose granulite

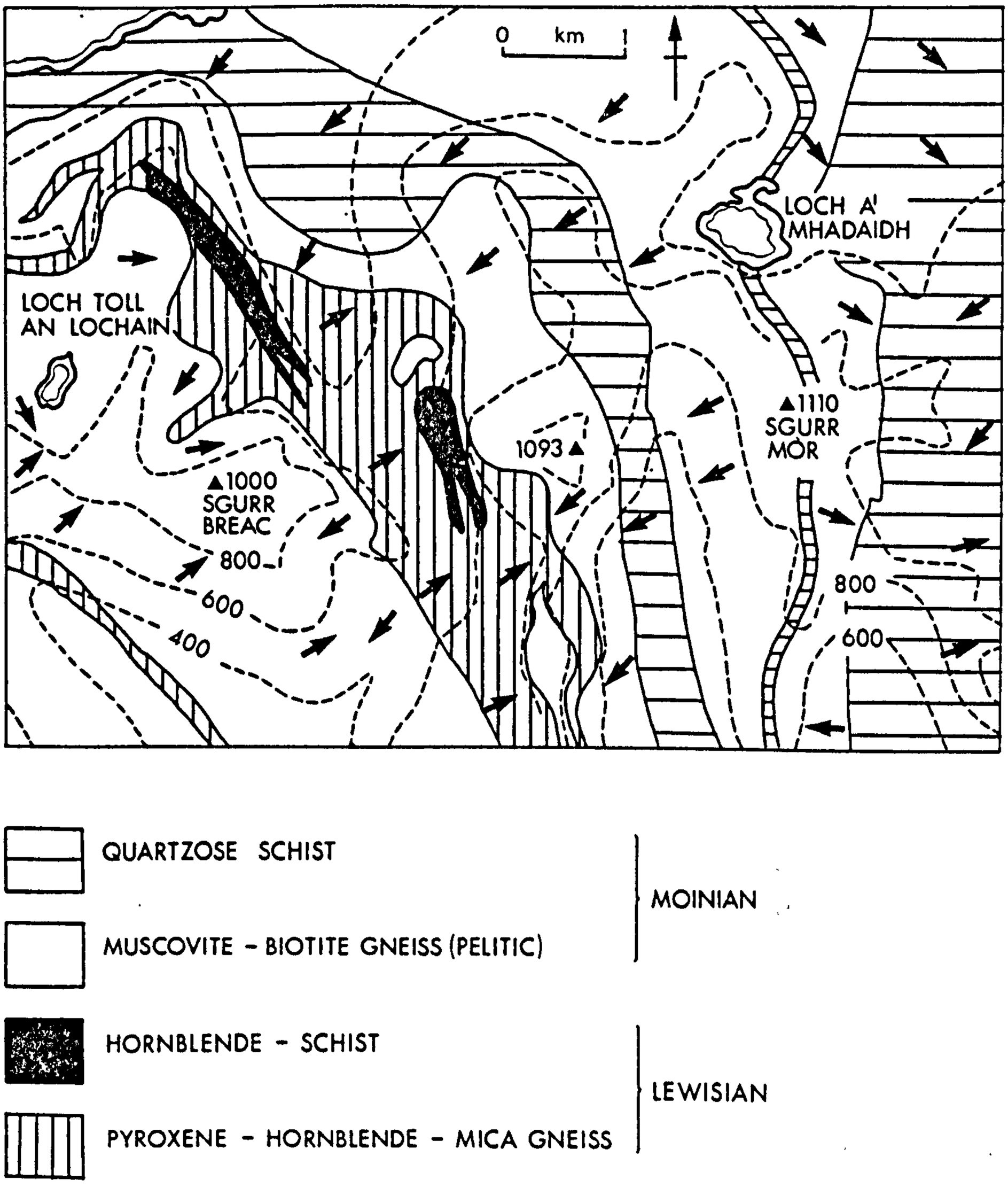


Figure 4.3: The geology of the Fannich Mountains (based on Geological Survey 1:63360 map). Arrows indicate direction of dip of cleavage planes.

schist that underlies Loch a'Mhadaidh; and (iv) flaggy granulitic quartzose mica-schist ("the typical Moine Schist") occupying the easternmost part of figure 4.3.

4.3.3 Ben Wyvis (figure 4.4)

Structurally, Ben Wyvis consists of a compound synclinal fold of Moine Schists, with the axis of the fold running parallel to the central ridge of the massif. The dominant lithology on either side of the fold axis consists of pelites (muscovite-biotite schists and gneisses) which enclose "islands" of siliceous rocks (flaggy to massive quartz biotite granulites) within smaller synclinal folds. To the N.W. and S.E. rocks of a lower siliceous zone (highly quartzose flaggy schists) pitch below the pelites. All three groups of rocks are characterized by considerable lithological variation (Peach et al., 1912). Particularly distinctive are the rocks that crop out on Càrn Gorm, S.E. of Loch Bealach Culaidh: these consist of felspathic granulites, and are much more massive than others in the group.

4.3.4 Discussion

The simplicity of the geologic maps (figures 4.2, 4.3, 4.4) belies the diversity of the rocks that are found within the study areas. Great lithological variety is particularly evident amongst the Moine Schists, which range in habit from the fissile mica-schists of the pelitic zones to the massive siliceous and felspathic rocks of the psammitic zones. Even the Torridon Sandstone of An Teallach displays considerable variety, ranging from pebbly conglomerate to shale over a few centimetres. It is not surprising, therefore, that the rocks of the study areas exhibit a very varied response to weathering under periglacial conditions, producing different types of surficial regolith. This topic is returned to in chapter 6.

4.4 Glacial History

The officers of the Geological Survey who mapped the Northern Highlands described evidence for three periods of glaciation:
(i) a period of maximum glaciation when ice covered all of the land

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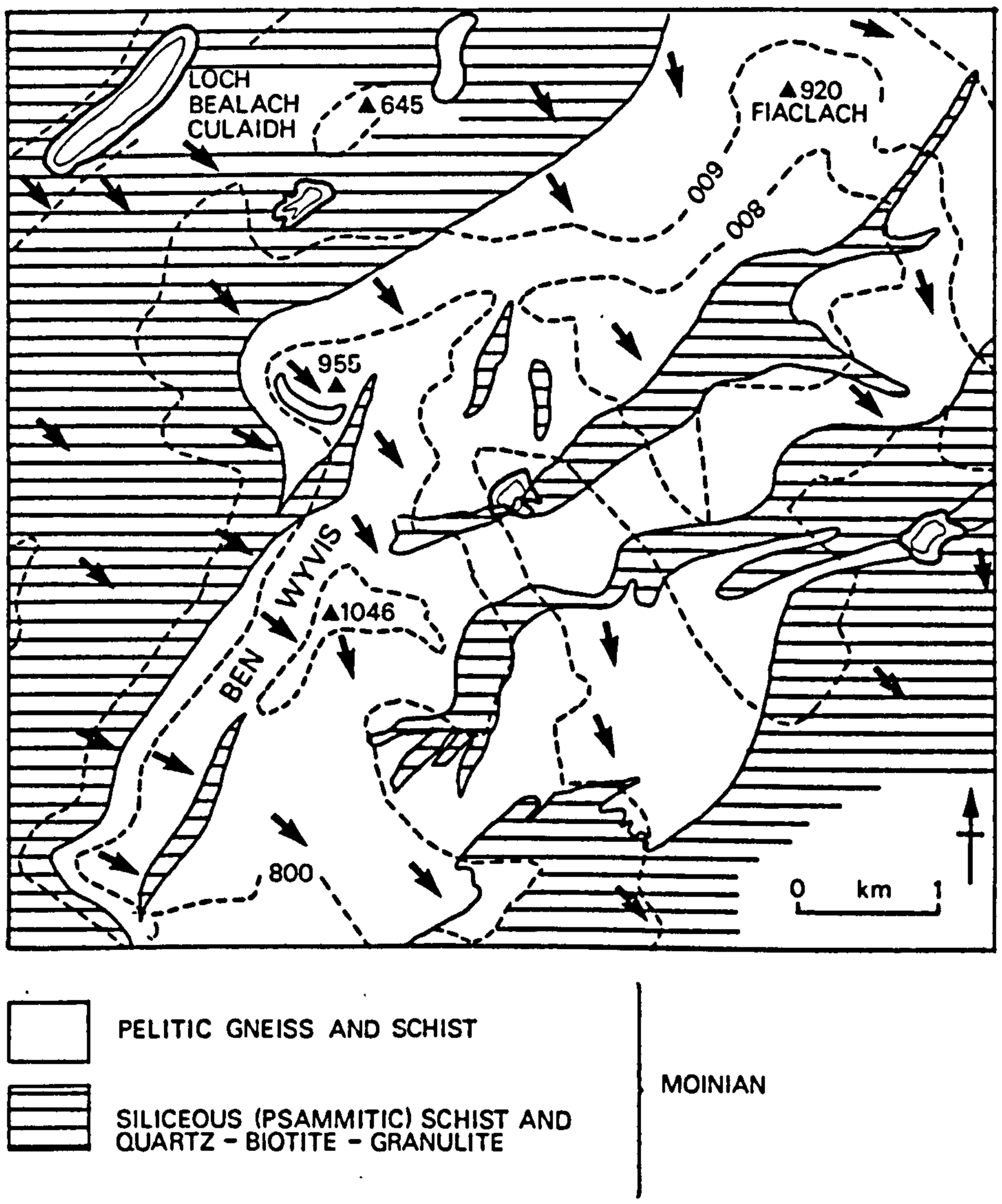


Figure 4.4: The geology of the Ben Wyvis massif (based on Geological Survey 1:63360 map). Arrows indicate direction of dip of cleavage planes.

and ice movement was partly independent of topography; (ii) a period of confluent glaciers when the higher summits were ice-free but glaciers radiating from central massifs occupied major valleys and coalesced on low ground; and (iii) a period of local corrie and valley glaciers. Although many of the details of the glacial history of the area remain to be resolved, this three-stage interpretation still appears reasonable in the light of present evidence. The "period of maximum glaciation" may be considered to represent the period during which the last ice-sheet was at its maximal extent, and Sissons (1977a) has demonstrated that many of the features of the "corrie and valley glaciers" stage were produced by the regeneration of glaciers during the Loch Lomond Stadial. The status of the intervening "confluent glacier" stage is less certain, although evidence in Wester Ross (Robinson and Ballantyne, 1979) suggests that a readvance may have interrupted the retreat of the Late-Devensian ice-sheet in that area.

4.4.1 The ice-sheet maximum

The presence by Loch Broom of augen-gneiss erratics derived from an outcrop *c.* 10 km east of the present watershed led Peach *et al.* (1912, 1913a) to conclude that when the ice-sheet was at its maximal extent its highest point lay well to the east of the present watershed (figure 4.1). Curiously, however, augen-gneiss erratics are found no higher than 730 m on the western flank of Ben Wyvis, and only below 540 m on the eastern side of the same hill. Peach *et al.* (1912) concluded from this that when the ice-sheet was at its thickest the ice-shed lay above the Wyvis massif, and that the deposition of augen-gneiss erratics below 730 m reflected westward migration of the ice-shed during subsequent downwastage. Striae above 600 m on Beinn Dearg, the Fannichs and An Teallach (Peach *et al.*, 1913a, figure 7) support west to north-west movement of the ice at its thickest, as does the presence of erratics of Moine Schist and "Thrust" Torridon Sandstone at nearly 900 m on the southernmost Cambrian Quartzite outcrop (Sàil Liath) on An Teallach. A band of quartzite erratics apparently derived from the northernmost outcrop crosses the northern plateau of An Teallach at NH 068860 and also indicates movement of the upper layers of the ice-sheet in a north-westerly direction.

In sum, the above evidence indicates that the last ice sheet at its maximal extent covered all of the mountains of the study area and that the ice-shed lay some distance east of the present watershed, possibly over Ben Wyvis, so that ice movement over the Fannichs and An Teallach was towards the west or north-west.

4.4.2 The period of confluent glaciers

Peach et al. (1913a) considered that the period of maximum glaciation was succeeded by one in which the Fannichs and Beinn Dearg acted as centres of dispersion of glaciers " ... which became confluent in the valleys and spread over the low plateaux as a continuous sheet". According to Phemister (1960, p. 97), " ... An Teallach and Ben Wyvis respectively split the west and east movements from the Fannich centre", and thereby formed nunataks standing above coalescent valley glaciers. The evidence for a distinct "confluent glacier stage" is fourfold. First, striae on low ground outside the limit of the later Loch Lomond Advance around An Teallach and Ben Wyvis indicate that the last ice movement was around rather than over these massifs. This could have resulted, however, from deflection of the basal ice during the ice-sheet stage. More significantly, Phemister (1960) identified two sets of striae in Wester Ross, one group reflecting the passage of the ice-sheet, the second a later period of glaciation. Secondly, the distribution of augen-gneiss erratics around Ben Wyvis described above suggests that these were deposited by valley glaciers that emanated from west of the source outcrop (i.e. in the vicinity of the Fannichs) and encircled the massif. Thirdly, lateral moraines on Ben Wyvis (Peach et al., 1912) and An Teallach (Robinson and Ballantyne, 1979) indicate that at some time following the ice-sheet maximum glaciers occupied the valleys surrounding these massifs but were absent from high ground. On Ben Wyvis, the limits of augen-gneiss erratics appear to coincide with lateral moraines (Peach et al., 1912). The presence of such moraines suggests that retreat of the ice margin was halted or even reversed. Finally, Robinson and Ballantyne (1979) described end and lateral moraines on low ground around Lochs Gairloch and Ewe, and they suggested that these and the An Teallach laterals represent the limits of a readvance (the Wester Ross Readvance)

that interrupted the retreat of the last ice sheet. Other authors (Kirk et al., 1966; Smith, 1977; Synge, 1977) have identified glacial limits both in the west (e.g. near Ullapool) and in the east (e.g. in Glen Glass, near Garve and in the Moray Firth), but the limited evidence available makes the validity of these limits difficult to assess.

4.4.3 The Loch Lomond Advance

Although the age and duration of the "confluent glacier stage" or Wester Ross Readvance is unknown, the radiocarbon date of 12,810 ± 155 B.P. obtained by Kirk and Godwin (1963) from Lateglacial Interstadial deposits at Loch Droma (between the Fannichs and Beinn Dearg) suggests deglaciation by c. 13,000 B.P. The glaciers of the subsequent Loch Lomond Advance were relatively small and confined to mountain areas. The limits of these glaciers were mapped and described by Sissons (1977a). On An Teallach he found evidence for six former glaciers facing between N.W. and east (figure 4.5) and he delimited in the Fannichs two corrie glaciers and a large ice mass produced by the coalescence of corrie glaciers on the N.E. side of the ridge east of Sgùrr Mór (figure 4.6). The author concurs with all of these limits except that to glacier 1 (figure 4.5), which is here interpreted as marked by a till ridge 2-3 m high and 210 m long some 500 m outside Sissons' postulated limit. It is also possible that fluted moraines and hummocky drift in other parts of the Fannichs (e.g. north of Sgùrr Breac) reflect the former presence of Loch Lomond glaciers, but this evidence is equivocal. Neither Sissons nor the author found evidence for local glaciers in the Wyvis massif. The absence of such glaciers here is consistent with a rise in former firn line altitudes from N.W. to S.E. across the Northern Highlands, interpreted by Sissons as reflecting decrease in former snowfall inland from the west coast.

4.4.4 Discussion

The evidence outlined above indicates that high ground in all three study areas was covered by the last ice sheet but exposed to the operation of periglacial processes both during subsequent downwastage (the "confluent glacier" stage) and during the Loch Lomond Stadial.

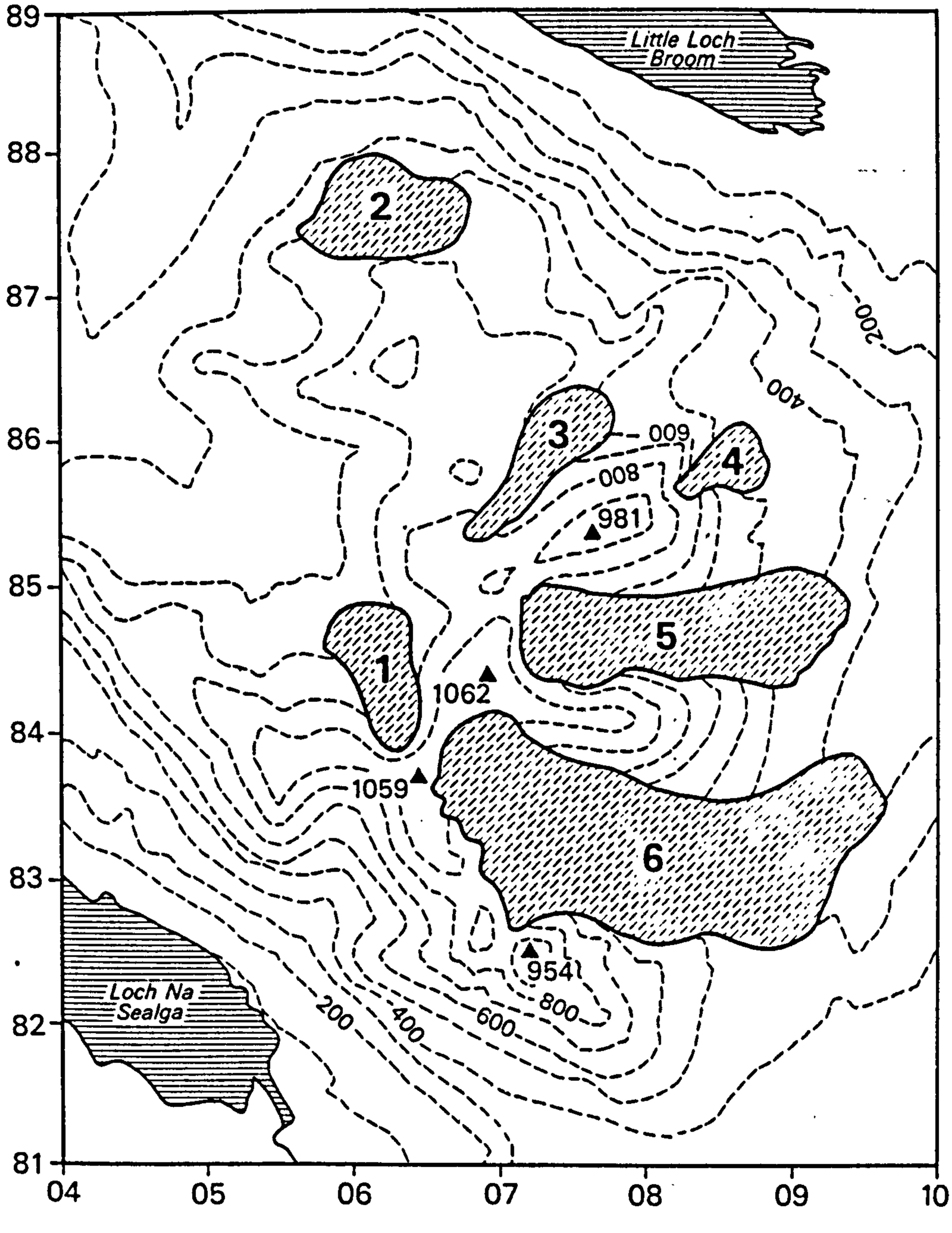


Figure 4.5: Loch Lomond Advance glaciers on An Teallach. The glacierized areas are stippled. Based on Sissons (1977a).

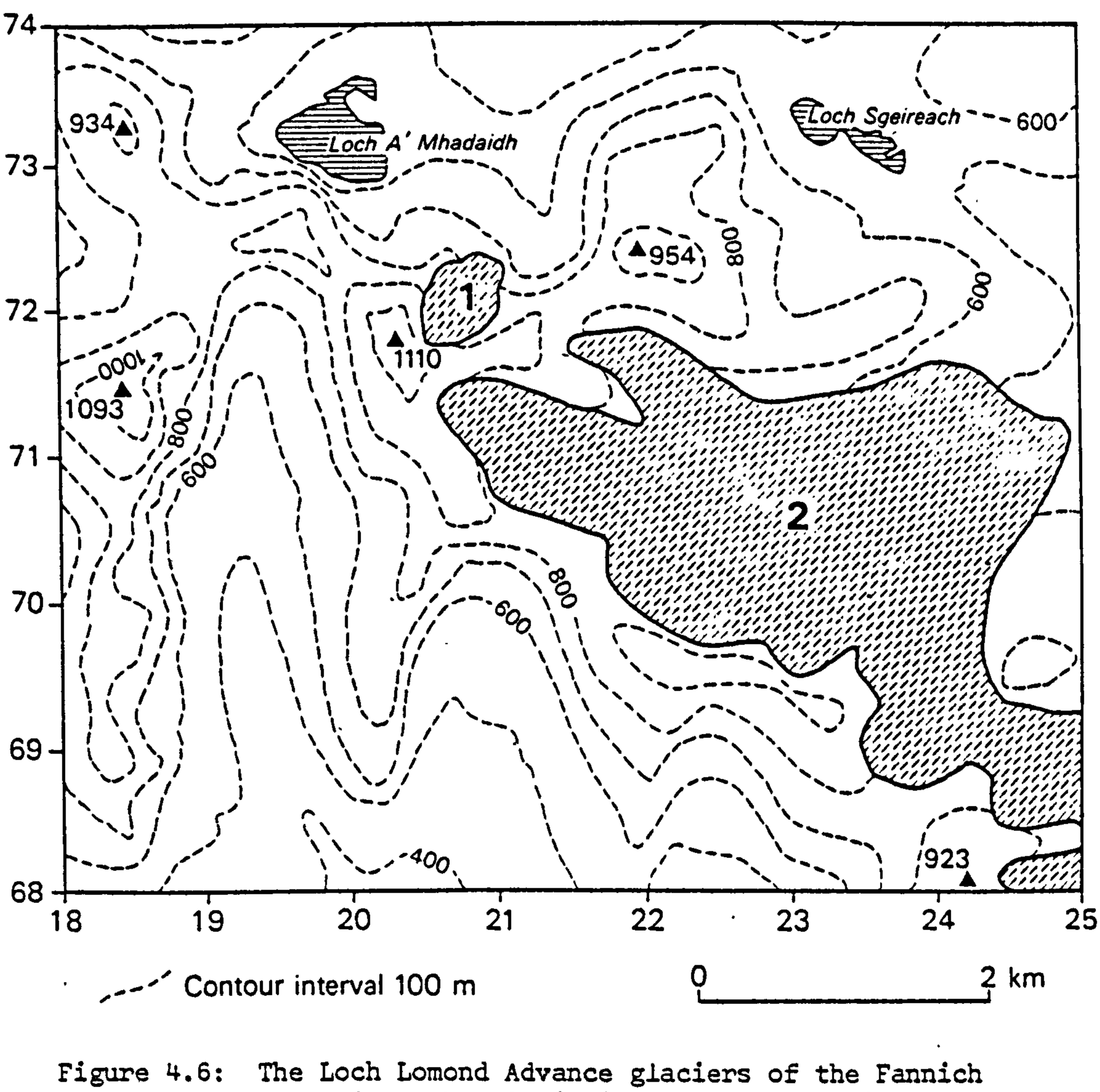


Figure 4.6: The Loch Lomond Advance glaciers of the Fannich Mountains. The glacierized areas are stippled. Based on Sissons (1977a).

The limits of the Loch Lomond glaciers are in general well-defined, particularly on An Teallach; as indicated in chapters 2 and 3, this is advantageous for establishing the inactivity of certain periglacial features during the Flandrian.

4.5 Soils and vegetation

The most important factor affecting the nature of upland soils in the study areas is lithology or, more specifically, the manner in which bedrock has broken down under periglacial weathering (chapter 6). On fine-grained Cambrian Quartzite, the granulite of Càrn Gorm and some psammitic schists weathering has produced a cover of coarse detritus and soil development is limited to isolated pockets of fines. In contrast, Torridon Sandstone has yielded a regolith of stones embedded in coarse felspathic sand with less than 2% clay and silt in the fine fraction. The soils of An Teallach are consequently well-drained, skeletal and azonal, with organic matter restricted to the shallow surface root zone. The absence of clay and humus and the coarseness of texture render such soils highly susceptible to erosion by wind and water. Only in poorly-drained corries below 700 m is there significant peat formation.

Mature zonal soils are restricted in distribution to the pelitic schists of the Wyvis plateau and parts of the eastern Fannichs. In these areas classic podzols are found, with thoroughly leached micaceous A horizons and organic B_1 horizons up to 20 cm thick. The soil texture in these areas is finer, and cohesion consequently greater, with 6-15% of the fine fraction composed of clay and silt. The lack of gleying indicates generally free drainage. With increasing slope the podzolic structure becomes less marked, eventually giving way to the homogeneous brown earths characteristic of solifluction soils. Immature rankers are also found on pelitic rocks, particularly where the regolith is shallow, as in the western Fannichs. Acid oligotrophic peat is common on gentle to moderate slopes up to 800 m on Moinian rocks, but is rare on exposed ridges and plateaux.

Contrasts in vegetation cover also reflect underlying lithology. Vegetation is patchy on coarse detritus and on exposed slopes and plateaux in Torridon Sandstone areas, but pelitic schists support a complete cover of vegetation. The northern plateau of An Teallach,

for example, resembles a desert, colonized only by small clumps of grass and heather, whereas bare ground is rare on the Wyvis plateau. The paucity of vegetation on the An Teallach plateau is attributable to the sterility and coarseness of the regolith and to deflation of surficial fines (chapters 8 and 11).

The species and associations characteristic of Ben Wyvis and An Teallach also differ. The summit plateau of Ben Wyvis and most of the Fannich ridge support a Rhacomitrium heath (Rhacomitrium lanuginosum and Carex bigelowii with few other species). On the N.W. flank of Ben Wyvis this grades into bilberry heath (Vaccinium myrtillus and Empetrum nigrum with grasses and sedges) that ends abruptly at 600 m, the upper limit of thicker peat and a Calluna-dominant association. The sheltered, gentler eastern slopes of Ben Wyvis support various grasses and mosses, such as Deschampsia and Rytidiadelphus species (Pearsall, 1968). On An Teallach the vegetation displays greater diversity. Below 600 m heath predominates (Erica tetralix, E. cinerea, Calluna vulgaris) and above this altitude the distribution of species reflects the nature of the terrain. On exposed plateaux up to 800 m clumps of Calluna form the main vegetation, but above 800 m Empetrum becomes dominant in exposed areas. On sheltered lee slopes and the sheltered risers of turf-banked terraces Deschampsia flexuosa, Festuca ovina, Juncus triffidus and Alchemilla alpina are common, with Nardus stricta occupying sites of late-lying snow. Vaccinium myrtillus is locally abundant, especially on coarse detritus. The above lists exclude many species that are common but unimportant in terms of cover.

4.6 Discussion

The three areas described above were selected for study on the basis of the range of the periglacial features they contain. In part, this range reflects the environmental variety in the three areas, which is conditioned principally by geology: the rocks in the study area have responded in different ways to periglacial weathering, producing distinctive types of regolith, and the nature of the regolith is the primary determinant of upland soil type, the nature of vegetation cover and, as will become apparent in later chapters, the types of periglacial landforms present in any area. The other major source of

environmental variation - climate - was not considered above as data on the upland climate of the northern Highlands are lacking. This lack prompted direct observations on weather conditions over a three year period; these form the subject of the next chapter.

CHAPTER 5 METEOROLOGICAL OBSERVATIONS

5.1 Introduction

The poverty of reliable routine observations of mountain weather in a country as weather-conscious as Great Britain is remarkable. Since the manned observatory on Ben Nevis closed on 1 October 1904 few records have been kept of conditions above 600 m. A list of data sources compiled by Taylor (1976, pp. 264-5) shows that most recent information relates to Wales and the Pennines, although some short-term observations have been made on the Cairngorms (Dybeck and Green, 1955; Baird, 1957) and on some other Scottish mountains such as Lowther Hill in the Southern Uplands (unpublished data, Meteorological Office, Edinburgh). An automatic weather station recently established on Cairngorm summit at 1245 m (Curran et al., 1977) has not been in effective operation for a sufficient period to have yielded much useful information.

Not only are existing data on mountain weather localized and fragmentary (none exist at all for ground above 600 m in the Northern Highlands) but such data as are available are often of limited value to the geomorphologist, whose concern is with moisture and temperature conditions at or under the ground surface and whose interests are best served by observations closely spaced in both time and distance. The lack of such data prompted the establishment of a network of meteorological instruments at altitudes up to 950 m on An Teallach with a view to relating the limits and nature of present-day periglaciation to climate. It must be stated at the outset, however, that this venture proved only partly successful owing to the difficulties of maintaining remote meteorological equipment under conditions of high exposure, prolonged snowcover and frequent freezing. A great deal of meteorological information was nonetheless obtained between April 1976 and September 1979, and this is described and discussed below under the headings of air temperature, surface and subsurface temperature, precipitation, snowcover and wind.

5.2 Air temperature

Air temperature provides the fundamental control on many forms of periglacial activity in that it determines the length, frequency

and intensity of freezing at and below the ground surface. Mean monthly air temperatures for a number of upland locations are summarized in figure 5.1. These show that mean monthly temperatures in upland areas are predominantly positive except on the very highest peaks where mean annual temperatures lie close to 0°C. Below 800-900 m sub-zero mean monthly temperatures occur in only four months of the year (December-March) and even above 1000 m average air temperatures fall below zero for only six months (November-April).

Such averages, however, conceal short-term temperature fluctuations that are of critical importance for the understanding of the operation of frost processes in upland Britain. An attempt was made to study these through the installation of Stevenson screens containing clockwork-powered thermographs at sites at approximately 600 m, 700 m, 800 m and 950 m on An Teallach. The record obtained from the higher and more exposed sites proved of little value, however, as the instruments frequently jammed or froze and the screens filled with snow, so attention is here focussed on the results obtained from screens installed at 610 m on the exposed summit of Meall Garbh (NH 081869; figures 5.2 and 5.3) and at 665 m in a sheltered location at the head of Coire a'Mhuilinn (NH 073859; figure 5.4). In both cases standard Stevenson screens were mounted about 1.5 m above ground level. Before 1 May 1978 the record at the first site was obtained with a Negretti-Zambra bimetallic thermograph and at the second with a Casella bimetallic thermohygrograph, both equipped with 32-day clocks. After that date the record at both sites was obtained using Grant model D recorders with thermistors mounted inside the screens as sensors. The accuracy of all of these instruments is given by their manufacturers as $\pm$ 0.6°C.

Both records (figures 5.3 and 5.4) are fragmented as a result of the difficulties of operation outlined above, but nonetheless provide sufficient information to allow some generalizations to be made about the nature of the air temperature regime. The overall range of recorded temperature was large. At 610 m the minimum recorded was -6.2°C (12 01 78), the maximum 25.5°C (10 07 76). The equivalent figures for 665 m were -10.6°C (01 01 79) and 26.2°C (10 07 76). Temperatures lower than -13°C were recorded at 665 m during the period January-April 1979 although the date of occurrence is unknown as the wind-on spool in the recorded jammed. The

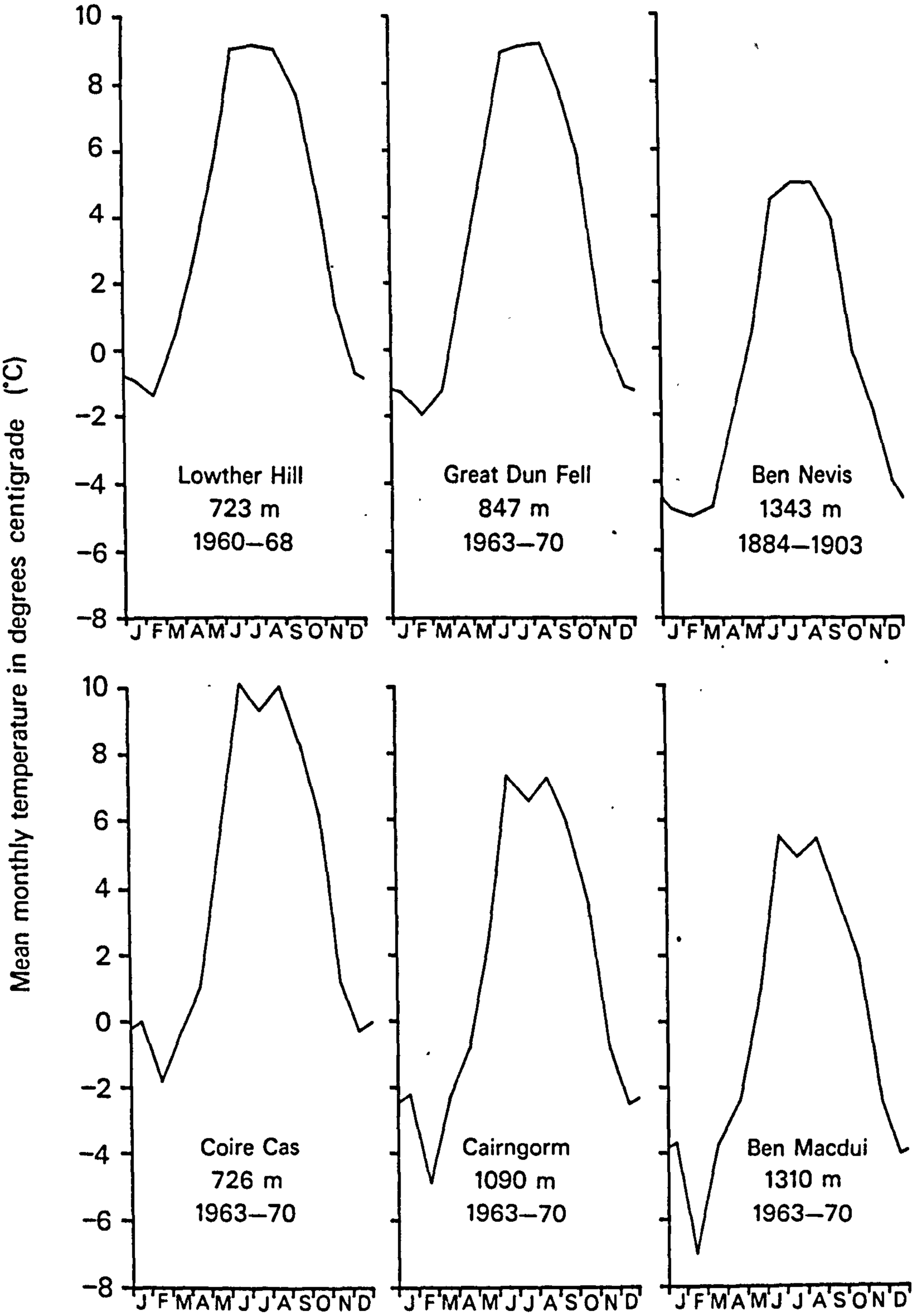


Figure 5.1: Mean monthly air temperatures for upland sites in Great Britain. Source: unpublished data, Meteorological Office, Edinburgh.



Figure 5.2: Stevenson screen mounted on the exposed summit of Meall Garbh, An Teallach, at an altitude of 610 m.

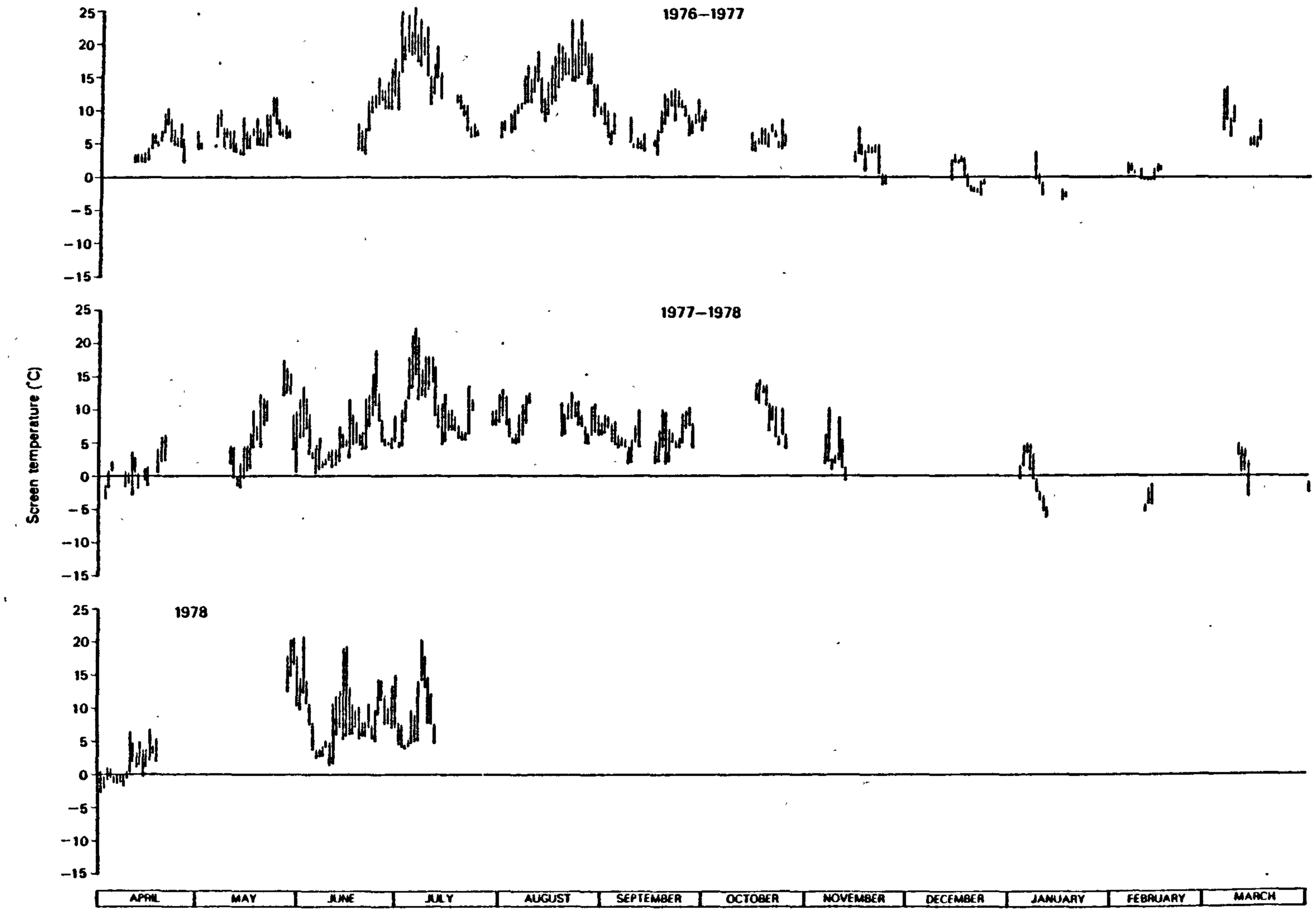


Figure 5.3: Diurnal screen temperature ranges at 610 m altitude on Mheall Garbh, An Teallach

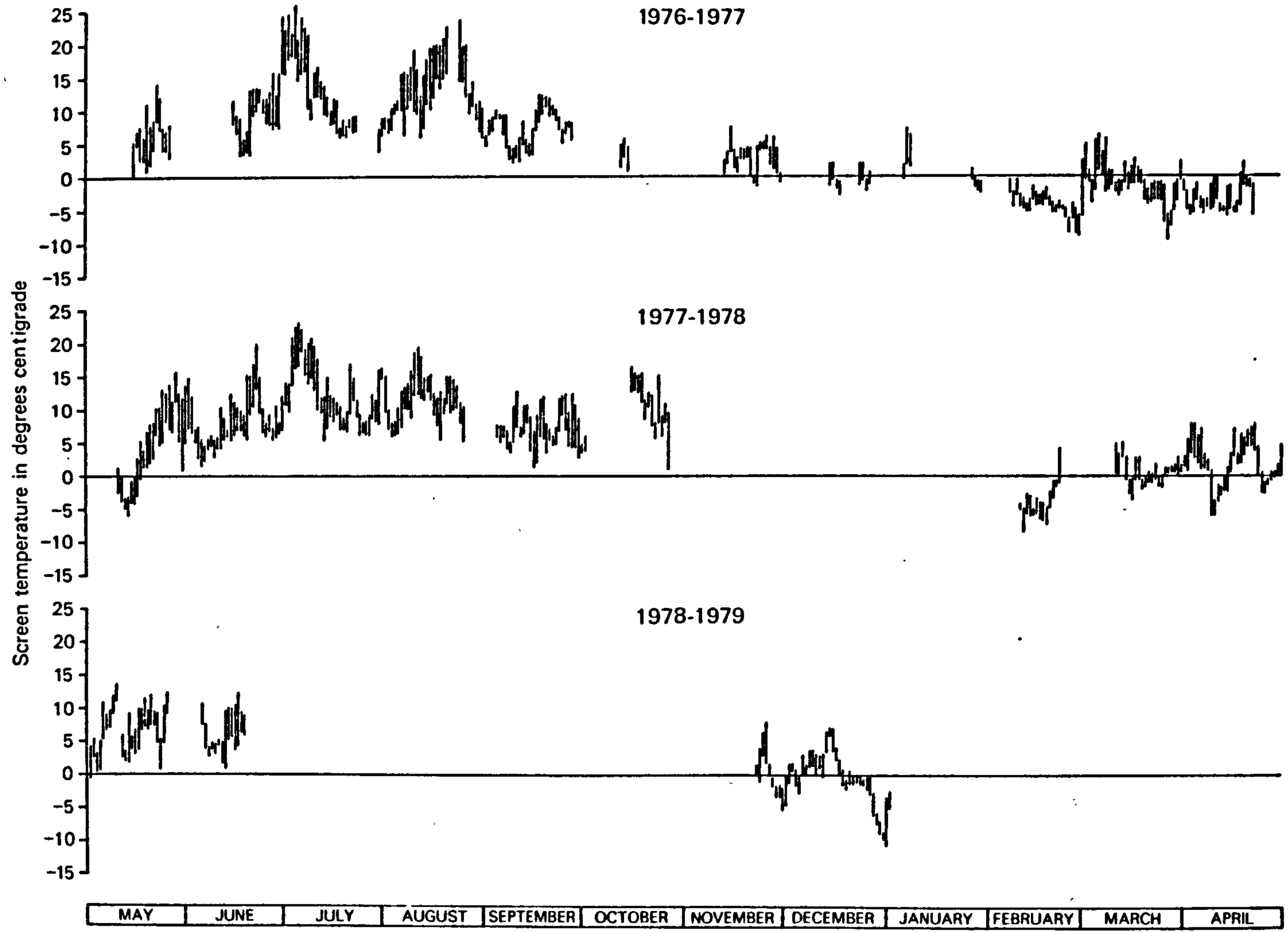


Figure 5.4: Measured diurnal temperature ranges at 665 m altitude in Coire a'Mhuilinn, An Teallach.

occurrence of significantly lower temperatures at the 665 m site probably reflects its location in an upland valley (into which cold air flows during winter anticyclones) rather than the differences in altitude between the two stations. The descent of cold air on calm cloudless winter nights often produces marked inversions so that temperatures in the valleys are lower than those on adjacent peaks. This is apparent in data from the Cairngorms (table 5.1) and explains the occurrence of much lower temperatures on low ground (e.g. -27°C recorded at Braemar in February 1895) compared with the mountain tops. (The lowest temperature measured on the summit of Ben Nevis between 1883 and 1904 was -17.3°C). The available data suggest that air temperatures lower than -10°C are rare on the slopes of An Teallach and other British mountains.

The temperature regime at 610 m and 665 m can be subdivided into two "seasons". The summer season is characterized by an absence of sub-zero air temperatures and lasts from June to October. During the period of measurement July and August temperatures did not fall below 4°C. Assuming a lapse rate of 0.006°C m^{-1} this implies that even the highest parts of the study areas did not experience sub-zero air temperatures at this time of year, although temperatures below 2°C in June, September and October suggest that sub-zero air temperatures occasionally occurred above 900-1000 m in these months. On sunny summer days high insolation produced diurnal temperature ranges of up to 10°C and sometimes greater. Diurnal temperature cycles were, however, superimposed on longer fluctuations of several days' duration (e.g. June-September 1977, figure 5.4) that reflect the passage of fronts and the alternation of high- and low-pressure systems.

The upland winter season at 610-665 m lasts from November to April inclusive as May generally marks the transition from winter to summer conditions. During these months the diurnal temperature range is generally much lower: at 610 m 71.5% of recorded winter days fell within the range -5°C to +5°C, and 91.4% fell between -7.5°C to +7.5°C, the corresponding figures for 665 m being 67.3% and 89.4%. The range of air temperature for any day rarely exceeded 5°C during these winter months, such "dampening" being attributable to the reduced effectiveness of direct insolation (particularly at the 665 m site, which lies in shadow from November

Table 5.1

Annual minimum temperatures recorded in the N.W. Cairngorms

Year	Minimum temperatures (°C) recorded at: Ptarmigan Restaurant Cairngorm (1090 m).	Coire Cas (763 m)	Glenmore Lodge (335 m)
1963	N.D.	N.D.	- 18.2
1964	- 12.2	- 11.1	- 13.3
1965	N.D.	N.D.	- 14.2
1966	- 12.2	- 11.7	- 14.7
1967	N.D.	N.D.	- 11.7
1968	- 12.2	- 10.0	- 14.9
1969	- 15.6	- 15.0	- 14.9
1970	- 12.2	- 10.6	- 13.9

Source: unpublished data, Meteorological Office, Edinburgh.

N.D. - No Data

to February). As many days have temperatures entirely above or below 0°C, however, days on which the freezing point was crossed were surprisingly few (19.8% of recorded winter days at 610 m, 23.6% at 665 m) and freezing cycles comprising an initial drop below 0°C and subsequent rise above 0°C are even rarer: the available data suggest that there are only about 30-35 freeze thaw cycles per winter season of c. 200 days at these altitudes. This accords with thermograph records for Coire Cas (763 m) in the Cairngorms, where 32 cycles were recorded in 1974 and 41 over the winter 1974-5 (figure 5.5). In midwinter (e.g. February 1977, figure 5.4; February 1975, figure 5.5) the frequency of such cycles is sometimes reduced as temperatures are persistently below freezing, whereas in November and April the low frequency of cycles at these altitudes reflects diurnal ranges that lie predominantly above the freezing point. It seems unlikely that higher altitudes experience a much greater number of freeze-thaw cycles as any increase resulting from the effective extension of the winter season and depression of November and April temperatures is likely to be at least partly offset by a reduction in thaw events in midwinter.

The amplitude of freeze-thaw cycles at the 610 m and 665 m sites is interesting. Fewer than 30% of the cycles recorded during the winter months were diurnal, and more than 40% had an amplitude of three days (below freezing) or longer. This indicates that most freezing cycles are related not to nocturnal cooling and diurnal heating, but to the alternating dominance of relatively cold and relatively warm airmasses and the passage of fronts associated with eastward movement of depressions across the North Atlantic. This interpretation is supported by the irregular timing of periods of freezing and thawing, which occur at all times of day or night. Typical long-term cycles are clearly illustrated on figure 5.4, particularly during March-April 1978 and November-December 1978. On the other hand, air temperatures at these altitudes display no true annual freeze-thaw cycle as freezing periods tend to be interrupted by periods of above-zero temperatures throughout the winter months. However, the occasional predominance of sub-zero temperatures over extended periods (e.g. February-April 1977, figure 5.4) suggests that higher altitudes may in some years support an annual freezing cycle 2-4 months in duration interrupted only briefly by above-zero temperatures.

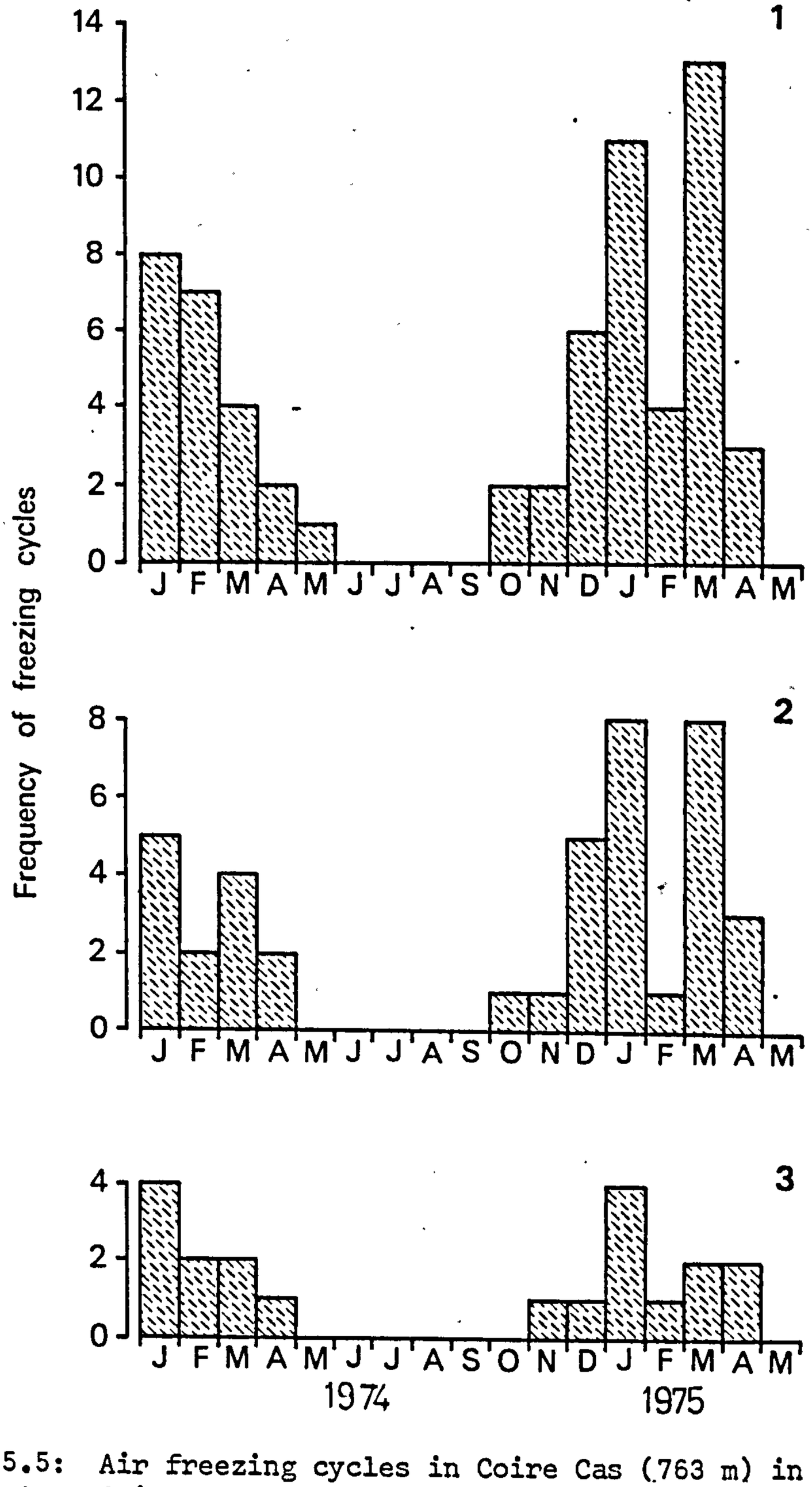


Figure 5.5: Air freezing cycles in Coire Cas (763 m) in the Cairngorms. 1 - frequency of occasions on which air temperature dropped below 0°C. 2 - frequency of freezing cycles with minimum amplitude +1.0°C to -1.0°C. 3 - frequency of freezing cycles with minimum amplitude +2.0°C to -2.0°C. Based on unpublished data, Meteorological Office, Edinburgh.

5.3 Ground surface and subsurface temperature

Most of the processes that characterize periglaciation are related to freezing and thawing of the ground. Although the depth, intensity and frequency of ground freezing are ultimately controlled by air temperature regime, the relationship between air temperature and ground surface temperature is often complex. The ground surface, for example, may be subject to the heating effect of direct solar radiation or insulated from air temperature changes by vegetation or snow. Moreover, the rate of heat transfer within soil and rock varies with air, water and ice content and is affected by void ratio, granulometry and composition. The characteristics of the surface and subsurface temperature regimes at any site cannot therefore be readily predicted from the air temperature regime and must be measured directly. This section describes the results of measurements of surface and subsurface temperature on An Teallach and elsewhere with a view to identifying some of the main characteristics of the ground temperature regime in upland Britain, particularly during the winter season when air temperatures fall below freezing point.

Surface and subsurface temperatures on An Teallach were measured during the winter of 1978-9 using two groups of thermistors buried at depths of a few millimetres to 20 cm. These were located about 5 m from the 610 m and 665 m weather stations and soil temperatures measured by the thermistors were recorded at one-hour (610 m) or three-hour (665 m) intervals on Grant model D recorders housed in the Stevenson screens. Unfortunately the poor performance of both recorders under winter conditions resulted in long gaps in the winter records, and although periods of successful measurement provided information on the effect of snowcover and insolation on the relationship between air and soil temperature (figure 5.6) these records are inadequate for investigating the characteristics of the upland ground temperature regime. However, a fuller record was obtained using similar instrumentation in the course of research into patterned ground development on a south-facing slope at 660 m on Tinto Hill in the Southern Uplands (NS 954352) during the winter of 1979-80 (Ballantyne et al., in prep.; figure 5.7).

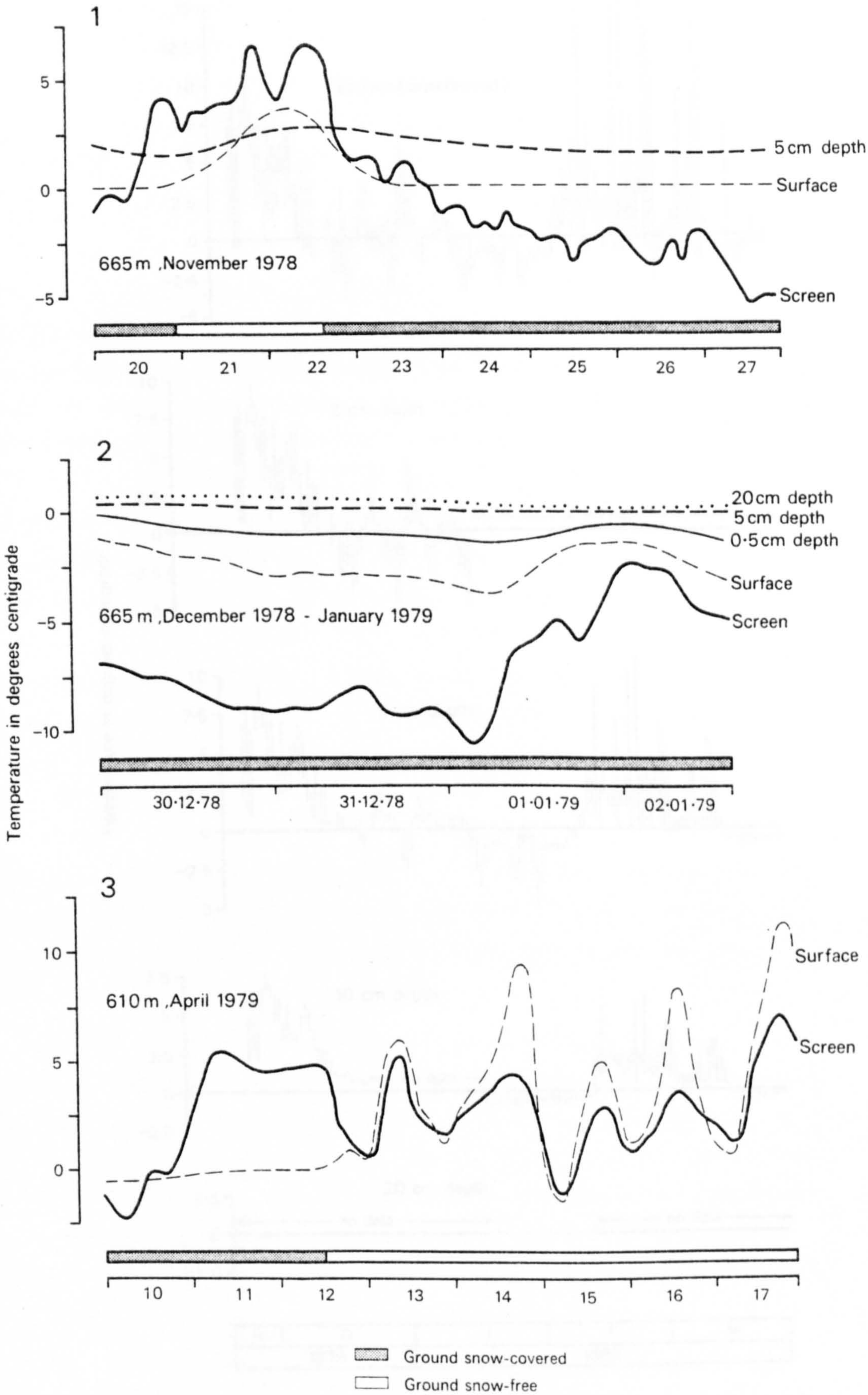


Figure 5.6: Screen, surface and subsurface temperatures on An Teallach, illustrating the insulating effect of snowcover (diagrams 1-3) and the effect of direct insolation (diagram 3) on ground temperatures.

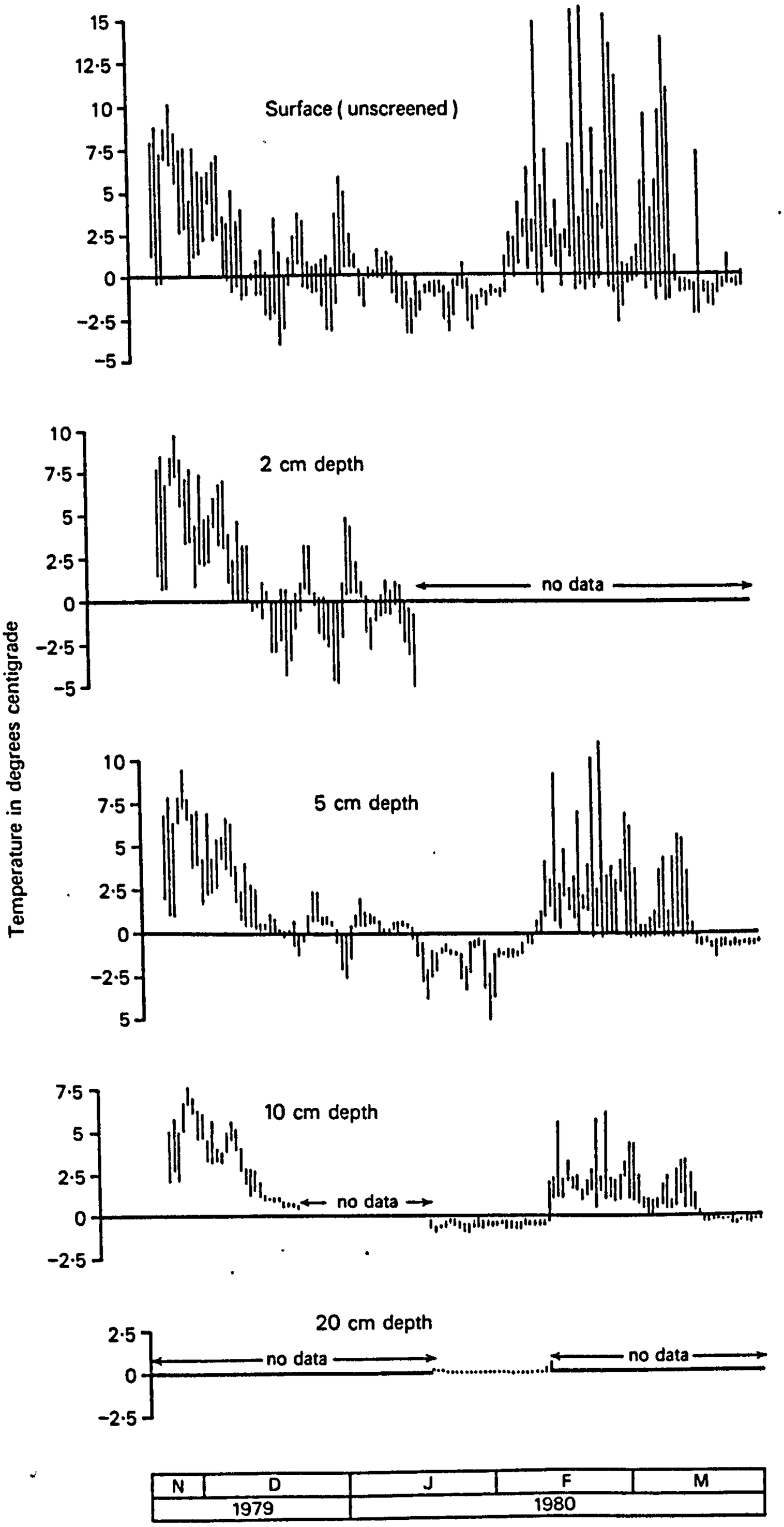


Figure 5.7: diurnal ranges of surface and subsurface temperatures at 660 m altitude on a south-facing slope cn Tinto Hill, southern Scotland, November 1979 - March 1980.

In a recent survey of ground temperature characteristics, Williams and Gold (1976) stated that in the absence of snowcover and direct solar radiation " ... the amplitude of a temperature variation at the ground surface is normally about equal to that of the corresponding one for air" (p. 3). They demonstrated that the amplitude of a given temperature change decreases exponentially with depth at a rate dictated by the time necessary for the completion of one complete cycle of cooling and heating (or vice versa).

The extent to which snowcover may disrupt these simple relationships is illustrated in figure 5.6. Diagram 1 contrasts the response of surface temperature to air temperature when the ground is snow-free (21-22 November 1978) with the lack of response following a heavy snowfall (23-27 November 1978) when surface temperature remained at 0°C despite a drop in air temperature to -5.4°C. Diagram 2 illustrates the insulating effect of shallow (5-10 cm) snowcover during a period of relatively severe freezing when air temperature fell to -10.6°C but surface temperatures dropped to only -4.0°C and ground below 5 cm depth remained unfrozen. The third diagram shows the effect of snowcover during a period of thaw when screen temperatures rose to over +5°C but surface temperatures remained at 0°C until the snow melted. On subsequent days the influence of direct insolation is apparent in that daytime temperatures at the ground surface rose well above screen temperatures. The effect of insolation is particularly marked in midsummer; on 9 July 1978, for example, temperatures at the ground surface at 610 m rose to almost 40°C and those at 5 cm depth reached 28°C even though screen temperatures reached only 20.4°C.

Both effects therefore tend to maintain ground surface temperature above the level of ambient air temperature. Even relatively shallow snowcover reduces the intensity of surface freezing and hence the depth to which the freezing plane penetrates in a single freeze-thaw cycle, and a thick snowcover may completely prevent ground freezing (diagram 1, figure 5.6). Given the high frequency of winter snowcover at altitudes above 700 m (section 5.5)

it seems likely that the frequency of geomorphologically effective freeze-thaw cycles at higher altitudes is much less than that of oscillations of air temperature about the freezing point. This is in accord with the conclusions of Harris (1974), who made measurements of winter ground temperatures in northern Norway and found that snow depth is of critical importance in determining the rate of soil freezing and that snowcover postponed thaw of the ground during the melt season and prevented the penetration of short-term (i.e. diurnal) freeze-thaw cycles into the ground.

A striking feature of the Tinto Hill record (figure 5.7) is the shallow depth to which diurnal freeze-thaw cycles penetrate the ground even in the absence of snowcover. During the period from mid-February to mid-March 1980 when the site was largely snow-free 19 diurnal freeze-thaw cycles were recorded at the surface but the ground temperature at 5 cm depth passed below freezing point only 8 times, on each occasion falling only a fraction of one degree centigrade below zero. This confirms findings from more northerly latitudes (Czeppe, 1960; Harris, 1974) that the limit of diurnal frost penetration rarely exceeds 5-6 cm in depth.

Conversely, longer periods of freezing resulting from the dominance of cold airmasses may penetrate the ground to much greater depths. During the prolonged cold spell from mid-January to mid-February 1980 temperatures at 10 cm depth remained below freezing for 26 days and temperatures at 20 cm depth dropped to freezing point. The length of this period of ground freezing was prolonged by snowfall after initial freezing. It seems, therefore, that only periods of prolonged freezing initiated under snow-free or almost snow-free conditions result in frost penetration to depths of more than a few centimetres. During the period of measurement on Tinto Hill such conditions occurred only three or four times. At higher altitudes and in snowier upland areas such as the Cairngorms the frequency of deep frost penetration may well be reduced by more persistent snowcover despite lower air temperatures except where the snow is removed by high winds (section 5.6).

Although measured subsurface freezing on An Teallach and Tinto Hill was restricted to the top 20 cm of the soil there is evidence that freezing to much greater depths sometimes occurs. Ragg

and Bibby (1966) reported ice lenses at a depth of 50 cm at the summit of Broad Law (830 m) in the Southern Uplands, and G.P. Chattapadhyay (pers. comm.) found ground ice at a similar depth on the Drumochter Hills following the relatively severe winter of 1978-9. The possible occurrence of ground freezing at still greater depths was predicted by Halstead (1974) from analysis of subsurface temperatures near the summit of Lowther Hill. Halstead's measurements extended from November 1967 to April 1968, and during this period (as on Tinto Hill) the only freezing events at depths greater than c. 5 cm accompanied prolonged periods of sub-zero air temperatures. The longest freezing event lasted from early February to late March 1968 and resulted in sub-zero temperatures at 30 cm depth for at least 30 days. Halstead extrapolated his data for this event to predict a maximum depth of frost penetration of 62 cm for this site (c. 750 m altitude) and corresponding depths of freezing of 22 cm for 500 m, 86 cm for 1000 m and 114 cm for 1250 m. However, the depths of freezing predicted for 1000 m and 1250 m take no account of probable increase in snowcover at these altitudes and therefore represent potential maxima rather than realistic values.

In summary, the available data suggest that the winter ground temperature regime in upland Britain is characterized by two types of freeze-thaw cycle: cycles penetrating to depths of 10 cm to more than 50 cm that result from prolonged sub-zero air temperatures and shallow cycles reaching less than 5-6 cm depth that result from nocturnal cooling. Snow cover tends to reduce the depth and intensity (but not necessarily the length) of freezing associated with the former type and inhibits the operation of the latter. Given these constraints it seems likely that geomorphologically effective freeze-thaw cycles are not frequent. The Tinto Hill and Lowther Hill data suggest that four or five "deep" freezing cycles per year are typical at 650-750 m in the Southern Uplands, but the data available do not permit accurate estimation of the frequency of "shallow" 24-hour freeze-thaw cycles. It seems likely that the latter will be most frequent in autumn and particularly spring when snowcover is less extensive and insolation is more effective.

Finally, although Ragg and Bibby (1966, p. 13) state that on British mountains " ... the soil is often frozen from late December until early April, apart from diurnal thawing of the upper few centimetres", they present no evidence in support of this statement and it conflicts with winter observations made at 800-950 m on An Teallach, where pits dug during the months cited occasionally revealed unfrozen or shallow frozen ground under snowcover. Whereas it is probable that high ground may sometimes remain frozen for several months (particularly if ground deep-frozen in November or December is subsequently protected from thaw by snowcover), instances of true "annual" subsurface freeze-thaw cycles in upland Britain have not been demonstrated.

5.4 Precipitation

In upland Britain precipitation increases with altitude, but the rate and nature of this increase are poorly documented. The 1:625,000 map of average annual precipitation for the period 1941-1970 (Meteorological Office, 1977) gives a general indication of the distribution and amount of precipitation on high ground (section 2.6) but as the isohyets represent estimated rather than measured precipitation, figures abstracted from the map must be treated with caution, particularly as precipitation gradients appear to vary greatly from area to area. For example, Unwin (1969) gave a gradient for 4.58 mm m^{-1} for Snowdonia, a figure that greatly exceeds Gloyne's (1958) estimates of 2.53 mm m^{-1} for west slope gradients and 0.83 mm m^{-1} for east slope gradients in the Scottish Highlands. Moreover, although these and other authors (e.g. Rodda, 1962; Taylor, 1976) provided data that indicate a roughly linear increase of precipitation with altitude, Harrison (1973) demonstrated that the gradient in some areas may be curvilinear and approximately exponential. This section describes the relationship between precipitation and altitude on An Teallach and the general characteristics of precipitation in the area from An Teallach to Ben Wyvis.

Daily measurements of precipitation at a gauge sited at 23 m near Dundonnell (at the foot of An Teallach) have been made since 1962. For the 15 y period 1964-1978 the mean annual precipitation recorded at this site was 1732 mm, ranging from 1292 mm (1972) to

2232 mm (1967). Precipitation at Dundonnell exhibits a pronounced winter maximum with mean monthly precipitation exceeding 200 mm in October, November and December yet falling below 110 mm during the period May to September. The primary cause of this pattern is increased cyclonic activity in winter.

During the period from 1 September 1976 to 31 August 1979 measurements of precipitation were made at four to six week intervals at three sites on An Teallach, two (at 605 m and 670 m) near to the weather stations on Mheall Garbh and at the head of Coire a'Mhuilinn and the third at 465 m (NH 110862). Precipitation was collected in standard 5" diameter Negretti-Zambra octapent gauges that were located at sites sheltered from westerly winds. When snow covered the gauges measurement was postponed until it melted. Only during the period January-May 1979 did postponement result in a gauge overflowing (at 670 m) with consequent loss of record.

The measurements made on An Teallach were adjusted in proportion to daily precipitation figures for Dundonnell to reconstruct the monthly precipitation totals at each gauge. The results (figure 5.8) highlight a marked increase in both precipitation and range of monthly precipitation with altitude, the latter being 23.6-414.8 mm at Dundonnell, 29.1-523.7 mm at 465 m, 30.8-678.4 mm at 605 m and 55.0-930.0 mm at 670 m.

The relationships between precipitation at the An Teallach sites and that at Dundonnell were assessed in three ways. First, the precipitation total for each An Teallach gauge during the measurement period was divided by the Dundonnell total. This yielded

$$P_{465} = 1.282\ P_o$$

$$P_{605} = 1.698\ P_o$$

$$P_{670} = 2.000\ P_o \text{ (excluding January-May 1979)}$$

where P_n is precipitation at altitude n and P_o is precipitation at Dundonnell. The second method involved adjustment of the mean monthly values for each An Teallach gauge to the 15 y monthly means for Dundonnell and summation of these to obtain estimates of 15 y

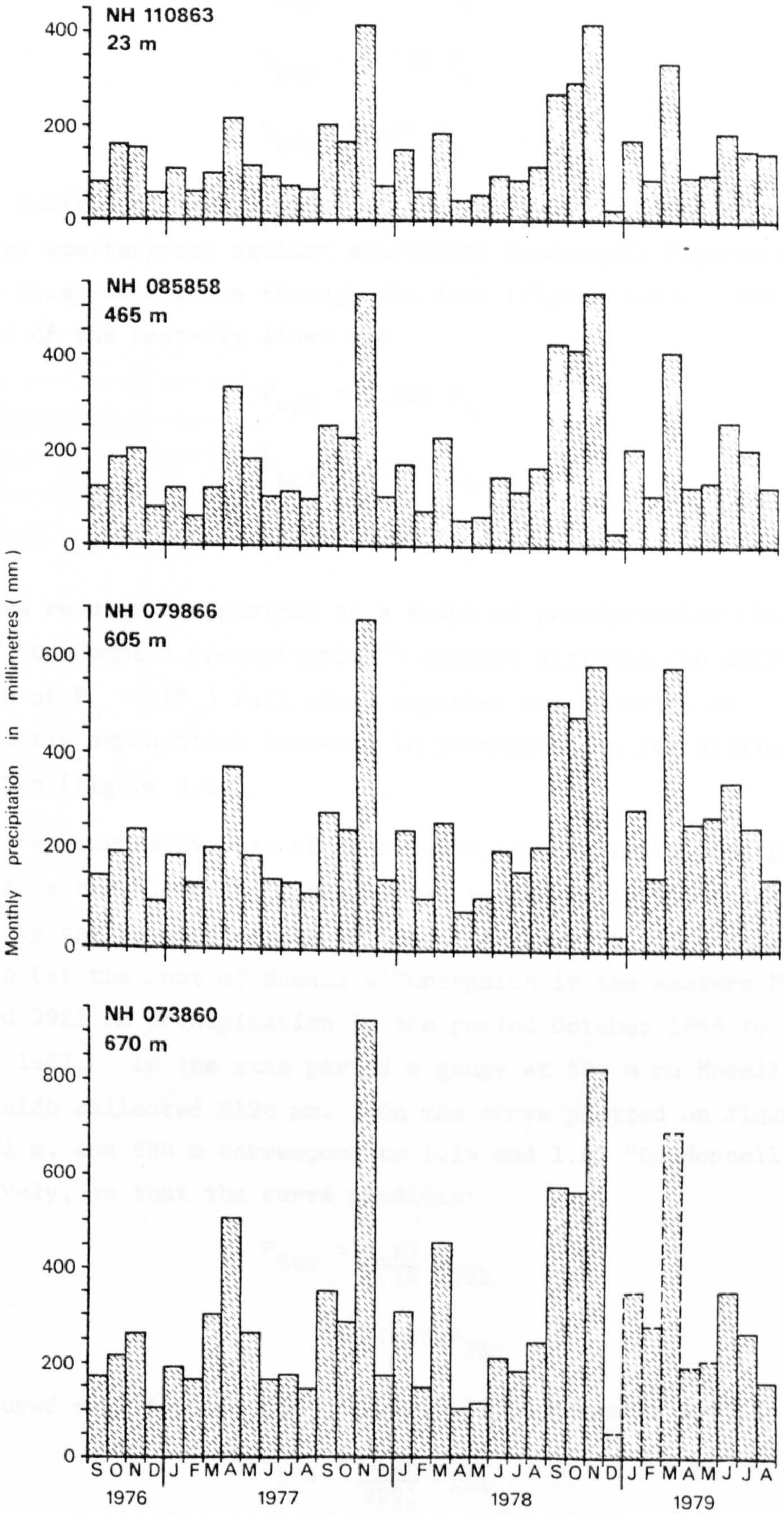


Figure 5.3: Monthly precipitation at 465 m, 605 m and 670 m on An Teallach, apportioned with reference to daily readings at Dundonnell (23 m), September 1976 to August 1979.

mean annual precipitation. Division by the Dundonnell 15 y mean (1732 mm) gave

$$P_{465} = 1.295\ P_o$$

$$P_{605} = 1.726\ P_o$$

$$P_{670} = 2.027\ P_o$$

Finally, individual measurements at the An Teallach sites were plotted as scattergrams against equivalent Dundonnell figures and best fit lines were drawn through the data (figure 5.9). The gradients of the best-fit lines are

$$P_{465} = 1.333\ P_o$$

$$P_{605} = 1.767\ P_o$$

$$P_{670} = 2.087\ P_o$$

When these results are plotted as a graph of precipitation (in units of "Dundonnell precipitation") against altitude the different estimates of $P_n = f(P_o)$ fall close together and describe an approximately exponential increase in precipitation for altitudes up to 670 m (figure 5.10).

The representativeness of this curve in describing precipitation gradients in the Northern Highlands was tested using unpublished data provided by the Meteorological Office. A gauge at 251 m beside Loch a'Bhraoin (at the foot of Mheall a'Chrasgaidh in the eastern Fannichs) collected 3921 mm precipitation in the period October 1965 to November 1967. In the same period a gauge at 594 m on Mheall a'Chrasgaidh collected 6126 mm. On the curve plotted on figure 5.10, 251 m, and 594 m correspond to 1.14 and 1.67 "Dundonnell units" respectively, so that the curve predicts:

$$P_{594} = \frac{1.67}{1.14} P_{251}$$

$$= 1.465\ P_{251}$$

The measured relationship for October 1965 to November 1967 is

$$P_{594} = \frac{6126}{3921} P_{251}$$

$$= 1.562\ P_{251}$$

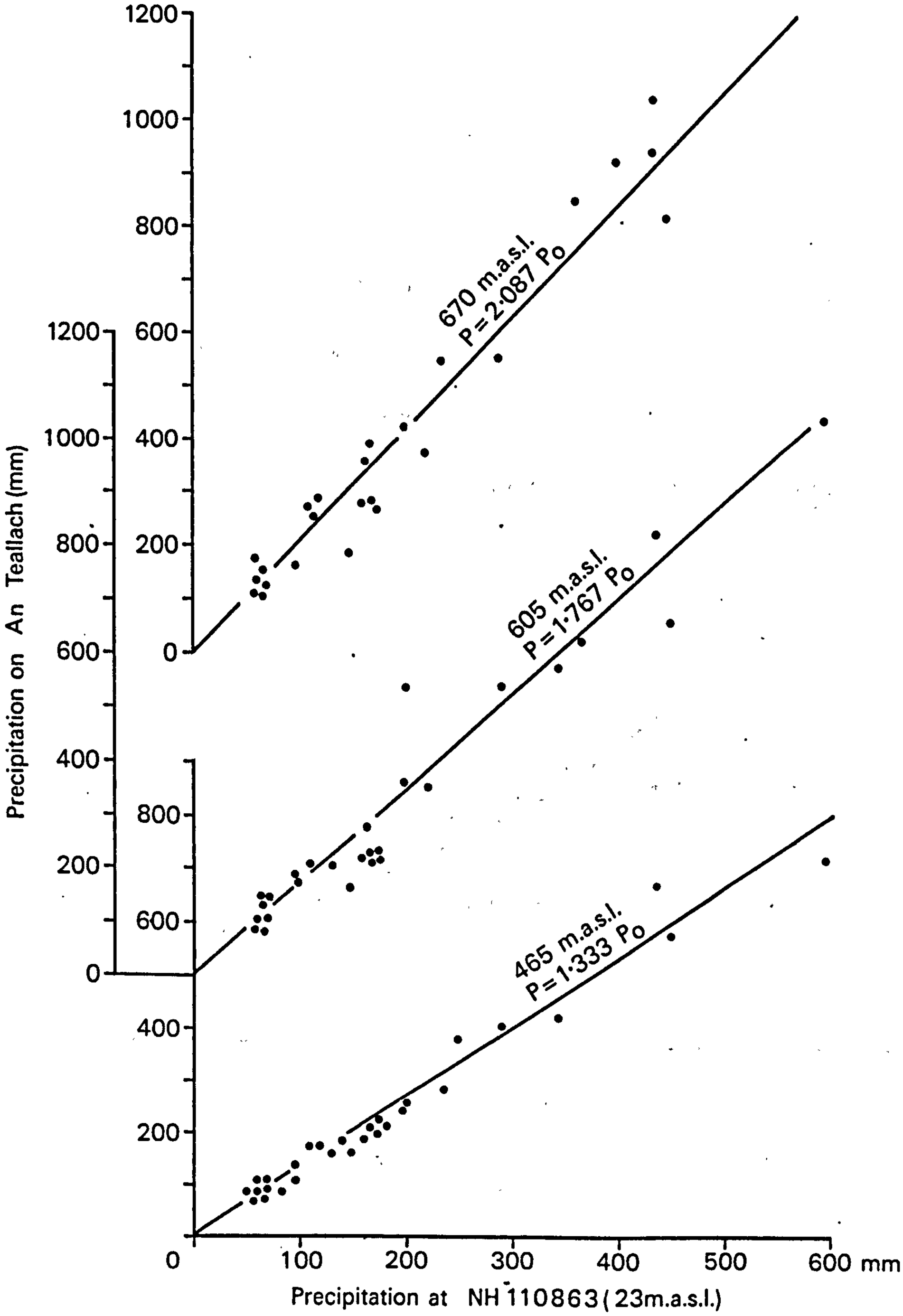


Figure 5.9: Best-fit lines drawn through scattergrams of precipitation readings measured at 465 m, 605 m and 670 m on An Teallach plotted against equivalent precipitation at Dundonnell (23 m).

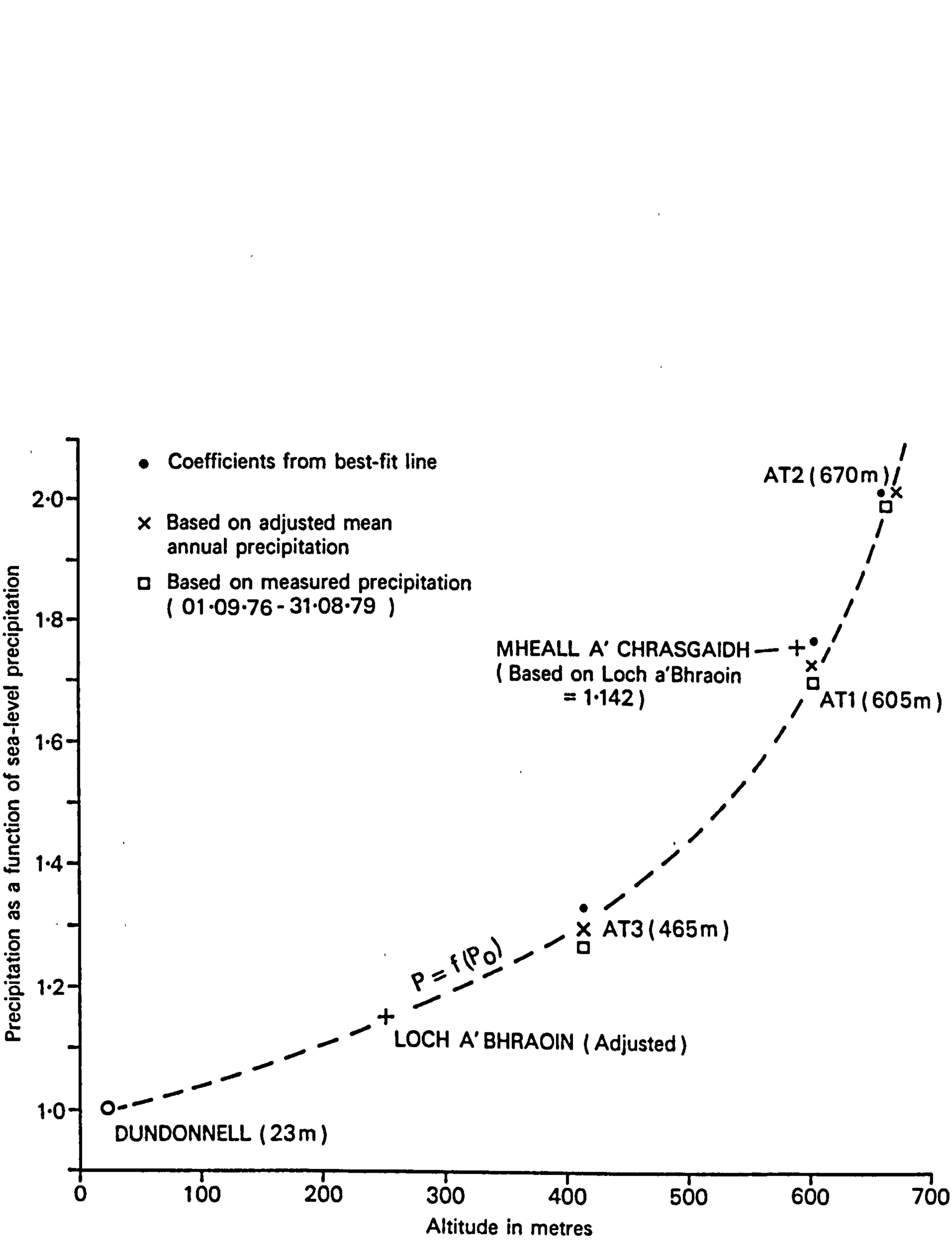


Figure 5.10: The precipitation gradient up to 670 m altitude on An Teallach. For explanation see text.

which although close to the estimate implies an even steeper precipitation gradient for the eastern Fannichs (figure 5.10).

Using this curve, mean annual precipitation at 670 m on Ben Wyvis was extrapolated from unpublished figures for Dubh Choille (351 m) and Loch Glass (219 m), which lie respectively 5 km west and 7 km east of the Wyvis ridge. The former yielded an estimate of 2052 mm y^{-1}, the latter 2200 mm y^{-1}. This compares with an adjusted 15 y mean of 3509 mm y^{-1} for 670 m on An Teallach and measured 5 y mean of 2579 mm y^{-1} for 594 m on Mheall a'Chrasgaidh, equivalent to 3200-3300 mm y^{-1} at 670 m. These figures indicate an eastward decline of precipitation of about 22 mm km^{-2} at 670 m between An Teallach and Mheall a'Chrasgaidh (west of the main watershed) and a steeper decline (c. 41 mm km^{-1}) at the same altitude between Mheall a'Chrasgaidh and Ben Wyvis (east of the watershed).

Details of all available precipitation data for the area between An Teallach and Ben Wyvis are summarized in table 5.2 and figure 5.11. Several points emerge from study of these data. First, the 1:625,000 rainfall map is unreliable in this area as it tends to overestimate mean annual precipitation on low ground and to underestimate (by up to 25% on An Teallach) for high ground. This in turn means that precipitation gradients in this area are much steeper than the map indicates, and suggests that high ground in parts of upland Britain may be considerably wetter than has hitherto been appreciated. Also, the data show that the east-west precipitation decline on high ground is much more marked than on low ground. For example, a drop of c. 1400 mm y^{-1} for 670 m altitude between An Teallach and Ben Wyvis compares with a drop of only c. 520 mm y^{-1} between Dundonnell and Loch Glass.

Finally, the steep precipitation gradient measured on An Teallach suggests that 24-hour rainfall on high ground may reach very high values indeed. Between 1 September 1976 and 31 August 1979 24-hour rainfall at Dundonnell exceeded 50 mm on five occasions (05 10 76, 54.6 mm; 06 05 77, 52.1 mm; 27 10 77, 50.8 mm; 24 11 77, 57.7 mm; 24 06 79, 57.4 mm) and on 4 October 1978 80.1 mm fell. If the precipitation gradient shown in figure 5.10 is valid for short term events these figures imply that falls of 100-120 mm d^{-1}

Table 5.2

Details of raingauge sites

Site	Altitude (m)	N.G.R. (NH)	Period of record (years)	Mean Annual Precipitation(mm) Measured	Isohyets[1]
Dundonnell	23	110862	010976-310879 (3y)	1776	1900
			1964-1978 (15 y)	1732	
An Teallach 1	605	079866	010976-310876 (3y)	3016	2400
			1964-1978 (15y)[2]	2990	
An Teallach 2	670	073860	010976-310876 (3y)	3553	2800
			1964-1978 (15y)[2]	3509	
An Teallach 3	465	085858	010976-310876 (3y)	2277	2400
			1964-1978 (15y)[2]	2243	
Loch a'Bhraoin	251	158752	1963-1967 (5y)	1405	2400
Loch Droma	282	258755	1966-1976 (11y)	1661	1900
Mheall a'Chrasgaidh	594	194735	1965-1969 (5y)	2579	2800
Loch Glass	219	534701	1961-1977 (17y)	1211	1300
Dubh Choille	351	397684	1961-1976 (16y)	1251	1200
Ben Wyvis 1	670	-	1961-1976 (16y)[3]	2052	1600
Ben Wyvis 2	670	-	1961-1977 (17y)[4]	2200	
Strone Nea	366	204843	1954-1978 (25y)	1623	1800
Luibachlaggan	259	350782	1954-1978 (25y)	1533	1700
Vaich Dam	230	347747	1962-1976 (15y)	1332	1500

[1] From Meteorological Office 1:625,000 map, based on period 1941-1970. Approximate.

[2] Adjusted according to Dundonnell 15y monthly means.

[3,4] Simulated from (3) Dubh Choille and (4) Loch Glass figures.

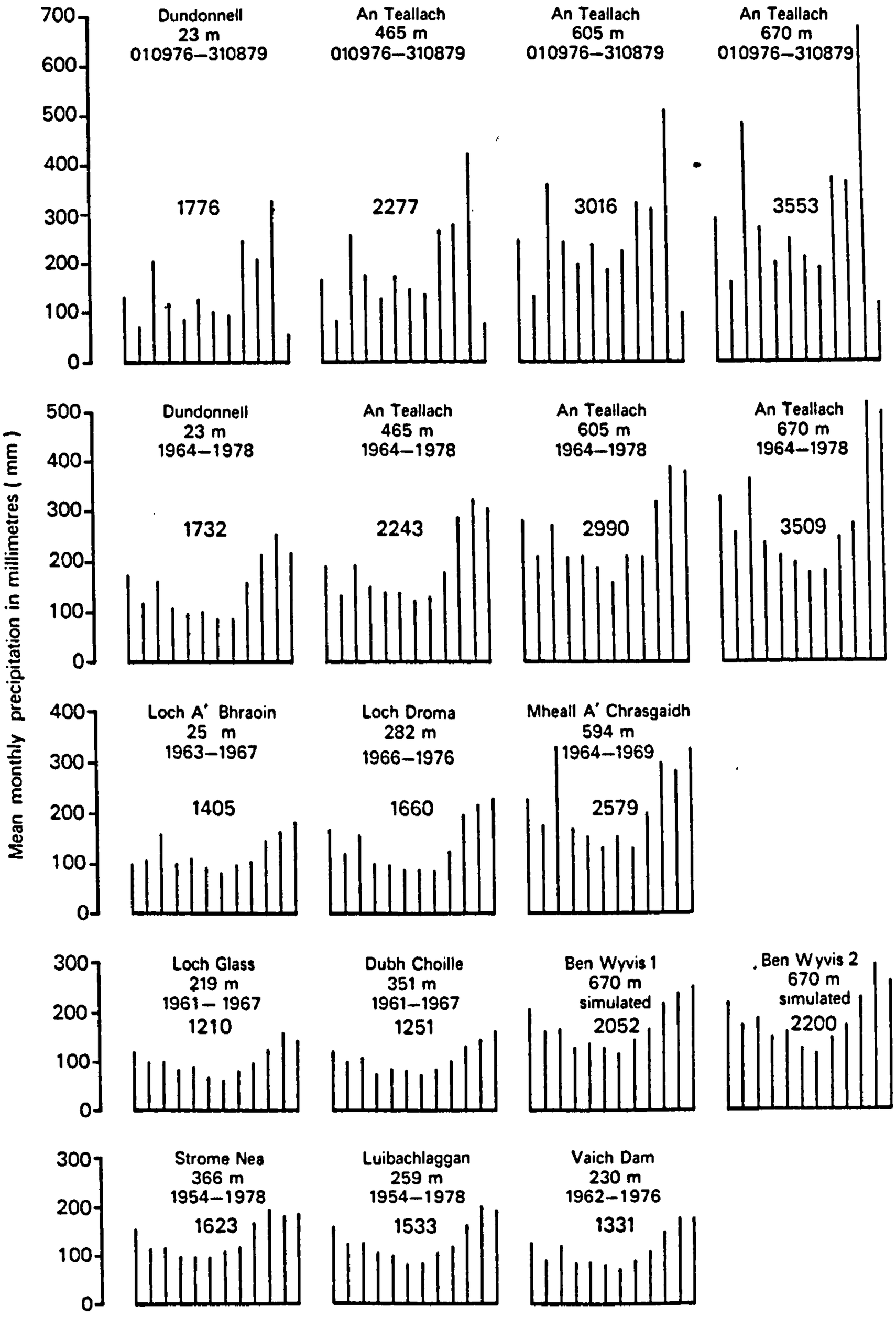


Figure 5.11: Mean monthly precipitation data for sites on both high and low ground between An Teallach and Ben Wyvis. The central figures represent mean annual precipitation in millimetres. The An Teallach figures in the second row are adjusted with reference to the Dundonnell 15 y record.

are not infrequent at 670 m, and that falls of over 160 mm d^{-1} occasionally occur. Under such circumstances it is not surprising that many upland areas display evidence of slope instability and failure triggered by high rainfall (chapter 9).

5.5 Snowcover

Snowcover on An Teallach during the period 1 September 1976 to 31 August 1979 was monitored on behalf of the author by observers at Braehouse Farm, which faces the N.E. flank of the mountain. Each morning they recorded snowcover conditions at the summits of Bidean a'Ghlas Thuill (1062 m), Glas Mheall Mór (981 m), Glas Mheall Liath (962 m) and Meall Garbh (600 m), the northern plateau at 074864 (730 m), a conspicuous rock bench at 088855 (500 m) and the farm itself (50 m) using a four-part scheme of assessment: (i) > 90% deep snowcover; (ii) > 90% shallow snowcover (thin scattering of snow); (iii) large snowpatches (> 50% of ground covered); and (iv) small snowpatches (< 50% of ground covered). Cover on days on which mist obscured some reference points was interpolated with reference to conditions on preceding and succeeding days, change in cover elsewhere, the Dundonnell precipitation record and available upland air temperature records. Most of the results are plotted in figure 5.12; the record from Glas Mheall Liath proved almost identical to that for Glas Mheall Mór and is excluded.

Small late-lying snowpatches and autumn snowshowers excepted, the length of the snow season at all altitudes above 500 m is from November to mid-May, corresponding to the "winter" season defined by air temperatures at 610-665 m (section 5.2). On the highest ground cover was rarely broken during this period (although all three years witnessed a conspicuous thaw in late February-early March), but at 500-600 m cover was frequently interrupted by melting. Snowmelt at 500-600 m often resulted from rain that fell as snow on higher ground, and only after mid-March did insolation cause appreciable melt. It is remarkable that the snowy winter of 1978-9 had little effect on snowcover at altitudes above 730 m despite a marked increase in the duration of snow-lie at Braehouse and to a lesser extent at 500 m and 600 m. This suggests that the duration of snow-lie varies less from year to year on high ground compared with low and intermediate altitudes.

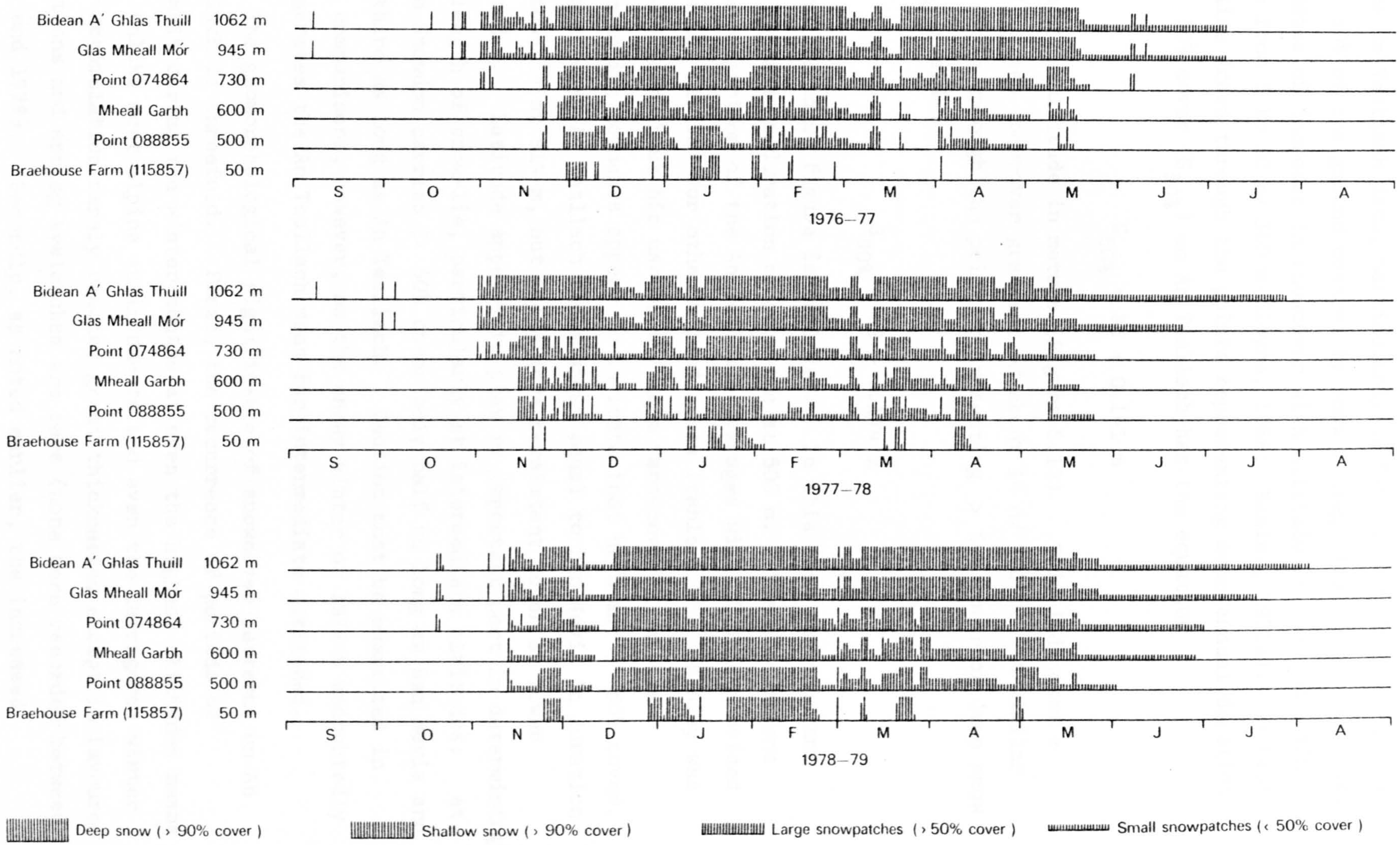


Figure 5.12: Snowcover at various points on An Teallach, recorded daily at Braehouse Farm.

Manley (1949) suggested that the duration of snowcover in Great Britain increases linearly with altitude. This is supported by the An Teallach data, particularly if "snowcover" is interpreted as > 50% of the ground covered by snow (categories (i) to (iii) above). Estimates of increase in snowcover with altitude in Great Britain range from 5 to 20 d/ 100 m (Gloyne, 1968; Manley, 1971a). A best-fit line drawn through the points representing mean annual duration of > 50% cover ($S_{50\%}$) on An Teallach has the equation

$$S_{50\%} = 36 + 0.141\ m$$

where m is altitude in metres (figure 5.13). This indicates a fairly steep snowcover gradient of about 14 d/ 100 m. A similar line drawn through the points representing > 90% cover of deep snow ($S_{90\%}$) yields

$$S_{90\%} = 59 + 0.174\ m$$

if the Braehouse figure is excluded. In this case the constant (59) represents duration of snow-lie at 500 m. These equations allow comparison of the An Teallach averages with figures provided by Manley (1971a) for other upland areas (table 5.3). Manley was not explicit about his use of the term "snowcover", but his accompanying comments appear to indicate that he meant > 50% cover, in which case An Teallach ranks about equal to Ben Nevis in duration of snow-lie at 914 m, but has a more persistent cover at lower altitudes. Latitude appears to play an important part in determining the length of snow-lie, particularly at intermediate altitudes: at 457 m Snowdon carries > 50% cover only half as long as Ben Nevis and one third as long as An Teallach. Caution must be exercised in such comparisons, however, as the snowy winter of 1978-9 undoubtedly exaggerates the An Teallach mean for intermediate altitudes.

The geomorphological significance of snowcover patterns on An Teallach is threefold. First, the occurrence of periods of snowmelt during the winter months at even the highest altitudes means that unlike most alpine environments and even the Cairngorms winter snow accumulation rarely reaches great thicknesses except in favoured locations and spring avalanches are rare (none were recorded between 1976 and 1979). Secondly, as noted earlier, the increased

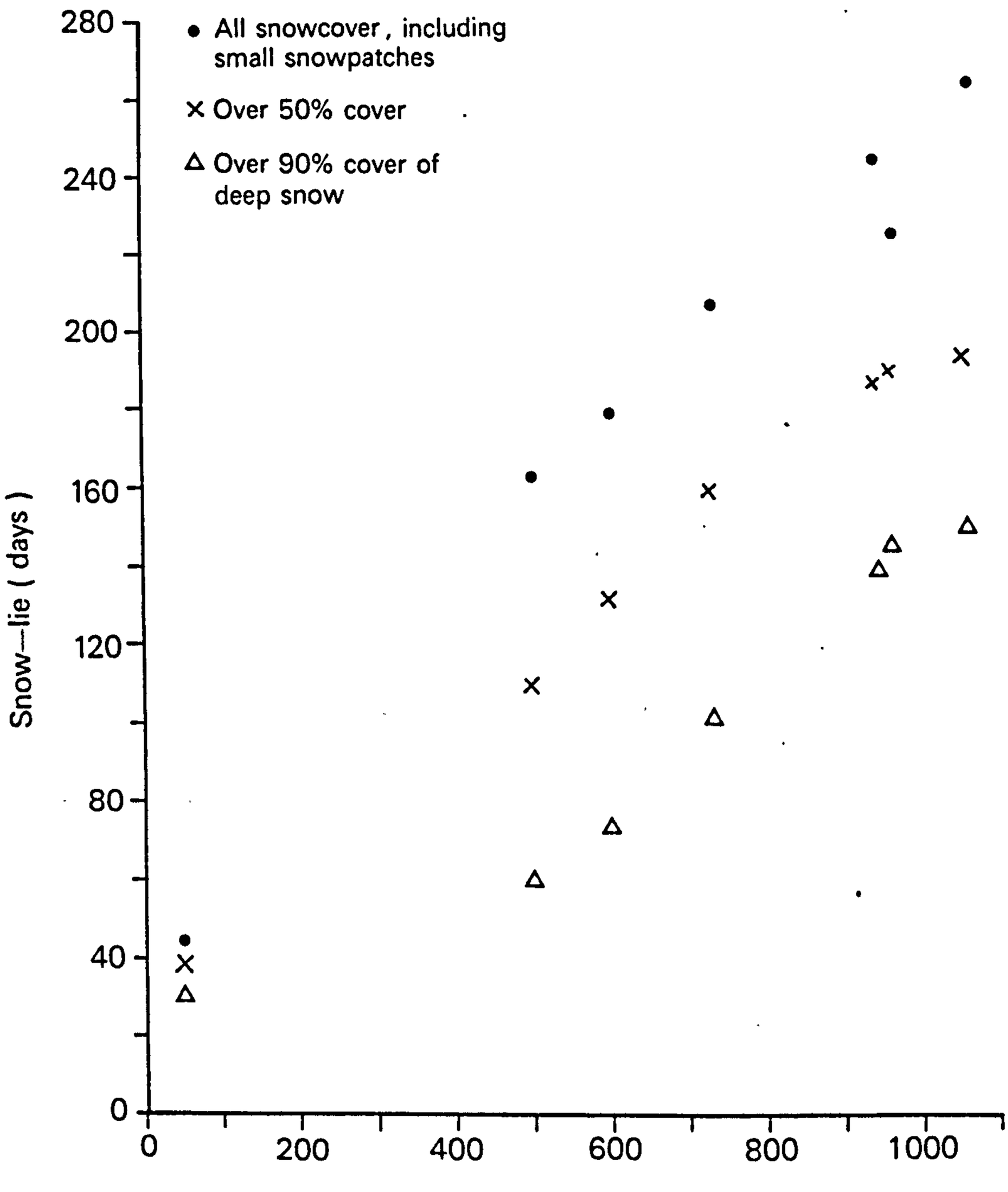


Figure 5.13: Snowcover gradients on An Teallach, based on data for the period 1976-1979.

Table 5.3

Annual duration of snowcover on An Teallach and some other mountain areas

Altitude:	1500' 457 m	2000' 610 m	2500' 762 m	3000' 914 m	3500' 1067 m
An Teallach 1976-9					
> 50% cover[1]	100	122	143	165	(186)
> 90% cover[2]	52	78	105	131	(158)
Glen Lyon Hills[3]	75	100	125	150	
Ben Nevis[3]	66	103	139	165	202
Snowdon[3]	32	44	54	72	

Notes:

1 Based on $S_{50\%} = 36 + 0.141\ m$ (see text)

2 Based on $S_{90\%} = 59 + 0.174\ m$ (see text)

3 From Manley, 1971a, p. 197. Manley apparently used the term "snowcover" to refer to > 50% cover.

persistence of extensive cover on high ground militates against frequent frost penetration of the ground, so that the number of effective freeze-thaw cycles (particularly diurnal cycles) is likely to be greater at intermediate altitudes. Thirdly, when winter and spring thaws occur the prolonged survival of snowpatches ensures that the ground will continue to receive a supply of nival meltwater that may result in surface wash (chapter 12) and, given renewed freezing, will promote more effective frost action.

5.6 Wind

Spot measurements of wind strength and direction at eight sites on An Teallach were made as frequently as possible using a hand-held anemometer and compass, strength being assessed as the average of six readings taken at 30-second intervals. No more than one reading per day was made at any site, but even so the data (summarized, with details of the sites, in table 5.4) cannot be treated as representative of "average" conditions as more readings were made in summer than winter and as the higher parts of the massif were rarely visited on stormy days. The data nonetheless provide insights into variations in wind strength and direction with altitude and aspect.

The influence of altitude on wind strength is partly obscured in table 5.4 as some sites (3, 6 and 7) occupy relatively sheltered locations and consequently experience fewer strong winds than exposed sites on lower ground. However, comparison of the data for Dundonnell with those for exposed sites at 610 m, 800 m and 950 m (figure 5.14) clearly demonstrates the changing shape of the wind speed frequency distribution with altitude, with an increase in the frequency of winds over 10 ms^{-1} (= 36 km h^{-1} = 22.4 m.p.h.) from 2% at 15 m to 27% at 610 m, 30% at 800 m and 33% at 950 m. More importantly, winds about or exceeding 30 ms^{-1} were recorded at all three exposed sites, with individual gusts exceeding the range of the anemometer (40 ms^{-1}). Given the limitations of spot sampling and the bias of the data towards calmer conditions, it seems likely that winds at exposed sites above 600 on An Teallach reach speeds at least equal to and probably exceeding those recorded at Cairngorm summit (1245 m; table 5.5).

Figures 5.14-5.16 highlight the dominance of westerlies and the influence of topography on wind strength and direction. The high

Table 5.4

Summary of measurements of wind strength, An Teallach

Site	N.G.R. (NH)	Altitude (m)	Sample	Percentage frequency of winds: > 5 ms^{-1}	>10 ms^{-1}	>15 ms^{-1}	> 20 ms^{-1}
1	088881	15	131	23	2	1	0
2	081869	610	107	56	27	15	2
3	072867	690	71	46	10	1	0
4	065863	800	63	54	30	18	5
5	064860	700	45	42	18	4	0
6	073860	670	68	32	2	1	0
7	085858	465	38	10	0	0	0
8	079865	950	33	55	33	9	6

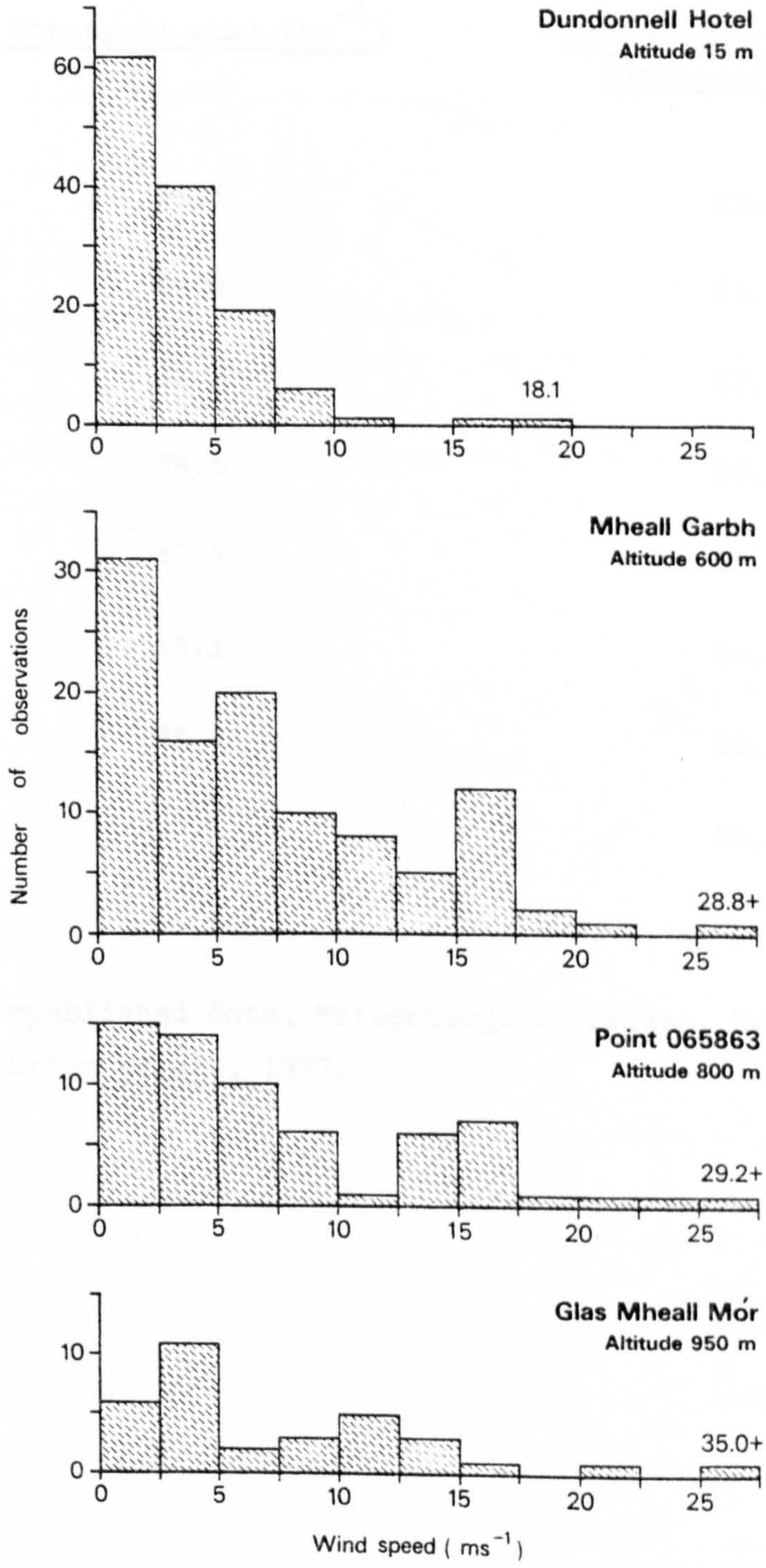


Figure 5.14: Wind speed frequency distributions for Dundonnell Hotel and exposed locations at 600 m, 800 m and 950 m on An Teallach. Each observation represents the average of six anemometer readings made at 30 second intervals.

Table 5.5

Strongest gusts and highest mean hourly wind speeds at the summit of Cairngorm (1215 m)

Year	Strongest gust (ms^{-1})	Highest mean hourly windspeed (ms^{-1})
1966	46.5	33.0
1967	64.4	32.5
1968	52.7	32.5
1969	54.5	33.0
1970	51.4	31.0
1971	55.1	34.0
1972	56.8	32.0
1976	47.9	36.7

Sources: Unpublished data, Meteorological Office, Edinburgh.
Curran et al., 1977.

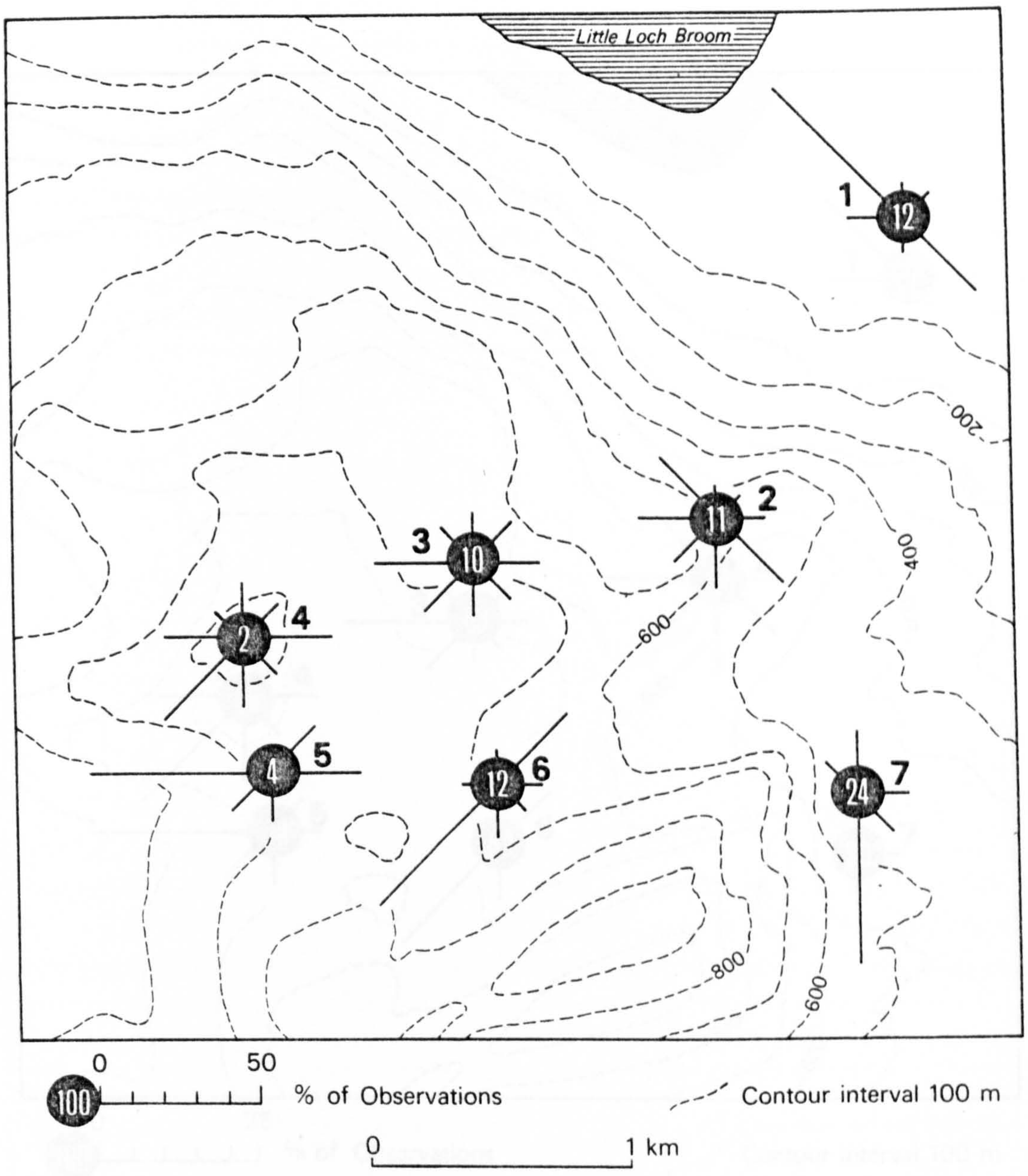


Figure 5.15: Wind roses depicting the frequency of winds from different directions at seven points on An Teallach. The figures in the circles represent the percentage frequency of calms.

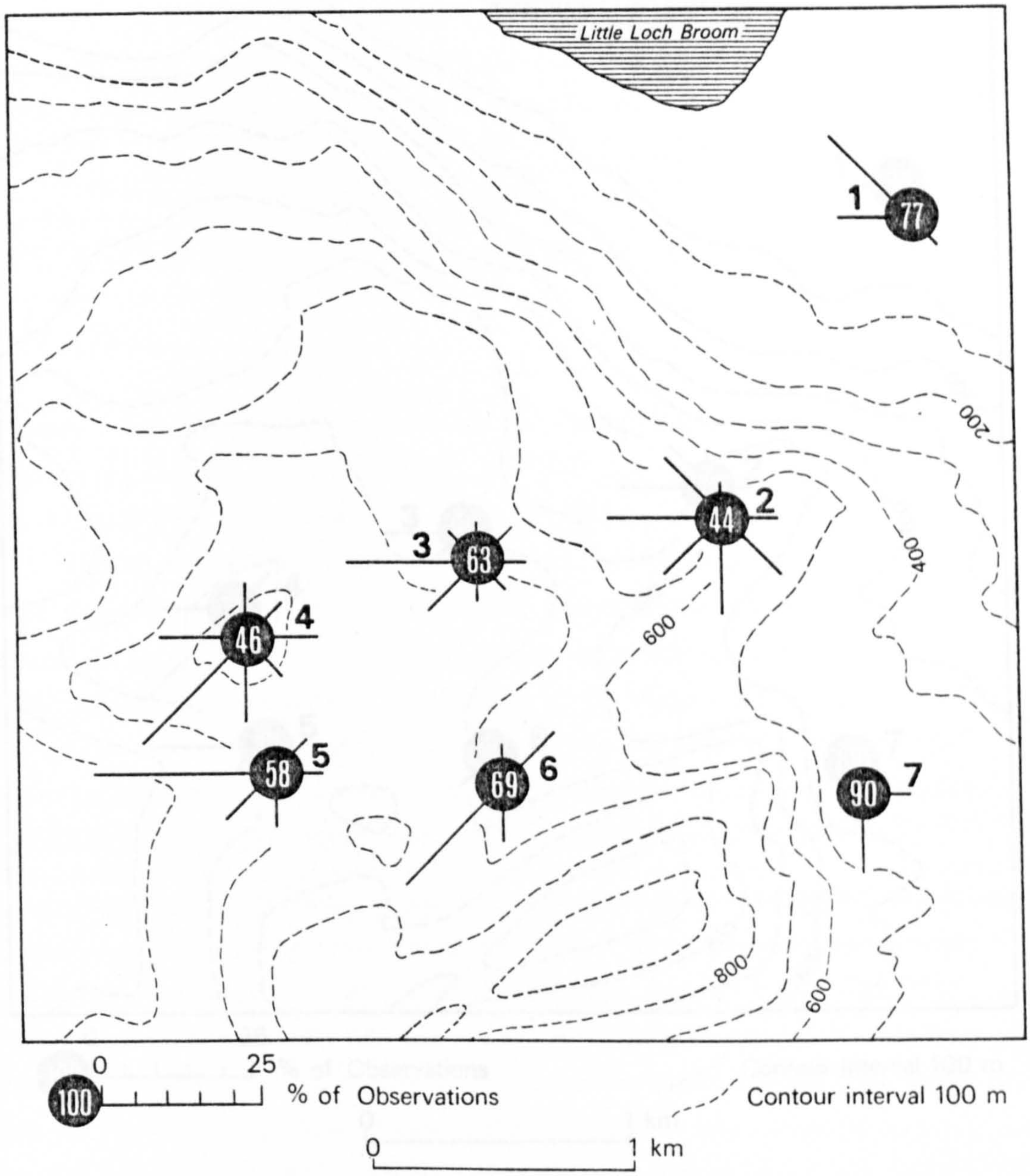


Figure 5.16: Wind roses depicting the frequency of winds exceeding 5 ms^{-1} from different directions at seven points on An Teallach. The figures in the circles represent the percentage frequency of winds of less than 5 ms^{-1}.

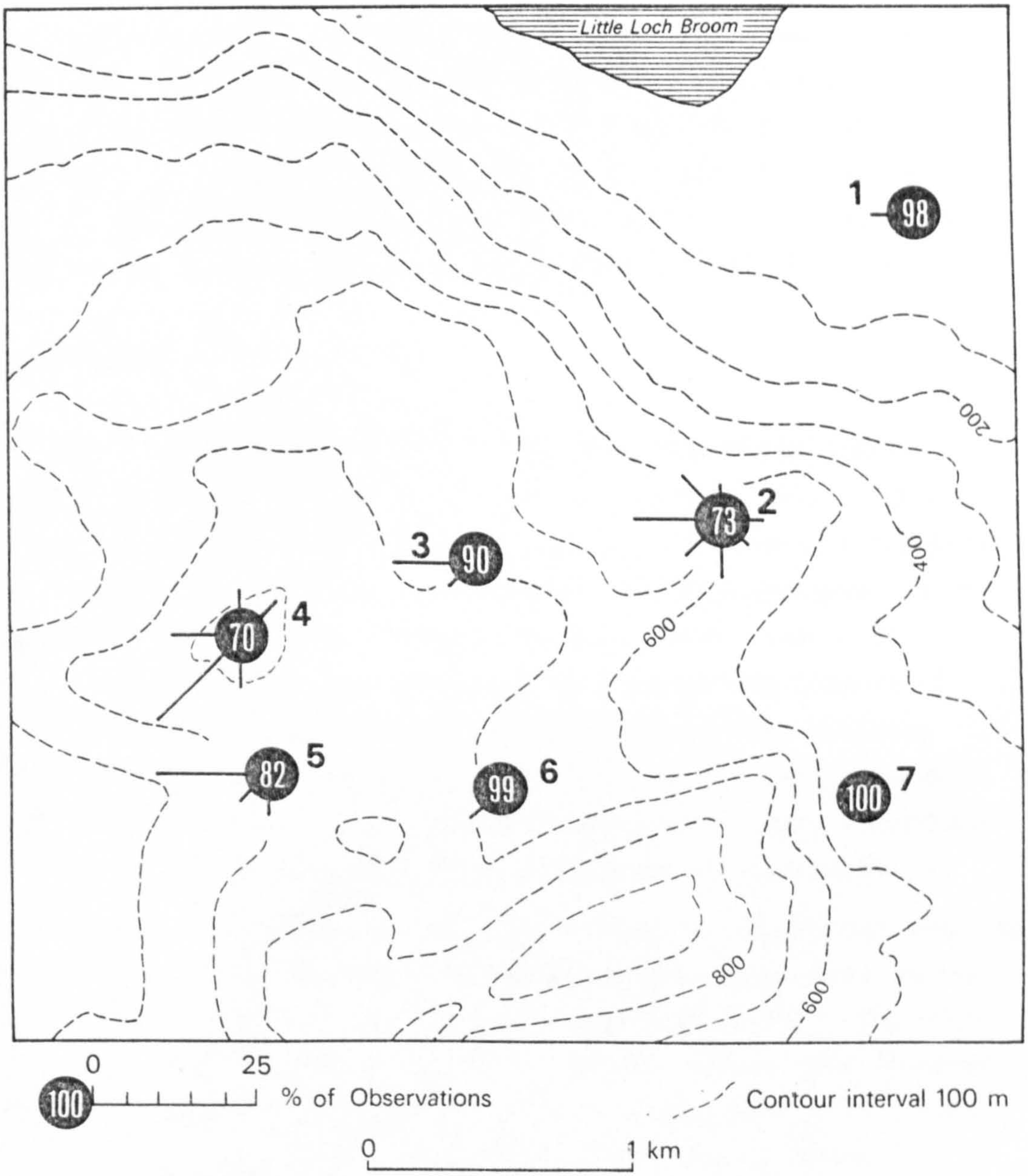


Figure 5.17: Wind roses depicting the frequency of winds exceeding 10 ms^{-1} from different directions at seven points on An Teallach. The figures in the circles represent the percentage frequency of winds of less than 10 ms^{-1}.

frequency of westerly winds is apparent even when winds of all strengths are considered (figure 5.15) although locally these are deflected around the mountain mass, approaching Dundonnell from the north-west and site 7 from the south. There is also a pronounced secondary mode representing east to south-easterly breezes. At site 6 the westerlies are funnelled through the valley of the Allt a'Mhuilinn and approach from the south-west, and some of the north-easterly component here is due to eddy effects as north-easterly gusts were recorded here on days when winds were blowing from the west. The dominance of westerlies is even more apparent when only winds > 5 ms^{-1} are considered (figure 5.16), and few winds other than westerlies had strengths exceeding 10 ms^{-1} (figure 5.17). A further striking feature of figure 5.16 is the absence of "deflected" westerlies at sites 1 and 7. All winds with recorded speeds in excess of 20 ms^{-1} came from between 210° and 300°.

5.7 Discussion

Perhaps the strangest feature of the "maritime periglacial" environment of upland Britain is its lack of extreme cold. The net balance of ambient air temperatures is positive, and even in midwinter severe freezing does not occur. Moreover, the data presented above indicate that even when air temperatures reach their lowest, the effect on the ground is often minimized by a protective blanket of snow that shields the surface completely from short-term freezing and restricts frost penetration to a few tens of centimetres during prolonged cold spells. Under such circumstances it seems remarkable that active periglacial landforms are widespread on British hills.

The present effectiveness and distribution of periglacial activity in upland Britain only becomes comprehensible when considered in the light of the precipitation and wind data presented above. The high precipitation recorded during the winter months ensures that whenever freezing does penetrate the ground moisture is abundant so that such effects as frost heave, frost weathering, frost creep, needle ice action and ice lens formation are maximized. The role of the high winds recorded on exposed summits and plateaux is even more important: such winds strip away vegetation and snow cover and thereby expose the surface to frost, rain and deflation. The direct effect of wind

on the upland landscape is evident in a wide range of landforms such as deflation surfaces, turf-banked terraces, wind stripes, sand scarps and sand deposits (chapters 8-11) but its indirect effect in exposing the ground surface to the action of frost is arguably the single most important climatological factor influencing - perhaps even permitting - the present-day periglaciation of upland Britain.

CHAPTER 6 PERIGLACIAL WEATHERING

6.1 Introduction

The regolith that mantles mountain slopes and plateaux in Scotland is largely composed of detritus weathered from the underlying bedrock, and the nature of this detritus is crucial in determining the mode and rate of geomorphological activity and hence the type, nature and morphology of resulting periglacial landforms. In consequence, geological boundaries often mark a transition from one type of periglacial landscape to one that is completely different. On An Teallach, for example, the unconformity that separates the Torridon Sandstone strata from those of Cambrian Quartzite also marks an abrupt change from the slab-covered deflated plateaux of the former to the chaotic openwork blockfields of the latter. Similarly, the basalt of Orval, in western Rhum, has weathered to give a completely different periglacial landscape from that of the microgranite on the neighbouring plateau of Sron an t-Saighdeir: the former is covered by a shallow soil of pebbles and fines in which small active stone polygons have developed, whereas the latter is buried under a blockfield of subangular boulders that form large fossil stone stripes, polygons and lobes. Such contrasts between adjacent areas can only be attributed to differences in the way in which periglacial weathering has affected different rock types.

The examples cited above indicate that periglacial weathering plays a fundamental role in determining the characteristic landforms to be found on any given rock type. Periglacial weathering is, however, equally important as an independent phenomenon, hitherto little-studied with regard to the mountain landscapes of Great Britain. The present chapter contains a discussion of the characteristics of past and present periglacial weathering in the study area, prefaced by a review of the theory of periglacial weathering and a summary of previous observations of periglacial weathering in Scotland.

6.2 The theory of periglacial weathering

6.2.1 Definitions

The term "periglacial weathering" is virtually synonymous with "frost weathering", which may be defined as the breaking up of consolidated rocks as a result of repeated temperature fluctuations about 0°C. Many different terms have been used to describe the operation of frost weathering, for example frost wedging (Washburn, 1979), frost splitting, congelifraction (Bryan, 1946), frost riving, frost blasting, frost cracking (Antevs, 1932), frost shattering, frost bursting (Dahl, 1966) and gelivation. This multiplicity of terms, however, masks a lack of understanding of the underlying processes, for although geomorphologists have long been aware of the dramatic effect of frost weathering on arctic and alpine landscapes, the mechanics of frost weathering are still incompletely understood (Ives, 1973; Washburn, 1973; White, 1976a). Tricart (1956) subdivided the process into "microgélivation", which he considered to operate independently of rock texture, producing material of silt size or finer, and "macrogélivation", which exploits structural and textural weaknesses in rocks (joints, bedding planes and intergranular voids) to produce coarse grains ("macrogélivation granulaire" or stones and boulders. In the present account these terms are redefined as follows: "microgelivation" is employed to refer to frost weathering that produces particles (grains, crystals or aggregates) generally less than 32 mm (-5ϕ) in length, or flakes not exceeding a few millimetres in thickness; and "macrogelivation" refers to frost weathering that produces larger fragments (pebbles and boulders).

6.2.2 Field Studies

The efficacy of frost weathering in breaking down bedrock to produce coarse angular debris on upland surfaces and slopes in both Pleistocene and present-day periglacial environments has long been recognised (Daly, 1905; Svenonius, 1909; Crampton and Carruthers, 1914; Högbom, 1914; Antevs, 1932), and most field

studies of the phenomenon relate to the problem of blockfield formation (Chapter 8). In general, field workers have attributed the production of both fine and coarse detritus to frost wedging, the prising apart of rocks through the freezing of water in pores or secondary geological structures such as joints or bedding planes (Washburn, 1979; French, 1976), although observations of the actual operation of frost wedging in the field are lacking. Griggs (1936) has, however, described evidence that frost wedging may operate very rapidly on the surface layers of freshly-exposed bedrock. Several studies have shown that the size range of weathered products is great, from "house-sized boulders" (Washburn, 1973) to fine sand, silt or even clay (Klatka, 1961, 1962; Caine, 1968) although chemical weathering may be partly responsible for the production of the last-mentioned.

Various field studies have demonstrated great differences in the susceptibility of different rocks to frost weathering. Bird (1953) pointed out that the limestones of Southampton Island (Canadian Arctic) are severely shattered, whereas the adjacent crystalline rocks are hardly affected, and the author noted similar differences on Ellesmere Island, where well-bedded sandstones have broken down into angular slabs whilst nearby outcrops of massive dolomite have apparently escaped frost weathering, and indeed display solution runnels (rillenkarren) and pitting that betoken effective chemical weathering. Some authors have, on the basis of field studies, ranked rocks according to susceptibility to frost weathering (Alexandre, 1958; Waters, 1964; St. Onge, 1969), but it is unlikely that their findings have more than local applicability.

6.2.3 Experimental studies

Few geomorphological processes can be simulated in the laboratory as successfully as frost weathering, and this is reflected by the extensive literature on this topic. The earliest experiments were carried out by Högbom (1899) but, with few exceptions, most research on experimental frost weathering postdates a classic study by Tricart (1956) that established a

tradition of such work in France (Lautridou, 1971b) and a methodology that has been widely imitated (e.g. Wiman, 1963; Martini, 1967; Potts, 1970). As White (1976a) has pointed out, however, all of the work to date has been concerned with microgelivation, in the sense defined above.

It is difficult to draw detailed conclusions from the literature, partly because of methodological variations, but also because of the wide range of behaviour that may be displayed by similar rocks, particularly when these are of varied provenance. A number of generalizations can be made, however, concerning: (i) conditions under which microgelivation is most effective; (ii) properties inherent in rocks that determine their susceptibility to microgelivation; (iii) vulnerability of different rock types; and (iv) nature of weathered products.

Experimenters are unanimous in agreeing that microgelivation is ineffective in the absence of water (Tricart, 1956; Masseport, 1959), and various studies have shown that rate of microgelivation increases with the amount of free water surrounding rocks at the time of freezing (Wiman, 1963; Martini, 1967; Guillien and Lautridou, 1970; Mellor, 1970; Potts, 1970). This indicates that the differential expansion and contraction of different minerals is not of great importance, and that microgelivation is likely to be most effective in humid periglacial environments. There is less agreement as to the type of temperature fluctuations under which microgelivation is most effective. Tricart (1956) found that 96 h cycles with a wide temperature range ("Siberian" cycles) were more effective than 24 h cycles with a relatively narrow range ("Icelandic" cycles), but these conclusions have been challenged by Wiman (1963) and Potts (1970). Summarizing the results of several years of research, Lautridou (1971b) concluded that freezing to -5°C, -10°C, -15°C and -30°C all produced similar results, although only when the freezing period was of sufficient length. Thomas (1938) showed that rocks vary in their reactions to an increased rate of freezing, but considered that, in general, rapid freezing is most effective in causing microgelivation. Similar conclusions were reached by Battle (1960) and Mellor (1970).

There is agreement that previously-weathered rocks disintegrate rapidly in the course of freeze-thaw experiments (Martini, 1967; Lautridou, 1971b), and this explains great variations in the susceptibility of different samples of the same rock type (Tricart, 1956). Wiman (1963) drew attention to the importance of planes of weakness that he termed "S-planes" or "shatter-planes", along which rocks tended to break up. These appear to be important in the microgelivation of schists (Coutard and Lautridou, 1971), in which belts of horizontally-oriented mica form "planes of fissility" (Washburn, 1973), and in well-bedded sedimentary rocks (Martini, 1967; Potts, 1970; Coutard and Lautridou, 1971; Lautridou, 1971b; Mainguet-Michel et al., 1971). In general, however, the two most important factors that determine the susceptibility of a rock to microgelivation appear to be the strength of the bonds between individual grains and crystals, and the porosity of the rock. Tricart (1956) noted that feebly-consolidated rocks are particularly susceptible, Martini (1967) that calcareous cement is weaker than siliceous cement; later authors have concluded that bond strength is the most important inherent quality in determining rock resistance to frost weathering (Lautridou, 1971b; Mainguet-Michel et al., 1971). Porosity (void space) has been shown by several investigators to correlate positively with rate of rock disintegration, although the relationship is complicated by variations in cement strength, such that "strong" rocks with high porosity may undergo limited disintegration (Wiman, 1963). This relationship is strongest for rocks of a single type (Martini, 1967; Kieffer and Lautridou, 1971; Lautridou, 1971a) and may be poor or absent when tested for samples including several rock types (Tricart, 1956; Potts, 1970). The relationship is sometimes nonlinear: Coutard and Lautridou (1971) showed that schists with 2-4% porosity undergo little disintegration, whereas those with 5% porosity are often rapidly destroyed. These results again emphasise the importance of high water content in promoting microgelivation.

Generalizations concerning the susceptibility of different classes of rock are conditioned by the inherent properties discussed above. All three main classes of rock show great variations.

In general, sedimentary rocks such as shales and sandstones are relatively susceptible, unless they have low porosity; igneous rocks such as granites and lavas are resistant unless they have been chemically weathered; and metamorphic rocks, especially those that have undergone repeated recrystallization, are the most resistant, with low-porosity schists, gneisses and quartzites often proving least susceptible to microgelivation (Tricart, 1956; Wiman, 1963; Martini, 1967; Potts, 1970; Coutard and Lautridou, 1971).

Although microgelivation can apparently operate on fragments as small as one micron (McDowall, 1960), and occasionally produces clay-size material (Guillien and Lautridou, 1970), most rocks rarely yield fragments of less than 6 μm (Dylik and Klatka, 1952), and experimental studies have shown that many rocks give only small quantities of detritus of less than 60 μm (Masseport, 1959; Wiman, 1963; Martini, 1967; Potts, 1970; Mainguet-Michel et al., 1971). The size of the weathered particles is often largely determined by the size of grains and crystals in the parent rock (Hopkins and Sigafoos, 1951; Martini, 1967).

6.2.4 Mechanisms

Frost weathering has traditionally been attributed to stresses set up by the expansion of water that freezes in joints, bedding planes and intergranular voids. Bridgeman (1912) showed that freezing at 0°C was accompanied by a 9% increase in volume, and that a maximum tensile stress of 2,115 kg cm^{-2} could be developed at -22°C. The maximum tensile stress that can be withstood by rocks is around 250 kg cm^{-2} (White, 1976a). However, Grawe (1936) pointed out that actual stresses set up by the freezing of interstitial water are likely to be much less than the theoretical maximum: freezing from the outside inwards is necessary to create a closed system (Battle, 1960; Mellor, 1973); air trapped within a rock, being compressible, greatly reduces effective freezing stress; and temperatures of -22°C are common only at high latitudes and altitudes. Battle (1960) considered that, in a crack 1 mm wide and 10 cm deep, freezing of water at -0.3°C would produce pressures of only 70 kg cm^{-2}, and Tricart (1963) suggested

an even lower figure (14 kg cm^{-2}) for effective tensile stress. Taber (1943) proposed that ice crystal growth is more destructive than simple expansion on freezing, a proposition supported by Martini (1967). However, experimental work by Connell and Tombs (1971) produced ice crystal growth pressures of only 0.2 kg cm^{-2}, indicating that this mechanism is less important than simple expansion of water on freezing.

Various authors have contended that the behaviour of unfrozen water in a frozen rock is of critical importance. Helmuth (1960) suggested that osmotic pressure produced by the migration of unfrozen water to the freezing plane causes microgelivation, and more recently Hudec (1973a, b) and White (1976a) have argued that hydration, the pressure of unfrozen adsorbed water on mineral surfaces, is sufficient to cause frost-shattering. Temperature fluctuations cause adsorption and the release of adsorbed water that is drawn between grains by ice nucleation or electro-osmosis (resulting from differences in the thicknesses of adsorbed layers on different minerals), producing stresses of up to 2,000 kg cm^{-2}. The effectiveness of hydration in causing microgelivation is supported by experiments by Falconer (1969), who showed that rocks containing clay minerals on to which unfrozen water is adsorbed are most susceptible to breakdown.

Although the fundamental process responsible for microgelivation remains in question, macrogelivation can only be attributed to some form of large-scale frost-wedging, presumably through expansion of water on freezing, along pre-existing lines of weakness (Washburn, 1979; White, 1976a). Rocks can, however, be split by frost weathering even in the absence of obvious structural weaknesses: Mellor (1970, 1973) has suggested that if saturation by water exceeds 50%, rocks may crack internally, although the mechanism responsible for cracking is unknown (White, 1976a).

6.2.5 Chemical Weathering

Most rates of chemical activity decline with temperature, but several authors have reported small-scale features indicative of

chemical weathering in arctic areas. These include exfoliation structures (Sim, 1962; Czeppe, 1964), tafoni (Cailleux and Calkin, 1963; Cailleux, 1968; Washburn, 1969a; Selby, 1971), oxide rinds (Hill and Tedrow, 1961; Selby, 1971) and carbonate coatings (Washburn, 1969a; Woo and Marsh, 1977), the last three types being largely restricted to arid areas, where frost weathering is inhibited.

Solute concentration in runoff waters can be considered a crude index of the amount of chemical activity in an area. For Northern Sweden, Rapp (1960a) concluded that solutional loss is the principal agent of denudation, but his measurements of solute concentration cannot be considered representative. Low concentrations (generally less than 25 ppm) were measured by Church (1972) in rivers draining metamorphic rocks on Baffin Island. Rivers draining carbonate terrain in periglacial areas carry higher concentrations (Cogley, 1972; Ballantyne, 1975), but although interflow waters in such areas may have a total hardness of over 200 ppm (Woo and Marsh, 1977), solution in the periglacial realm appears to be generally less effective than elsewhere.

In sum, the role of chemical weathering in periglacial environments is uncertain, though generally subsidiary to that of frost weathering, and apparently very dependent on local factors such as rock type and precipitation.

6.3 Periglacial weathering in upland Scotland

The importance of frost as an agent of weathering on the mountains of Scotland was recognised by officers of the Geological Survey in the early years of this century. Harker (1901) attributed the formation of talus slopes on Skye to a period of intensive frost shattering, and the quartzite and conglomerate detritus that mantles the hills of Caithness was considered by Crampton (1911) and Crampton and Carruthers (1914) to be the product of "the splitting action of frost" acting on well-jointed rocks. A similar origin was suggested for the blockfields and blockslopes of Ben Nevis and the Mamores by Bailey and Maufe (1916).

All of these authors attributed the production of coarse frost-shattered debris to a period of intensive frost action that accompanied the downwastage of the last ice sheet.

Several subsequent writers have, however, supported the view that present frost action at high altitudes in Scotland is capable of causing macrogelivation, producing coarse shattered detritus. Thompson (1950) considered frost shattering to be active on high summits in the N.W. Highlands, especially in quartzite areas, and FitzPatrick (1958) described the quartzite cap on Beinn Eighe as being "rapidly comminuted". Godard (1958, 1959, 1965) found evidence of "recent splitting by frost" on the volcanic rocks of Ben Buie and Ben More in Mull, but concluded that frost shattering on crystalline rocks, such as those of Beinn Mheadon (Morvern) was of Lateglacial age. However, Galloway (1958, 1961b) proposed that all coarse mountain-top detritus had been produced by intensive frost action during the Lateglacial period, arguing that the present climate is insufficiently severe to produce the "disruption" that he had observed on high plateaux, such as those of the Cairngorms and Ben Nevis. He also pointed out that the products of macrogelivation are not restricted to high ground in Scotland, as frost-shattered debris occurs in Orkney and Buchan but not, significantly, in the glaciated valleys of the N.W. Highlands. Sissons (1967a, 1977b) subsequently noted a marked contrast in degree of frost weathering between adjacent areas outside and inside the limit of the Loch Lomond Advance, with reference to the Torridon Sandstone rocks of the Applecross Penninsula and the metamorphic rocks of the Glen Moriston area; outside the Loch Lomond Advance limit in these areas the rock has been completely shattered, producing large slabs and blocks, whereas inside the limit frost action has apparently been restricted to granular disintegration and small-scale wedging. This contrast partly explains the lack of frost shattering in the glens of N.W. Scotland noted by Galloway. Further evidence in support of a Lateglacial age for the products of macrogelivation in the Scottish Mountains is provided by Sissons' (1976a) observation that much high-altitude frost shattered detritus is peat-covered.

Although Sissons' evidence indicates that frost riving capable of producing large slabs and blocks has not operated since the Loch Lomond Stadial, some researchers have maintained that smaller clasts are displaced from bedrock under present conditions. For example, King (1968) considered that fragments of schist are being produced by macrogelivation at altitudes of over 900 m in the Cairngorms, Hills (1969) described "freshly weathered" granite pebbles on the same massif, and Ryder (1968) was of the opinion that the peridotite of Ruinsival (524 m) is presently subject to "considerable shattering".

Together, these accounts suggest that the operation of macrogelivation since the end of the Loch Lomond Stadial has been limited, although the untested observations of a few researchers indicate that, on certain rocks, frost riving under present conditions is capable of producing clastic fragments. The possibility that the Lateglacial cold periods were of insufficient length to allow the production of a cover of coarse, frost-shattered detritus, implying that some mountain tops remained ice-free during the Late Devensian glaciation (Ragg and Bibby, 1966) or longer (Romans et al., 1966) is inconsistent with evidence relating to the extent and thickness of the last ice sheet and may be discounted (Sissons, 1974a).

The action of microgelivation on Scottish hills has received much less attention, probably because the products are less spectacular. The granulometry of frost-weathered fines is closely related to the size of the grains or crystals in the parent rock (Hopkins and Sigafoos, 1951; Martini, 1967), and the abundance of such fines reflects the susceptibility of the parent rock to granular breakdown and/or flaking. Quartzite, for example, appears to be generally resistant to microgelivation, producing relatively little fine material (Whyte, 1970), usually of silt size (Crampton and Carruthers, 1914). On other rock types an extensive cover of fine material testifies to the efficacy of microgelivation. King (1968) attributed the widespread coarse sand of the Cairngorms to small-scale frost disintegration of granite, and a similar origin may be inferred for the sand deposits of Rhum, derived from peridotite (Ryder, 1968)

and those of the Torridon Sandstone mountains in the N.W. Highlands (Peach et al., 1913a; Godard, 1965; Sissons, 1967a, 1976a; Ballantyne, 1977). In all these areas the thickest sand deposits result from accumulation by wind, but it seems probable that this sand was initially produced by microgelivation (see chapters 8 and 11). Similarly, the deep stony regolith that mantles the greywacke hills of the Southern Uplands was attributed to frost weathering by Ragg and Bibby (1966), who argued that the low percentage of clays and silts in these deposits indicated that they were the product of mechanical breakdown rather than chemical weathering.

The roundness of the parent rocks often gives an indication of their susceptibility to granular disintegration. Rock outcrops of granite and Torridon Sandstone are commonly well-rounded (Galloway, 1958; King, 1968) whereas the quartzite blocks of Scaraben in Caithness were described by Crampton and Carruthers as "highly angular". This generalization is not universal, however, for a comparison of granite pebbles on the Cairngorms with quartzite debris on Schichallion revealed no significant difference in roundness (Hills, 1969).

Direct evidence for the present operation of microgelivation in Scotland is lacking, although several authors have expressed a belief in its occurrence (e.g. Galloway, 1958; King, 1968) and its present effectiveness is implied in the writings of others (e.g. Ryder, 1968; Hills, 1969). That granular disintegration was operative during the Loch Lomond Stadial is suggested by the results of a study by Shaw (1977) on the roundness and weathering characteristics of boulders inside and outside the Loch Lomond Advance limits in the Eastern Grampians. Although Shaw's results are in some cases equivocal, they suggest that the boulders outside this glacial limit are more rounded and weathered than those inside, which in turn indicates that microgelivation was effective during this period.

From the above discussion, it is apparent that systematic studies of the nature and products of past and present periglacial weathering on the Scottish hills are lacking, despite the

importance of such weathering in determining the nature of periglacial landforms on different rock types. This lack of existing evidence prompted a detailed study of periglacial weathering in the study areas. Observations were concentrated on the effect of periglacial weathering on the Torridon Sandstone of An Teallach, which serves as a standard against which the effectiveness of periglacial weathering on other rock types may be compared. For convenience, microgelivation and macrogelivation are treated separately.

6.4 Microgelivation

6.4.1 Microgelivation on Torridon Sandstone

There is abundant evidence for the operation of granular disaggregation on Torridon Sandstone rocks. The surface grains of some exposed rocks have been loosened and can be removed by rubbing, and loose grains are found *in situ* on rock surfaces. Exposed rock outcrops and slabs are almost invariably rounded, usually well-rounded, a form characteristic of granular disintegration of the surface layers of rock (figure 6.1). In contrast, buried flags and bedrock surfaces are usually angular (figure 6.2). This difference in roundness may be observed for individual flags embedded in sandy deposits: the exposed surfaces of such rocks are generally rounded, the buried portions generally angular, a contrast observed by King (1968) on the granite plateaux of the Cairngorms. Differences in roundness were assessed by measuring the minimum radius of curvature (r) on the principal plane of each of 300 clasts from three different environments (*in situ* weathered clasts; clasts on top of sandy deposits on the plateau surface; buried clasts). Two samples of 50 clasts were measured for each environment, and mean roundness ($\overline{R}$) was assessed using the Cailleux roundness formula (King, 1966) in the form

$$\overline{R} = 1000 \sum_{i=1}^{50} \left[\frac{2r_i}{a_i} \right] /50 \qquad (6.1)$$

where a is length of the principal axis of each clast. The results,



Figure 6.1: The Cambrian Quartzite - Torridon Sandstone contact at 960 m altitude on Glas Mheall Mór, An Teallach, showing the effects on both lithologies of macrogelivation (riving apart of slabs along joint planes and other lines of weakness) and microgelivation (rounding of exposed surfaces).



Figure 6.2: Pit in debris surface at c. 800 m on An Teallach plateau, showing the contrast between well-rounded surface clasts and highly-angular subsurface clasts.

given in table 6.1, were tested for paired samples using the t-test and showed differences between the exposed and buried clasts significant at the .001 level. These differences suggest (i) that microgelivation in the form of granular disintegration has operated only on exposed surfaces, and (ii) that microgelivation largely postdates the deposition of plateau-surface detritus (chapter 8).

The effectiveness of microgelivation in Torridon Sandstone areas is also suggested by the lack of openwork blockfields on plateau surfaces, all clastic detritus being embedded in or resting upon a matrix of locally-derived, poorly-sorted sand up to 80 cm deep, and by the accumulation of blown sand deposits up to 3.8 m deep on sheltered lee slopes, the most likely source of this sand being frost-weathered material on the plateau surface (chapter 11). The accumulation of blown sand on snowfields downwind (east) of the An Teallach plateau suggests that microgelivation is actively producing a supply of granular detritus at present, and there is evidence (chapter 11) that it has done so for much or all of the Flandrian.

Microgelivation in the form of small-scale flaking is apparently restricted to outcrops of fine-grained sandstone, and also appears to be active at present. Judging by the roundness of rock outcrops, the effectiveness of microgelivation is much reduced below an altitude of c. 500 m, although there is no well-defined altitudinal limit to its operation.

6.4.2 Microgelivation on other rocks

Nowhere within the study areas is there evidence for microgelivation as effective or widespread as that on Torridon Sandstone. In part this results from the fact that Ben Wyvis and the Fannichs are extensively covered by a vegetation mat and a stable protective soil that inhibits frost weathering. On Torridon Sandstone a similar protective cover cannot form as sand is apparently blown away from exposed surfaces as soon as it is produced.

Table 6.1

Roundness of periglacially-weathered detritus

Sample	Description	$\bar{r}$	$\bar{R}$
P 1	*In situ* exposed weathered clasts	1.55	131.7
P 2	*In situ* exposed weathered clasts	1.30	102.4
B 2 a	Clasts on plateau surface	1.16	169.9
B 3 a	Clasts on plateau surface	1.02	145.1
B 2 b	Buried clasts	0.22	47.7
B 3 b	Buried clasts	0.22	56.1

$\bar{r}$ = mean minimum radius of curvature on the principal plane of sampled clasts.

$\bar{R}$ = mean roundness of sampled clasts (see text, equation 6.1).

Sample size (n) = 50 in each case.

The metamorphic rocks of the study area vary greatly in their response to microgelivation. Three basic categories were identified in the field:

1. Rocks that had been rounded, like Torridon Sandstone, through granular disintegration. Only the pebbly basal Cambrian Quartzite of An Teallach falls into this category, although it is generally less rounded than Torridon Sandstone (figure 6.1). The coarse sand produced by ~~microgelivation~~ granular disaggregation forms, on plateau surfaces, a shallow matrix in which clastic detritus is embedded. Buried clasts and bedrock are angular. Some siliceous schists, such as those that crop out on An Socach, Ben Wyvis (NH 471682) also exhibit some rounding.

2. Rocks apparently resistant to microgelivation. These include the fine-grained "pipe rock" (quartzite) of the Cambrian sequence that crops out on Sàil Liath, An Teallach (NH 072825) and the massive granulite of Càrn Gorm, Ben Wyvis (NH 455705). Outcrops and exposed boulders are generally subangular, and plateau surfaces are covered by blockfields in which fines are absent at the surface. The greater angularity of subsurface blocks in these blockfields indicates limited operation of microgelivation on the surface. The flaggy quartz-schist at the summit of Mheall a'Chrasgaidh, in the Fannichs (NH 185733), a less micaceous rock than most Moine schists (Peach et al., 1913a) shows similar characteristics, as do some siliceous schists on Ben Wyvis.

3. Rocks with well-developed cleavage. This group includes the abundant mica-schists of the Fannichs and Ben Wyvis. Surface clasts are in general rounder (though often fairly angular) and flatter than buried clasts, and occasionally exhibit flaking along cleavage planes. This suggests that microgelivation is operative, although much less effective than on Torridon Sandstone.

Although present microgelivation is apparently limited on Moine rocks, the protective soil "blanket" that inhibits its operation over wide areas itself appears to be the product of past microgelivation. The lack of clay (figure 6.3) and angular appearance of grains under the microscope indicate that these soils

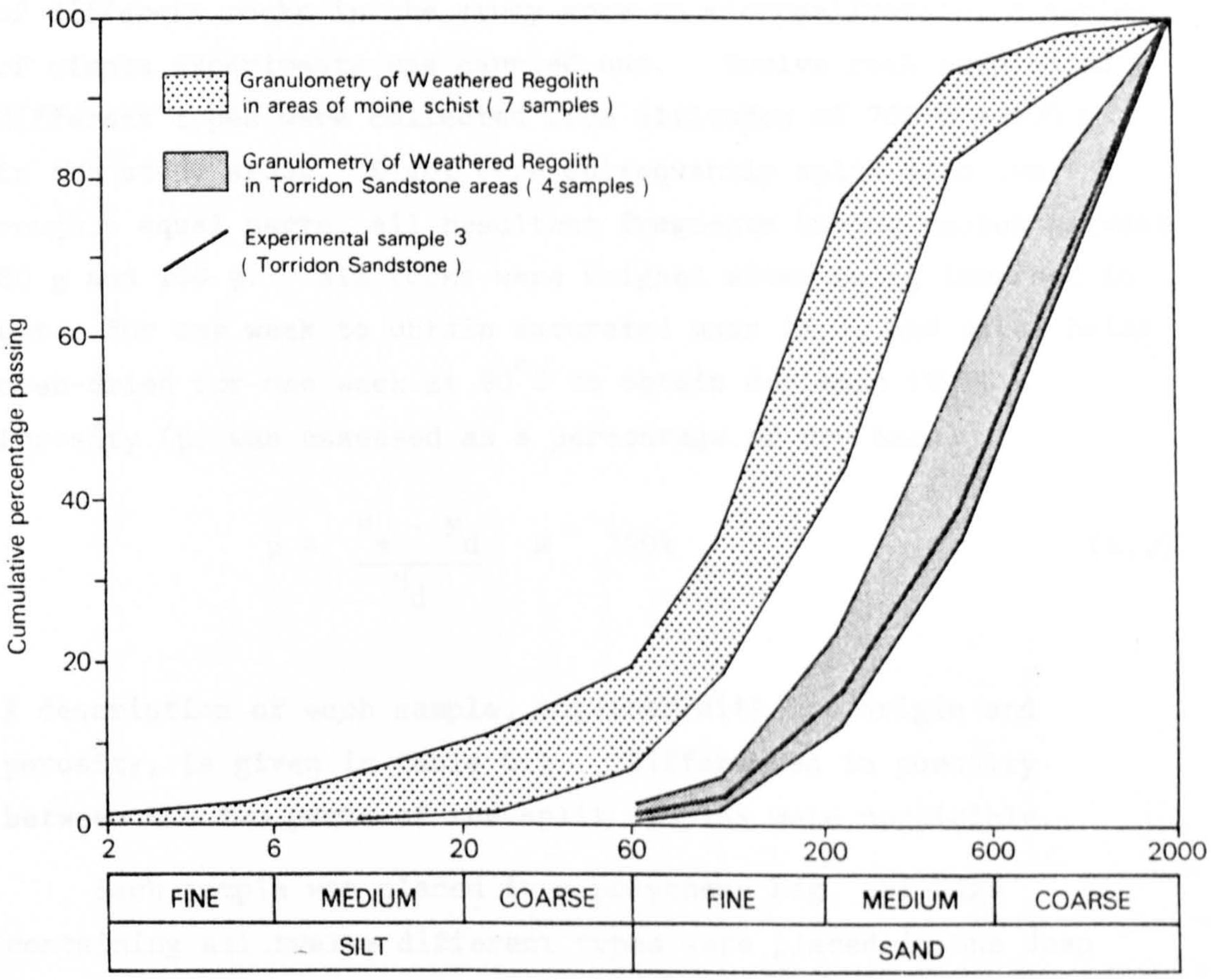


Figure 6.3: The granulometric characteristics of frost-weathered regolith in areas of Torridon Sandstone and Moine schist.

formed as a result of frost weathering (cf Ragg and Bibby, 1966). The formation of a well-developed podzol in these deposits on the Ben Wyvis plateau rules out a recent origin, and the incorporation of angular clasts produced by macrogelivation indicates that these soils probably represent the products of microgelivation during the Lateglacial cold periods (see below and chapter 8).

6.4.3 Laboratory experiments

To test the above conclusions concerning the susceptibility of different rocks in the study area to microgelivation, a series of simple experiments was carried out. Twelve rock samples of different types were collected from altitudes of 700 m - 900 m in the study areas. Eight were subsequently split into two roughly equal parts, all resultant fragments having masses between 50 g and 200 g. All rocks were weighed after being immersed in water for one week to obtain saturated mass (M_s), and after being oven-dried for one week at 90°C to obtain dry mass (M_d). Porosity (p) was assessed as a percentage of dry mass:

$$p = \frac{M_s - M_d}{M_d} \times 100\% \qquad (6.2)$$

A description of each sample, together with its origin and porosity, is given in table 6.2. Differences in porosity between the two parts of the split samples were negligible.

Each sample was placed in a polythene bag, and bags containing all twelve different types were placed in one deep tray, whilst bags containing the eight duplicate samples were placed in another. The bags and the spaces between the bags were filled with distilled water until all rocks were approximately three-quarters immersed. This is not unrealistic for freeze-thaw experiments, as during winter exposed rocks in upland areas of Scotland are often completely encased by ice. The tray of twelve samples was subjected to 250 freeze-thaw cycles of 24 h duration, with a mean freezing period of 10 h at -18°C, and a mean thaw period of 14 h at +14°C. The tray of eight samples was subjected to 45 freeze-thaw cycles of 168 h duration, with a mean freezing period

Table 6.2: Results of freeze-thaw experiments

Sample	Origin	Description	Porosity %	Percentage disintegration				
				24 h cycles			168 h cycles	
				50	100	250	20	45
1	A.T.	Coarse-grained Torridon Sandstone	1.25	1.46	2.84	4.48	0.15	0.86
2	A.T.	Medium-grained Torridon Sandstone	1.89	1.58	3.16	9.05	0.23	0.98
3	A.T.	Weathered Torridon Sandstone	3.32	8.45	32.07	100.00	-	-
4	A.T.	Fine-grained Cambrian Quartzite	0.31	0.04	0.06	0.15	-	-
5	A.T.	Medium-grained Cambrian Quartzite	2.27	0.10	0.19	0.75	0.44	0.59
6	A.T.	Pebbly basal Cambrian Quartzite	1.68	1.93	2.29	2.62	0.03	-
7	F.	Micaceous schist	0.87	0.27	0.46	0.80	0.05	0.12
8	F.	Siliceous schist	1.24	0.13	0.22	0.26	0.17	0.34
9	B.W.	Granulite	0.49	0.02	0.03	0.08	0.02	-
10	B.W.	Coarse-grained siliceous schist	1.63	0.08	0.12	0.30	-	-
11	B.W.	Coarse-grained siliceous schist	2.91	0.07	0.13	0.26	0.12	-
12	B.W.	Fine-grained siliceous schist	0.48	0.02	0.03	0.11	-	-

A.T. - An Teallach F. - Fannich Mountains B.W. - Ben Wyvis

of 144 h at -18°C, and a mean thaw period of 24 h at +14°C. The bags were periodically removed, and weathered products washed out, dried and weighed.

The results of these experiments are summarised in table 6.2 and shown diagrammatically on figures 6.4 and 6.5. They provide strong support for the conclusions reached in the field. Samples of rock in category 1 (above), including Torridon Sandstone (samples 1, 2 and 3) and basal Cambrian Quartzite (sample 6) disintegrated far more rapidly than any others. This indicates that roundness of exposed rock surfaces and abundance of granular detritus are successful criteria for identifying rapid microgelivation. Sample 3, a weakly-consolidated sample of Torridon Sandstone with a whitish rind suggestive of chemical weathering, disintegrated completely in 192 cycles, indicating the importance of other forms of weathering in weakening rocks (Martini, 1967; Lautridou, 1971b). The rate of disintegration of the Torridon Sandstone samples is rapid in comparison with other sandstones. Potts (1970), for example, measured less than 1% disintegration over 200 cycles for fresh sandstones from Wales. Interestingly, Torridon Sandstones subjected to repeated freezing to -5°C by Mainguet-Michel et al. (1971) yielded no detritus whatever. Their samples, however, consisted of fresh rock from low altitudes, which suggests that the samples used in the present analysis had been significantly weakened by previous microgelivation and/or chemical weathering before the experiments were carried out. Alternatively, freezing to -5°C may be insufficient to cause microgelivation on Torridon Sandstone, although this seems unlikely in light of the conclusions reached by Lautridou (1971b) concerning the equal effectiveness of cycles of different severity.

Rocks of category 2 (above), which are angular or subangular in the field, proved the most resistant to microgelivation in the laboratory. Samples 4 (fine-grained quartzite), 9 (granulite) and 12 (fine-grained siliceous schist) all yielded less than 0.15% weathered products. These results are in accord with those of previous experimenters (Tricart, 1956; Wiman, 1963; Martini, 1967; Lautridou, 1971a). Coarser siliceous schists (samples 8, 10 and 11) yielded similar percentages of detritus (0.26-0.30)

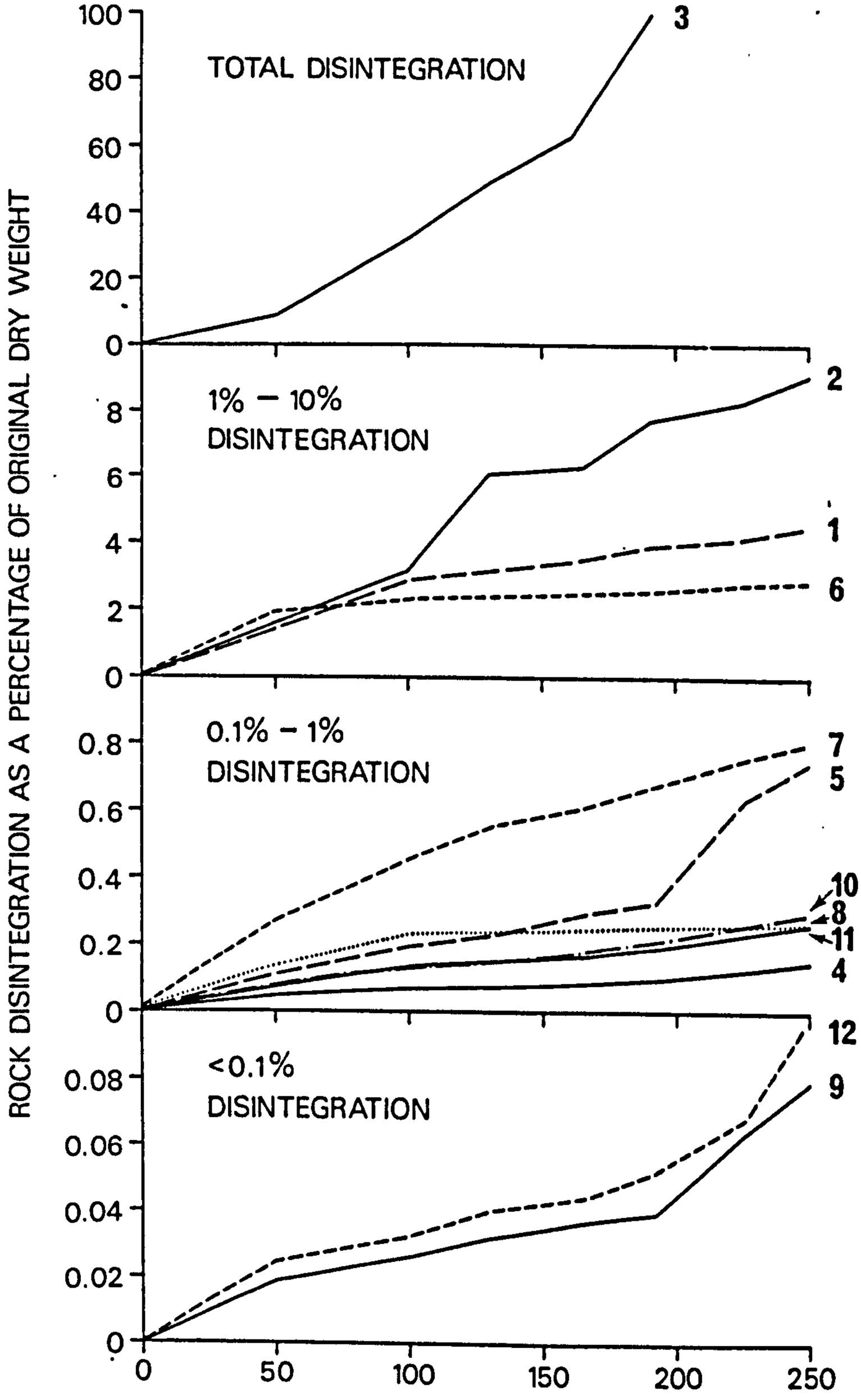


Figure 6.4: Percentage (by weight) disintegration of rock samples over 250 freeze-thaw cycles of 24 h duration. The sample numbers are identified in table 6.2.

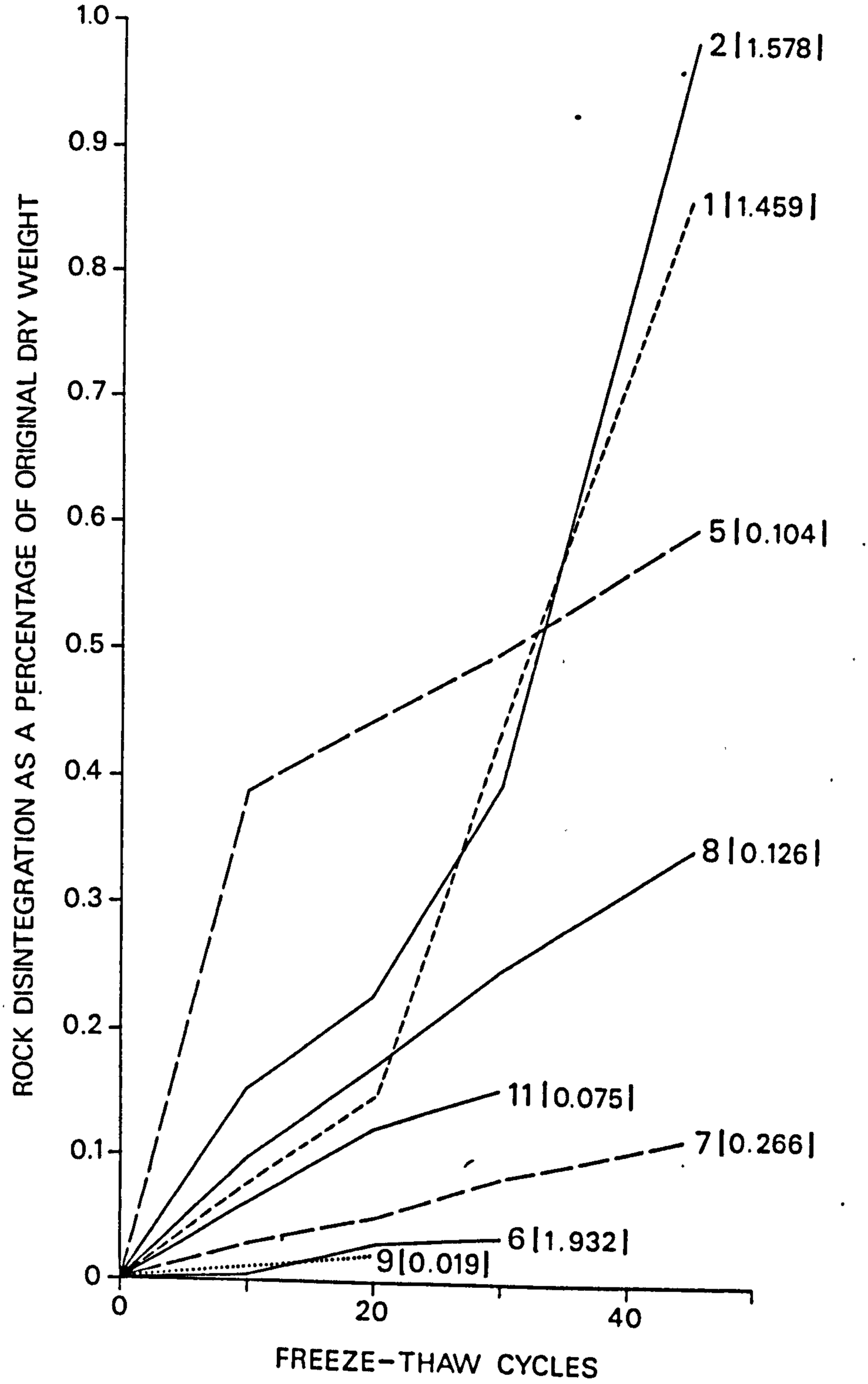


Figure 6.5: Percentage (by weight) disintegration of rock samples over 45 freeze-thaw cycles of 168 h duration. The sample numbers are identified in table 6.2.

under the 24 h cycles despite considerable variations in porosity, but a very micaceous schist (sample 7) yielded 0.80% despite low porosity, suggesting that susceptibility to microgelivation increases with mica content. Medium-grained Cambrian Quartzite also proved moderately susceptible; amongst the Cambrian Quartzites (4, 5 and 6) susceptibility apparently increases with grain size.

Comparison of the results for the 24 h and 168 h cycles reveals no consistent trend. The longer cycle is apparently less effective in disintegrating Torridon Sandstone, basal Cambrian Quartzite and mica-schist, but more effective in breaking down medium-grained Cambrian Quartzite and siliceous schists. This may indicate that a longer period of freezing is more effective in the disaggregation of strongly-bonded rocks.

For all twelve samples there is a moderately strong positive exponential relationship between disintegration after 192 24 h cycles (Dn) and porosity (figure 6.6), represented by the regression:

$$Dn = 0.0605\ e^{1.525\ p} \qquad (r^2 = 0.401;\ p < 0.015). \quad (6.3)$$

The most likely interpretation of the residuals is in terms of the strength of the bonds between individual crystals and grains (Lautridou, 1971b; Mainguet-Michel et al., 1971). Samples with positive residuals (Torridon Sandstone, basal Cambrian Quartzite and mica-schist) appear to be relatively weakly bonded, whereas those with negative residuals (granulite, siliceous schist and medium-grained quartzite) appear to be strongly bonded.

Although only sample 3 yielded sufficient weathered material to allow particle-size analysis (figure 6.3), inspection of the weathered products of the other samples showed that size of weathered products was dependent on grain size of the parent rock. The Torridon Sandstone and basal Cambrian Quartzite samples yielded mainly coarse sand or small granules; the siliceous schists, granulite and medium-grained quartzite gave mainly medium sand; and the fine-grained quartzite and schist produced fine sand and silt. Weathering of the mica-schist produced biotite flakes of various sizes finer than coarse sand.

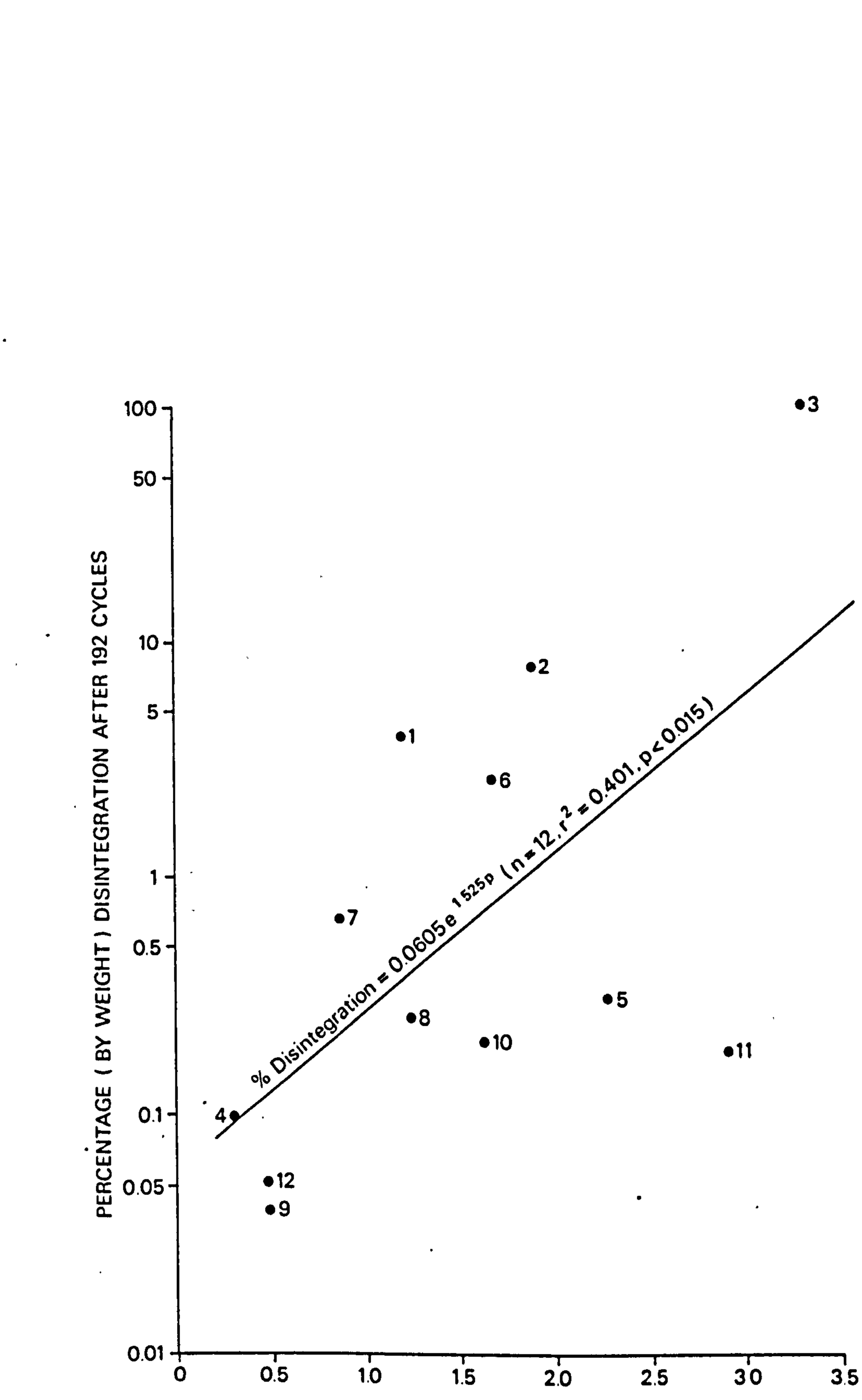


Figure 6.6: Relationship between percentage (by weight) disintegration of 12 rock samples after 192 freeze-thaw cycles and porosity.

6.5 Macrogelivation

6.5.1 Macrogelivation on Torridon Sandstone

In Torridon Sandstone areas, macrogelivation has operated through the riving apart of bedrock along pseudo-bedding (dilation jointing) planes to produce slabs that occasionally exceed 1 m in length (Godard, 1965), slab length and breadth being largely dependent on the density of joint systems that intersect pseudo-bedding at angles around 90°. Displacement and tilting of joint-bound blocks suggest that riving has also opened up joints. Not all pseudobedding planes have been affected by macrogelivation. Those that have been tend to increase in frequency and the depth to which they have been "etched out" towards the top of outcrops. On top of many outcrops are completely severed slabs, some in situ, others having tilted or toppled over the edges of the outcrops. The principal plane of some slabs is concave, a feature considered by Tivy (1962) and King (1968) to be symptomatic of frost weathering.

Several points of evidence suggest that the intensive frost weathering that produced opening of joints and pseudobedding planes no longer operates:

(i) Under winter conditions, ice often encases rock outcrops, but rarely fills open joints or pseudobedding planes.

(ii) There is a lack of fresh, sharp-edged facets. Virtually all exposed surfaces have been rounded by ~~microgelivation~~ granular disaggregation.

(iii) The exposed surfaces of outcrops and slabs are often lichen-covered.

(iv) There is little evidence for the "etching out" of pseudo-bedding in the source areas of recent rockfalls.

Sissons (1967a) inferred from dramatic contrasts in the degree of frost weathering of Torridon Sandstone inside and outside the limits of the Loch Lomond Advance that macrogelivation

had not taken place since the end of the Lateglacial period. This conclusion was tested on An Teallach by sampling the depth to which pseudobedding planes have been etched out or prised open in comparable sites (near-vertical outcrops) inside and outside the Loch Lomond Advance limit. The results (figure 6.7) show that depth of etching for sites inside the limit at 600 m and 650 m does not exceed 15 cm, whereas for sites outside the limit, at 600 m and higher, maximum depth of etching exceeds 50 cm. This suggests that little etching has taken place since the disappearance of the Loch Lomond Advance glaciers.

Although the prising apart of Torridon Sandstone along pseudobedding planes may be ascribed to frost wedging, the initial etching of pseudobedding planes (or weaker beds) must have preceded wedging, forming an opening in which ice could accumulate. From figure 6.7 it is apparent that maximum depth of etching for sites outside the Loch Lomond Advance limit increases with altitude. Field observations suggest that, above 500 m, exposure is the major controlling factor, as exposed ridge-crest and plateau outcrops generally display greater etching than relatively sheltered valley-side outcrops at similar altitudes. Significantly, etching of pseudobedding planes is absent on buried rocks. These observations suggest that initial etching of pseudobedding results from differential microgelivation, weaker beds being weathered through granular disintegration to provide openings suitable for the operation of wedging. This explanation accounts for the fact that not all pseudobedding planes have been affected by macrogelivation.

Whereas displacement of joint-bound blocks is telling evidence for the operation of frost-wedging in joints, it is probable that this process was aided by joint enlargement through pressure release concomitant with the downwastage of the last ice sheet. Open joints are found at all altitudes, but at shallow depth the joint system is largely latent. It is suggested that wedging exploited the open joints produced by pressure release. In both a horizontal and a vertical sense, therefore, the efficacy of wedging was apparently dependent on the opening of suitable cavities by other mechanisms.

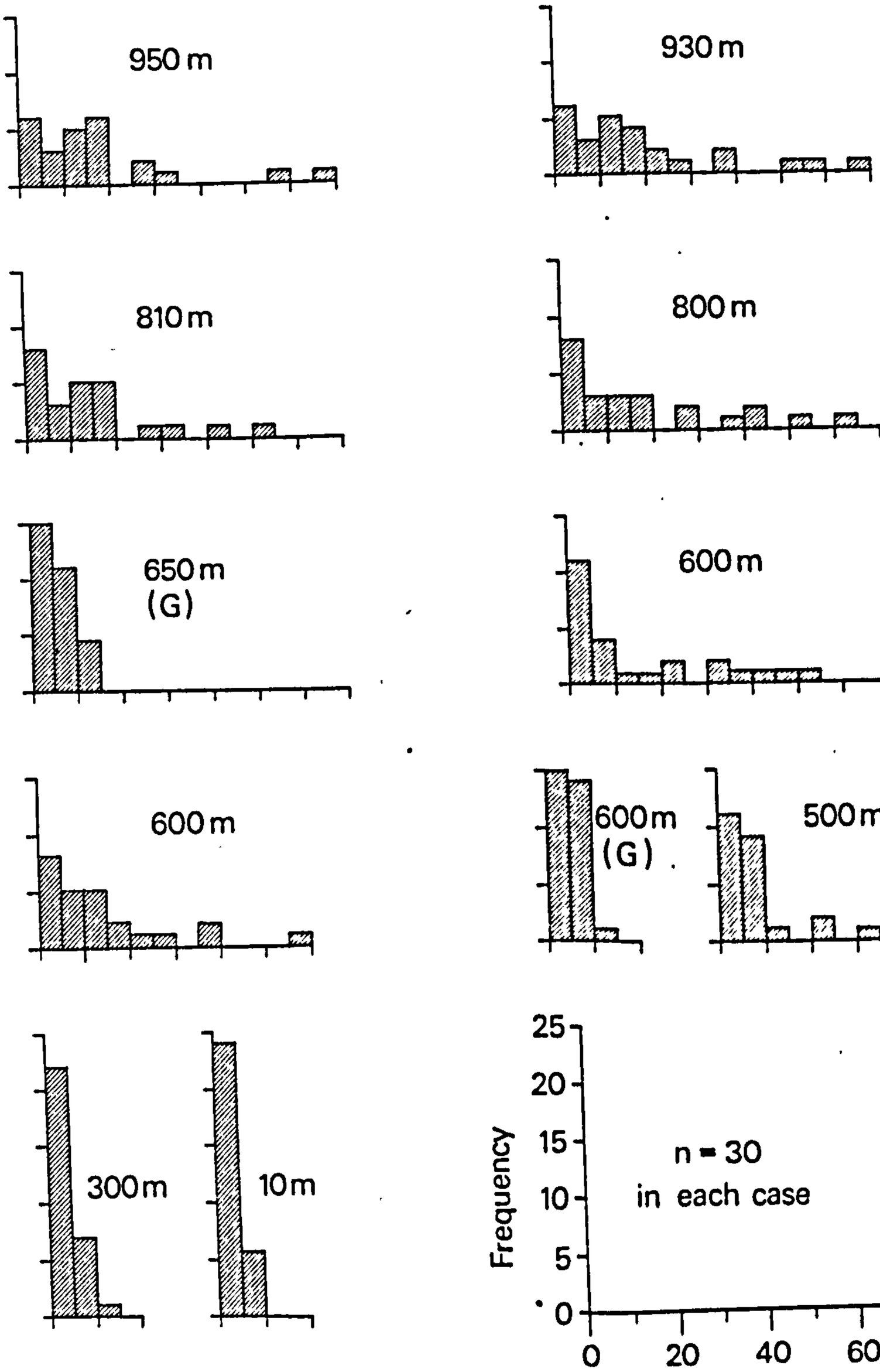


Figure 6.7: depth of "etching" or open jointing in pseudobedding in vertical outcrops of Torridon Sandstone at different altitudes on An Teallach. Outcrops inside the limit of the Loch Lomond Advance glaciers are identified by (G).

Although frost wedging is apparently no longer active, freshly "cracked" slabs are found on top of exposed outcrops above 800 m. These slabs have been split across their principal planes where no apparent line of weakness exists. This type of fracture corresponds to that obtained experimentally by Mellor (1970, 1973) under conditions of > 50% saturation, and suggests that limited macrogelivation capable of producing relatively small clasts still occurs at altitudes over 800 m in Torridon Sandstone areas. Similar conclusions were reached by King (1968) for schists above 900 m in the Cairngorms.

6.5.2 Macrogelivation on other rocks

Evidence for macrogelivation is found on all rock types in the study areas above an altitude of about 700 m. Its effects on different rocks vary in two respects: in the size and shape of weathered detritus, and in the depth to which the process has operated.

The size of weathered fragments reflects the density of planes of weakness in the parent rock. The siliceous schists of Mheall a'Chrasgaidh, for example, are densely jointed, so that few weathered clasts exceed 30 cm in length. Shape is determined by the relative density of vertical or near-vertical planes of weakness (principally joints) and horizontal or near-horizontal structures, such as bedding planes, joints, pseudobedding planes and cleavage planes. Where these densities are similar blocky clasts are produced, such as the joint-bound blocks of fine-grained Cambrian Quartzite on Sàil Liath, An Teallach, and of granulite on Càrn Gorm, Ben Wyvis. Where they differ, slabby or elongate clasts result. This occurs on the coarse-grained Cambrian Quartzite of An Teallach (figure 6.1), where the density of bedding planes exceeds that of the joints that intersect them, and is particularly noticeable on some mica-schists in which closely-spaced cleavage planes have been exploited by macrogelivation to give parallel-sided slices ranging in thickness from a few millimetres to 25 cm.

On Torridon Sandstone, macrogelivation is essentially a

surface phenomenon. This is also true of basal quartzite and schist areas. Deep macrogelivation is restricted to areas of fine-grained quartzite and granulite, on which exposures of fresh rock are rare (cf Crampton and Carruthers, 1914). A pit dug in the granulite blockfield on Càrn Gorm, Ben Wyvis, showed that this weathered deposit is at least 1.6 m deep. Differences in depth of weathering apparently reflect differences in the width of joints at depth. Those in fine-grained quartzite and granulite are open to greater depths than those on other rocks, so that frost wedging operated to greater depths.

As on Torridon Sandstone, there is abundant evidence to indicate that macrogelivation on other rocks is no longer active. All exposed rocks display a certain amount of surface rounding that is absent on buried rock surfaces, and large areas of shattered bedrock are covered by vegetation and by the products of microgelivation. Gneiss exposed at the surface by the construction of a stalkers' track in the Fannichs shows no sign of breaking up after over 70 years of exposure.

6.5.3 Macrogelivation and the delimitation of former glaciers

Various authors (e.g. Sissons, 1977b; Ballantyne and Wain-Hobson, 1980) have employed the limit of pronounced macrogelivation to identify the former limits of Loch Lomond Advance glaciers. Measurement of depth of etching along lines of weakness, as described above (figure 6.7) offers one method of testing visual impressions as to the limit of pronounced macrogelivation.

An alternative method, applicable on rocks that yield slabby detritus, involves consideration of the shape of cobbles and boulders in the area of the proposed glacial limit. On An Teallach, glacially-transported clasts are generally blocky (i.e. have high b:a and c:a axis ratios) whereas the products of macrogelivation are predominantly slabby (b:a high, c:a low). These differences are immediately apparent when samples of the two types of deposit are plotted on ternary shape diagrams (figure 6.8) and compared. Samples of <u>in situ</u> weathered slabs (P1 - P3) and weathered slabs lying on the plateau surface

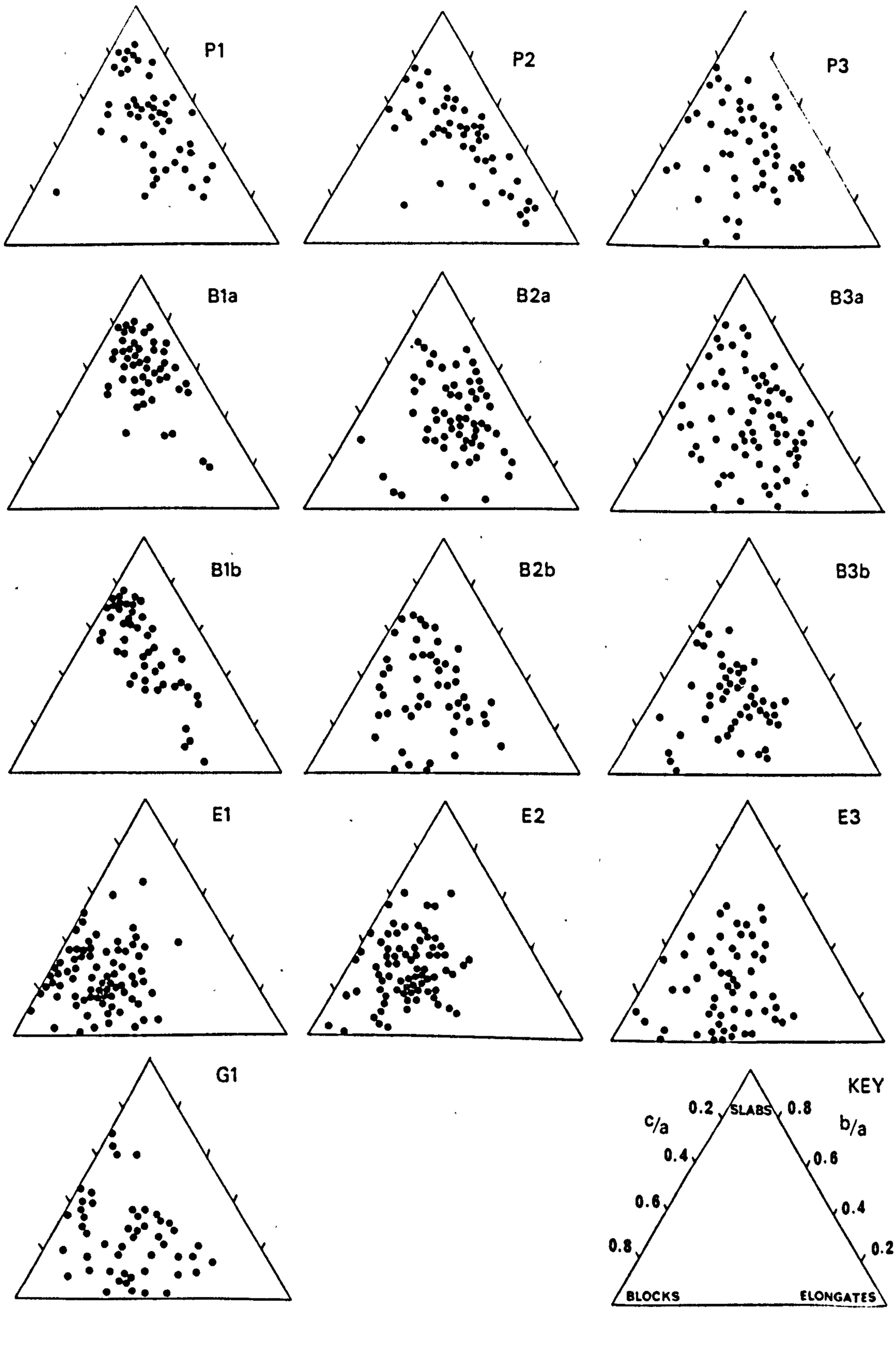


Figure 6.8: Ternary diagrams depicting characteristic clast-shape distributions for various deposits. P1, P2, P3: in situ weathered slabs. B1a, B2a, B3a, B1b: weathered slabs on the plateau surface or immediate subsurface. B2b, B3b: weathered bedrock under plateau-surface detritus. E1, E2, E3, G1: clasts from a Loch Lomond Advance end moraine (E) or fluted moraine (F)

(B1a, B2a, B3a) or immediate subsurface (B1b) have a wide range of b:a axis ratios, but c:a axis values tend to be less than 0.4. For samples taken from the Loch Lomond Advance end moraine in Coire a'Mhuilinn (E1 - 3) and from a fluted moraine in Glas Tholl (G1), c:a axis ratios of less than 0.4 are rare, giving concentrations of points in the left-hand corners of the diagrams. In this case the differences in the shape characteristics of the two types of debris are sufficiently great to obviate the need for statistical testing. In areas of Torridon Sandstone that have suffered Lateglacial macrogelivation, therefore, it would seem to be possible to identify former glacial limits through establishment of the characteristic shape of the detritus.

This property is not limited to Torridonian rocks. In Alaska, Seppala (1976) found that glacially-transported clasts have a significantly lower flatness index than those on a nearby talus slope, and the writer (Ballantyne, in preparation) found similar differences in a study of over 2000 glacially-transported and frost-shattered gneissic clasts in the Jotunheimen massif, Norway. Many rock types, including some in the study areas, yield slabby detritus on weathering, so the analysis of particle shape may find applicability in determining former glacial limits in many areas where macrogelivation accompanied glaciation.

The simplest explanation for the high "blockiness" of glacial debris is that rocks initially produced by macrogelivation tend to be cracked across their principal axes in the course of transportation within glacier ice. Alternatively, high "blockiness" may result from glacial entrainment of large numbers of buried clasts. It was noted above that etching of pseudobedding is absent on buried rock, and this is reflected in the blocky nature of buried clasts (e.g. samples B2b and B3b, both taken from the base of pits on the plateau surface). On An Teallach, a rockfall origin for blocky glacial detritus can be discounted as there are no free faces in Coire a'Mhuilinn upvalley of the Loch Lomond Advance moraine.

6.6 Chemical weathering

Chemical weathering appears to have been of relatively little importance in effecting rock breakdown in the study areas. The lack of clay in weathered deposits (figure 6.3) indicates that these are the products of mechanical rather than chemical disaggregation (cf Ragg and Bibby, 1966). Chemical weathering may, however, have weakened the binding cement or minerals in some Torridonian rocks (e.g. those with whitish weathering rinds) thereby rendering these rocks more susceptible to microgelivation.

6.7 Conclusions

Microgelivation is active in the study areas, but its present effects are important only on weakly-bonded, relatively porous rocks such as Torridon Sandstone and basal Cambrian Quartzite. Moine schists suffer relatively little microgelivation at present, but there is evidence that during the Lateglacial period they underwent intense granular breakdown. Strongly-bonded rocks with low porosity, such as fine-grained quartzite and granulite, have yielded little to microgelivation. In general, the granulometry of the products of microgelivation reflects that of the parent rock.

Macrogelivation has operated along planes of weakness on all rock types, and the density of such structures has determined the size and shape characteristics of weathered clasts. The operation of frost-wedging was apparently dependent on the pre-existence of fissures produced by other mechanisms, hence it operated only at the surface except where open joints occurred at depth. Macrogelivation in the form of frost-wedging apparently ceased at the end of the Loch Lomond Stadial, although there is evidence for limited recent frost-cracking of Torridon Sandstone. In areas in which macro-gelivation produced slabby detritus, analysis of particle shape can be employed to determine the former limits of Loch Lomond Advance glaciers. The lack of clay in weathered regolith suggests that chemical weathering has been of relatively little importance in effecting rock breakdown in the study areas.

As noted in the introduction to this chapter, the varying response of different lithologies to periglacial weathering has produced a detritus mantle of widely differing physical characteristics, ranging from a smooth, protective soil "blanket" in some areas to chaotic openwork blockfields in others. The nature of this detritus mantle is discussed in greater detail below (chapter 8).

CHAPTER 7 CLASSIFICATION AND MAPPING

7.1 Introduction

In chapter 3 it was argued that comprehensive classification of the periglacial phenomena of upland Britain is overdue, not least because the establishment of a suitable scheme would facilitate the production of detailed large-scale maps showing the distribution of such features. A classification scheme based on the author's observations is presented in the present chapter, together with definitions of all features represented and the system of mapping symbols that was used to depict these features on the large-scale maps contained in the back pocket of this thesis.

7.1.1 The need for classification

Classification of periglacial features in upland Britain is not new; it is implicit in every study that has differentiated between recognisably distinct types from Hollingworth (1934) on. The construction of a formal, explicit scheme, however, possesses important advantages over such ad hoc categorization:

(i) The construction of a comprehensive scheme creates a conceptual framework within which the researcher can order ideas, data and methodological strategy.

(ii) The researcher is forced to recognize and make explicit the criteria on which he bases differentiation of different types and classes of phenomena.

(iii) Periglacial phenomena, like other natural phenomena (e.g. plants, animals and rocks) can be ordered in a hierarchical scheme: "oblique turf-banked terraces", for example, represent a sub-group within the group "turf-banked terraces"; this group in turn forms a subset of "all terrace features", itself a division of the major group encompassing "all mass-movement features". Formal classification makes clear the nature of such hierarchies.

(iv) The taxonomic organization involved in classification

requires precise definition of terms, which ensures (or should ensure) consistency of nomenclature. Standardization of terminology is highly desirable if the semantic confusion that presently exists in this field is to be ended (chapter 3).

(v) Formal classification is prerequisite to the production of meaningful maps depicting periglacial phenomena.

(vi) If a classificatory and taxonomic scheme receives widespread acceptance, comparison of the results obtained by different researchers is facilitated.

7.1.2 Previous classification schemes

Periglacial geomorphology has a long-established tradition of classification, but no universally-accepted comprehensive scheme has yet emerged and, despite attempts at standardization (e.g. Bryan, 1946), periglacial nomenclature remains "irrational, imprecise, incomplete and non-systematic" (Hamelin and Cook, 1967, p. 11). Even classification of single groups of features has proved problematic. Patterned ground, for example, was first divided into three or four types on an entirely empirical basis (Meinardus, 1912; Bergstrom, 1912; Högbom, 1914), but later classifications attempted to relate different types of patterned ground to formative influences such as slope and boulder concentration (Beskow, 1930), climate and age (Steche, 1933) and regolith, slope and moisture supply (Sorenson, 1935). The over-simplifying generalizations involved in such schemes, however, led many workers to adopt Washburn's (1950, 1956) relatively simple classification, which was based on morphology (polygons, circles, nets, steps and stripes) and presence or absence of lateral sorting. Even this is unsatisfactory, as most of the ten resulting combinations may be used to represent two or more different features: the term "nonsorted stripes", for example, may imply a "ridge and furrow" pattern or alternating stripes of bare ground and vegetation. Such considerations have recently led Nicholson (1976) to propose a more comprehensive scheme based on form, grouping, pattern marking and size. It is perhaps significant that Lundqvist (1962) in his excellent review of

patterned ground features in Sweden abandoned any rigid form of classification and resorted to an unstructured empirical scheme that reflected Washburn's system only in terminology.

All of the classifications proposed for periglacial features in upland Britain (King, 1968; Ball and Goodier, 1970, 1974; Goodier and Ball, 1975; Mathieson, 1977) have been local in scope. Of these, that proposed by Ball and Goodier (1970) for frost action features in Snowdonia is the most comprehensive and the most ingenious. It involves a three-part division of periglacial phenomena into major physiographic features (e.g. mountain-top detritus and scree slopes), minor physiographic features (subdivided into lobate forms, regular patterned forms and irregular forms) and modifications of drift fabric (e.g. frost shattering, involutions and downslope-oriented fabrics). For each type of feature Ball and Goodier attempted to specify the dominant formative process or processes (gelifraction, gelifluction and/or cryoturbation) and period(s) of formation (Lateglacial, recent and/or present). As with Lundqvist's scheme, the categories were selected on an empirical, largely morphological basis. A similar approach is adopted here.

7.2 Principles of classification

"Classification is the grouping of objects into classes on the basis of properties or relationships they have in common" (Grigg, 1965, p. 466). There are two main approaches: the divisive approach ("logical division") and the agglomerative approach (Harvey, 1969; Johnston, 1976). The former assumes that all individual types can be derived by repeated subdivision of the universal set of phenomena under consideration using a series of predetermined criteria, and is inappropriate in the present context. The classification described here is agglomerative, in that it was constructed through grouping a large number of observed "types" on the basis of similar characteristics. Of necessity, the groupings used here are highly subjective because (i) the range of types identified reflects the author's field experience and interpretation of the published

literature, (ii) the criteria employed for grouping features are non-systematic, and (iii) most features identified here as individual types could be subdivided into sub-groups according to further criteria.

The final point is probably the most important. Ploughing boulders, for example, are here regarded as one "type" of feature, although Tufnell (1972) described several different types of ploughing boulder. On the other hand, more "types" of detritus lobe are differentiated here than in the classification by Ball and Goodier (1970). Recognition of the "lowest common denominator" is thus entirely subjective, particularly as many of the features under consideration are not discrete types but part of a continuum of related forms (Matthews and Petch, 1970); for such features the process of classification involves recognising such continua (e.g. lobes, sheets, terraces) and subdividing these according to definable criteria.

The classification presented here is essentially a priori in that it was drawn up in its original form before the results of the study were known in order to provide a conceptual framework within which research could be carried out. Some of the criteria that might have been used to group features (e.g. age, genesis) could not therefore be employed as these were largely unknown at the start of the study. The guiding principles that were employed were as follows.

(i) That the classification should encompass all recognizably different types of phenomena, including those that are not strictly "periglacial" (e.g. slump features, debris chutes, gulleys) and upland terrain types (e.g. vegetation-covered plateau surfaces).

(ii) That where possible the terminology employed should reflect established nomenclature yet be descriptive rather than genetic (e.g. vegetation-covered lobes rather than solifluction lobes). These two aims could not always be reconciled, however, and some of the terms used (e.g. nivation hollow, aeolian sand deposits) presuppose mode of formation.

(iii) That grouping of features should be made on the basis of

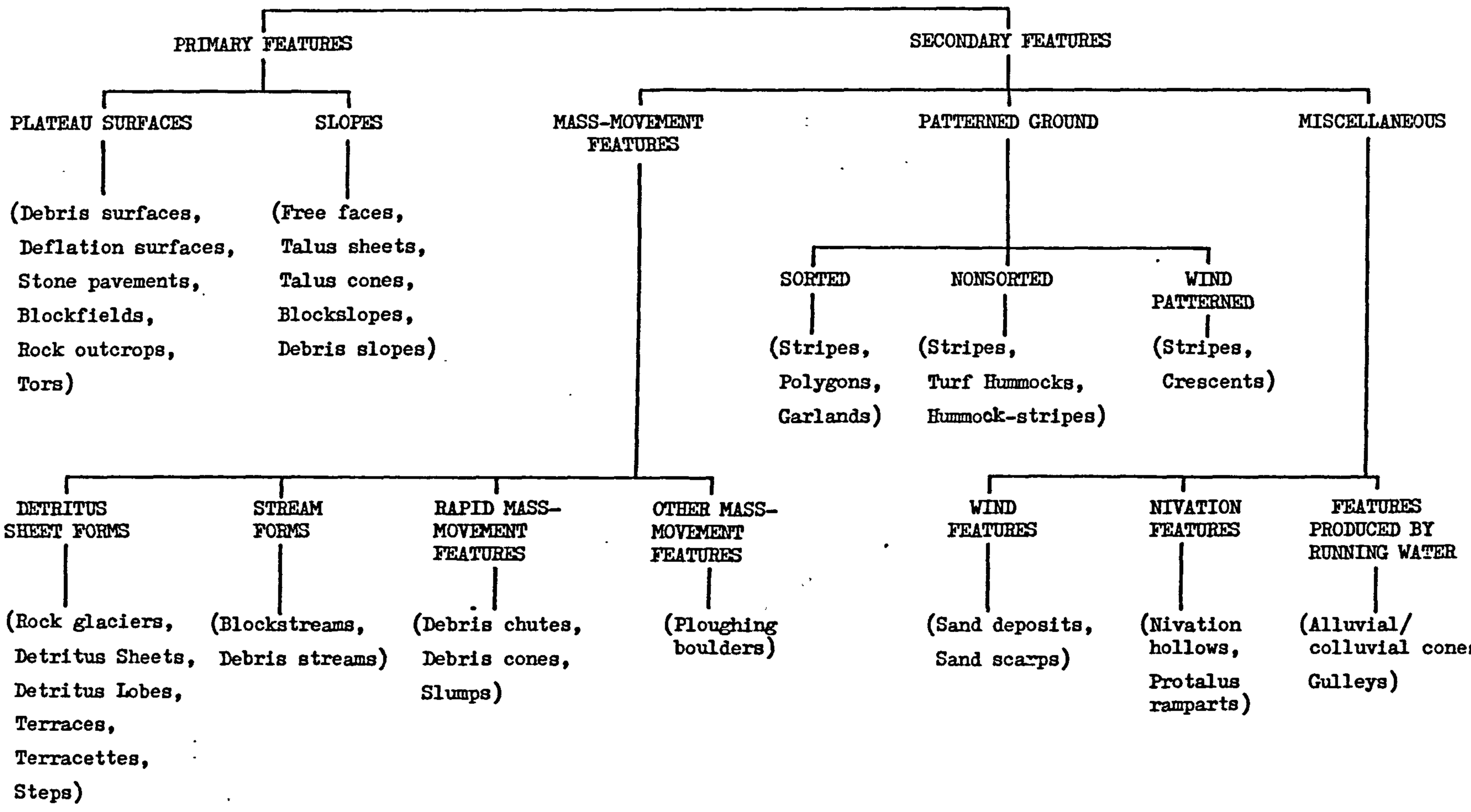
definable non-genetic characteristics such as morphology, particle-size composition and vegetation cover (again not always possible).

(iv) That the classification should be sufficiently detailed to allow recognition of all distinct types, yet sufficiently concise to allow representation of all types on a 1:10,000 geomorphological map.

The structure of the final classification scheme is shown in table 7.1. The basic unit upon which this classification is built is the "morphological type". The features listed in brackets in table 7.1 constitute such "morphological types" (e.g. debris surfaces, tors, lobes, sorted stripes, ploughing boulders, nivation hollows), each of which is morphologically and/or structurally distinct from all other types. In some cases it proved convenient to subdivide these basic units according to further criteria, such as particle-size composition (e.g. boulder lobes and debris lobes), vegetation cover (e.g. unvegetated talus, vegetation-covered talus), size (e.g. large-scale patterned ground, small-scale patterned ground) and even morphology (e.g. horizontal terraces, oblique terraces, intersecting terraces, lobate terraces). Such subdivisions are described in the following section, and the complete classification scheme is outlined on the map key contained in the back pocket of the thesis (figure 7.1).

The grouping of features in table 7.1 is one of convenience and does not reflect a systematic classification strategy. The basic "morphological types" were placed in initial classes on the basis of a number of subjective criteria (principally morphology and/or assumed genetic relationships) and these initial classes were then combined to form five major families of features (plateau surface features, slopes, mass-movement features, patterned ground and miscellaneous). These five families were further grouped into "primary" and "secondary" features. The former group consists of rock outcrops and various types of detritus cover, the latter of features developed on the detritus cover. This distinction is broadly equivalent to that drawn by

Table 7.1 : Classification of periglacial landforms



Ball and Goodier (1970) between "major physiographic features" and "minor physiographic features".

The scheme outlined above may be criticized for its non-systematic approach and for the inclusion of certain features in some categories rather than others. Wind stripes, for example, could be accommodated equally well in the sections on patterned ground and wind action features. Classification, however, should be regarded " ... as a flexible device, one to be altered to meet our needs in any given situation" (Harvey, 1969, p. 327). The present classification represents a convenient research tool; the organization of periglacial phenomena on the basis of more systematic criteria (e.g. age, formative influences) is discussed in the final chapter.

7.3 Definitions

One of the advantages of explicit classification is that it necessitates precise definition of all terms used. A complete list of such definitions is given in the present section. For convenience, conciseness and clarity these definitions are stated as in a dictionary and each "morphological type" is listed within subsections devoted to each of the five "families" identified above. After each definition appears a list of sub-types (if any). Terms used to differentiate these sub-types are defined at the start of each sub-section.

7.3.1 Plateau surface features

(Definition of sub-types: "unvegetated" implies less than approximately 20% vegetation cover; "partly-vegetated" implies roughly 20%-80% cover).

Debris surface: regolith mantle of intermingled fine and coarse detritus on flat or gently-sloping terrain. Subdivisions: unvegetated, partly-vegetated, vegetation-covered.

Deflation surface: debris surface from which surficial fines have been largely removed by wind, leaving a gravel lag deposit. Subdivisions: unvegetated, partly-vegetated.

Stone pavement: debris surface with a concentration of cobbles and boulders (embedded in fine material) at the ground surface. Subdivisions: unvegetated, partly vegetated.

Blockfield: regolith mantle of coarse detritus, with no fine material in its upper part, on flat or gently-sloping terrain. Subdivisions: unvegetated, partly-vegetated, vegetation-covered.

Rock outcrop: bare bedrock protruding through the regolith on a flat or gently-sloping plateau surface.

Tor: "... a residual of bare bedrock, isolated by free faces on all sides, the result of differential weathering and mass slope-wasting" (Caine, 1967, p. 418).

7.3.2 Slopes

(Definitions of sub-types: as for section 7.3.1).

Free face: steep (> 40^{o}) rock face with little or no detritus cover.

Talus sheet: sheet-like accumulation of coarse rockfall detritus at the base of a free face. Subdivisions: unvegetated, partly-vegetated, vegetation-covered.

Talus cone: fan-like accumulation of coarse rockfall detritus at the base of a free face, usually below a gully. Subdivisions: unvegetated, partly-vegetated, vegetation-covered.

Debris slope: regolith mantle of both fine and coarse detritus on slopes greater than about 5^{o}. Subdivisions: unvegetated, partly-vegetated, vegetation-covered.

Blockslope: regolith mantle of coarse detritus with no fine material in its upper parts on slopes greater than about 5^{o}. Subdivisions: unvegetated, partly-vegetated, vegetation-covered.

7.3.3 Mass-movement features

(The classification of detritus sheet forms (table 7.1) is complex, as three major types (terraces, sheets and lobes) form members of a continuum. In order to encompass all possible morphological combinations, several additional sub-types are introduced below: lobe-fronted sheets, oblique terraces, interconnecting terraces and lobe-fronted terraces. The relationship between these forms is made clear in chapter 9. Further subdivision of these features is made in terms of vegetation cover or, where apparent, constituent material).

Debris features: forms composed of both fine and coarse detritus; partly vegetated.

Boulder features: forms composed of coarse detritus with little or no fine material evident; partly vegetated.

Turf-banked features: forms with a vegetation-covered risers and unvegetated treads or surfaces.

Stone-banked features: forms with a riser of unvegetated coarse detritus and vegetation-covered treads or surfaces.

Vegetation-covered features: forms characterised by a complete or nearly-complete cover of vegetation.

Fossil rock glacier: "... glacierlike tongue of angular rock waste" (Sharpe, 1938, p. 43).

Detritus sheet: uniform hillslope regolith mantle that follows the general angle of the slope but terminates downslope in a fairly straight (in plan) frontal riser. Subdivisions: boulder sheet, debris sheet, turf-banked sheet, stone-banked sheet, vegetation-covered sheet.

Lobe-fronted sheet: detritus sheet with a front that is crenulate in plan. Subdivisions: as for detritus sheet.

Detritus lobe: hillslope detritus form bounded by a riser that is U-shaped in plan. Subdivisions: boulder lobe, debris lobe, turf-banked lobe, stone-banked lobe, vegetation-covered lobe.

Horizontal terrace: hillslope detritus form with a steep, straight across-slope riser and a tread that is nearly horizontal both across-slope and downslope. Subdivisions: debris terrace, turf-banked terrace, vegetation-covered terrace.

Oblique terrace: As above, but with riser and tread aligned oblique to the slope. Subdivisions: oblique turf-banked terrace, oblique vegetation-covered terrace.

Interconnecting terrace: hillslope detritus form produced by the intersection of horizontal and oblique terraces and characterised by truncated horizontal risers and interconnecting oblique and horizontal treads. Subdivisions: interconnecting turf-banked terrace, interconnecting vegetation-covered terrace.

Lobe-fronted terrace: horizontal terrace whose riser is lobate or crenulate in plan. Subdivisions: lobate turf-banked terrace, lobate debris terrace, lobate vegetation-covered terrace.

Step: small-scale lobate turf-banked terrace, with riser not exceeding a few centimetres in height.

Terracette: turf-covered horizontal terrace, generally not exceeding a few decimetres in width.

Blockstream: band of unvegetated or partly-vegetated coarse detritus, generally aligned directly downslope and sometimes terminating in a turf-banked lobe.

Debris stream: band of unvegetated or partly-vegetated detritus comprising both coarse and fine material, generally aligned directly downslope and sometimes terminating in a turf-banked lobe.

Debris chute: track of a debris flow (mudflow), generally consisting of a gulleyed upper part that gives

way downslope to two parallel levées of coarse detritus. These enclose a band of relatively undisturbed ground and usually terminate in a detritus lobe.

Debris cone: fan of detritus produced at the base of a steep slope by repeated debris flow activity.

Slump feature: concave scar and associated tongue of detritus produced by localized slope failure.

Ploughing boulder: boulder located at the downslope end of a furrow-like depression.

7.3.4 Patterned ground features

(Definitions of subdivisions: "large-scale" implies a width or diameter greater than 1 m; "small-scale" implies a width or diameter less than 1 m.)

Sorted stripes: alternating bands of coarse and fine detritus that follow the line of maximum slope.
Subdivisions: large-scale sorted stripes, small-scale sorted stripes.

Sorted polygons: circles or polygons formed of coarse detritus that encloses relatively fine material.
Subdivisions: large-scale sorted polygons; small-scale sorted polygons.

Sorted garlands: elongate patterned ground with relatively fine material enclosed on at least three sides by coarse detritus. Subdivisions: large-scale sorted garlands; small-scale sorted garlands.

Nonsorted stripes: "ridge and furrow" features aligned downslope along the line of maximum gradient.

Turf hummocks: vegetation-covered domes of predominantly fine material separated by a network of depressions.

Hummock stripes: turf hummocks aligned in bands along the line of maximum slope.

Wind stripe: elongate deflation scar on otherwise vegetated terrain.

Wind crescent: semicircular vegetation pattern on deflation surface, with the "horns" of the semicircle facing the dominant wind.

7.3.5 Miscellaneous features

Aeolian sand deposit: accumulation of wind-transported fine detritus, principally of sand size.

Sand scarp: eroded, unvegetated margin of a sand deposit.

Nivation hollow or bench: site of a late-lying snowpatch with a relatively steep backslope and gently-sloping frontal "wash" slope.

Protalus rampart: ridge-like deposit, generally of coarse material, produced by detritus sliding or rolling down a perennial snowpatch and accumulating at its foot.

Alluvial/colluvial cone: fan of water-transported detritus deposited at the foot of a steep slope.

Gulley: stream course deeply incised in drift or bedrock on a steep hillslope.

7.4 Mapping

The construction of detailed large-scale maps of the periglacial features in the study areas posed three initial problems. First, it was impossible to begin mapping until a provisional classification of landforms had been established. Secondly, it was found that the scale of available aerial photographs

(c. 1:26,000 and c. 1:10,000 panchromatic imagery of variable tonal contrast and resolution) precluded identification of many periglacial features (Ballantyne, 1974) so that preliminary airphoto mapping was of limited value. Finally, the system of mapping symbols could not be based on existing schemes, many of which were designed for small-scale maps (e.g. Sekyra, 1961) or for use in arctic environments (e.g. Dutkiewicz, 1962; Hamelin, 1963; St-Onge, 1965). None of the existing maps of periglacial features in upland Britain (Galloway, 1958, 1961a; King, 1968; Ryder, 1968, 1975; Kelletat, 1970a; Sugden, 1970b; Whyte, 1970) was considered to offer a suitable starting point, as most of these maps are relatively small in scale (1:25,000 or smaller) and all depict a rather limited range of features, usually in a very generalized way.

These difficulties necessitated extensive preliminary reconnaissance of the study areas, following which a provisional classification of features and mapping key were drawn up. This provisional scheme was used during initial mapping of Ben Wyvis and An Teallach, in the course of which both the classification and the mapping symbols were extensively modified. A revised scheme was then drawn up based on both field experience and descriptions in the literature. This was designed to incorporate all of the periglacial features found in upland Britain, and formed the basis of the classification described above. A new set of symbols was designed according to the criteria listed below, and all three study areas were then mapped or re-mapped. This was done through repeated traverses through each mountain area, in the course of which the appropriate symbols were plotted on 1:10,560 Ordnance Survey maps. During mapping aerial photographs were used to ensure planimetric accuracy, partly through the plotting of major features (particularly detritus sheets) directly on to the airphotos. The applicability of the scheme to other areas was checked by mapping periglacial features in two mountain areas on Rhum.

Further amendments to the mapping symbols were made during the drawing of preliminary pencil drafts based on the field maps. When a satisfactory cartographic balance had been achieved the

final maps were drawn at scales of 1:10,560 (An Teallach, Fannichs, Ben Wyvis) or 1:10,000 (Rhum). These maps were then photographically reduced to 1:16,000 and appear at this scale in the back pocket as maps 2 (An Teallach), 3 (West Fannichs), 4 (East Fannichs), 5 (Ben Wyvis), 6 (Ben Wyvis North-East), 7 (Rhum Cuillin) and 8 (Western Hills of Rhum). Also contained in the back pocket is the complete mapping key (figure 7.1).

7.4.2 Symbols

The classification scheme outlined above incorporates nearly 80 different types or sub-types. It was decided to include on the maps information on altitude (contours and summit points), glaciation (glacial limits, deposits and landforms) and terrain (e.g. peat cover, vegetation cover, alluvium), bringing the total number of symbols required to 98. The design of these symbols was guided by the following criteria:

1. All recognised types or sub-types should have an unambiguous identifying symbol, although related features should have related symbols. Thus lobes are always identified by a downslope-oriented "U" shape, but different types of lobe are distinguished through forming the "U" in different ways or adding symbols within the "U".

2. Where appropriate, sub-types should be designed to allow gradations between these to be depicted. An increase in the vegetation cover on debris surfaces, for example, can be depicted by increasing the number of "vegetation" symbols, and the transition from detritus sheet to individual detritus lobes can be shown by increasing the sinuosity of the symbol and then breaking its continuity (figure 7.1).

3. The key should permit superimposition of certain symbols, particularly the superimposition of "secondary" types on "primary" surfaces (e.g. lobes or ploughing boulders on a debris slope).

4. Where possible the symbols employed should reflect the

morphology and orientation (if any) of the features depicted, yet permit planimetric accuracy to be maintained.

5. The symbols used should be suitable for plotting on 1:10,000 maps and aerial photographs in the field and for photographic reduction to slightly smaller scales.

7.5 Discussion

The classification and mapping scheme outlined above is in the author's experience unique (i) in that it attempts to incorporate all periglacial features found in upland Britain, (ii) in the range of subdivisions recognised and (iii) in the detail and accuracy attempted in mapping. Subsequent observations on more than thirty other mountain areas in Scotland and successful use of the scheme by G.P. Chattopadhyay (personal communication) in mapping periglacial features in the Cairngorm, Creag Meaghaidh and Drumochter Pass areas have indicated that it is sufficiently comprehensive to be employed in many parts of upland Britain.

The maps were designed primarily as research tools rather than heuristic models, means rather than ends. As the former they facilitate recognition of spatial relationships that aid the understanding of many features (chapters 8-12). As the latter, however, they provide a striking demonstration of the richness and variety of the periglacial landforms of upland Britain.

CHAPTER 8 SLOPES AND PLATEAUX

8.1 Introduction

The mountain slopes and plateaux of Highland Scotland have undergone a long and complex evolution. The oldest recognisable elements of the mountain landscape consist of a series of planation surfaces, thought to be of Tertiary age (Sissons, 1976a). The plateaux of An Teallach and Ben Wyvis, for example, were interpreted by Godard (1965) as isolated remnants of a "surface supériere", identifiable throughout the Northern Highlands, and the high plateau of the Cairngorms offers a more striking (and more convincing) example of a similar type of feature (Sugden, 1968). Even the best-preserved of these surfaces have, however, been extensively modified by successive Pleistocene glaciations, and in view of the freshness of the forms of glacial erosion in Scotland it is reasonable to assume that, in broad outline at least, present mountain slopes are very similar to those that emerged from under the last ice sheets and valley glaciers.

In detail, however, the glaciated upper slopes of Scottish mountains show evidence of subsequent modification, largely attributable to periglacial activity. This modification has taken the form of a "cut and fill" process involving the erosion of rock protrusions and glacially-oversteepened outcrops, the infilling of depressions and the accumulation of detritus on rock benches and valley floors. The end product of such activity has been the mantling of all but the steepest slopes by a detritus cover of variable thickness, the "mountain-top detritus" and scree that were vividly described in the memoirs of the Geological Survey over 60 years ago (e.g. Peach et al., 1912, 1913a, b; Crampton and Carruthers, 1914; Bailey and Maufe, 1916). In the present account, the term "detritus" is used to represent any accumulation of unconsolidated material, irrespective of origin and manner of accumulation, except for unmodified glacial deposits. "Detritus slopes" and "detritus surfaces" are convenient terms to describe all types of debris-mantled slopes and plateaux.

In the previous chapter, several types of detritus slope and surface were described and defined. These fall into three main categories:

(i) Talus slopes. These include talus sheets and cones, and are distinguished from other types in that they accumulate primarily through detritus falling from (or over) a rock wall and piling up against the foot of this wall.

(ii) Blockfields and blockslopes, which are clastic deposits characterised by an absence of fine material at the ground surface.

(iii) Debris slopes and surfaces. This category consists of slopes and plateaux covered by a mixture of fine and coarse detritus, and includes deflation surfaces and stone pavements.

In the present chapter each of the three main categories is considered separately (sections 8.2-8.4) and their characteristics are compared in a concluding summary (section 8.5). Although various authors have suggested critical slope angles at which blockfields become blockslopes and debris surfaces become debris slopes (e.g. Dahl, 1966; Caine, 1968; Stromquist, 1973; Washburn, 1973, 1979; White, 1976b), the suggested angles are highly variable (5^{o}-25^{o}) and obviously arbitrary. In the present account, therefore, blockfields and blockslopes are treated together as parts of a continuum, as are debris surfaces and debris slopes. Each section is prefaced by a review of previous work on these features.

A fairly similar field procedure was employed at each of the sample detritus slopes investigated. Slope profiles were surveyed using an Abney level and tape, with sightings taken at 5m or 10m intervals. A grid of over 30 points was established over each sample slope, and at certain points on these grids the long (a), intermediate (b) and short (c) axes of 50 to 80 stones were measured to the nearest 0.5cm. The irregular sample size results from the necessity of including in the sample 50 stones with an elongation $(a/b) \geqslant 1.5$, on which orientation and dip were measured to the nearest degree in order to determine the nature of surficial fabrics. At the remainder of the points on each grid measurements

of intermediate axis length were made on 50 surface clasts. The intermediate axis was chosen for measurement as on all slope types mean intermediate axis length for samples of 50 or more stones is strongly correlated with mean clast diameter ((a+b+c)/3) (figure 8.1).

8.2 Talus slopes

8.2.1 Previous work

8.2.1.1 Introduction

The pioneering studies of Rapp (1960a, b) on talus slopes in Spitzbergen and Scandinavia initiated two decades of intensive research on the nature of talus in periglacial areas. Rapp (1960a) restricted use of the term "talus" to " ... rock debris which has fallen down more or less continuously from a weathering rock wall", and recognised three resultant forms: uniform talus sheets, indicative of regular rockwall retreat; talus cones at the feet of rockwall chutes or funnels; and compound talus, formed through the lateral coalescence of talus cones. The same three types have been described by many later authors (e.g. Caine, 1974; Slaymaker and McPherson, 1977; Church et al., 1979). The processes operating on talus are not everywhere the same, however; White (1967), for example, distinguished rockfall, avalanche and alluvial talus in the Colorado Front Range, each type displaying features characteristic of the dominant process. Most taluses are nevertheless associated with several distinct slope facets (Caine, 1974), including a convex interfluve zone, a free-face (often stepped and gullied), a talus slope proper and a talus foot zone consisting of a fringe of large boulders resting on or embedded in solifluction or wash deposits (Jahn, 1960).

Four approaches have been adopted in research on talus slopes namely studies of morphology (especially slope form), examination of sedimentological characteristics, observations

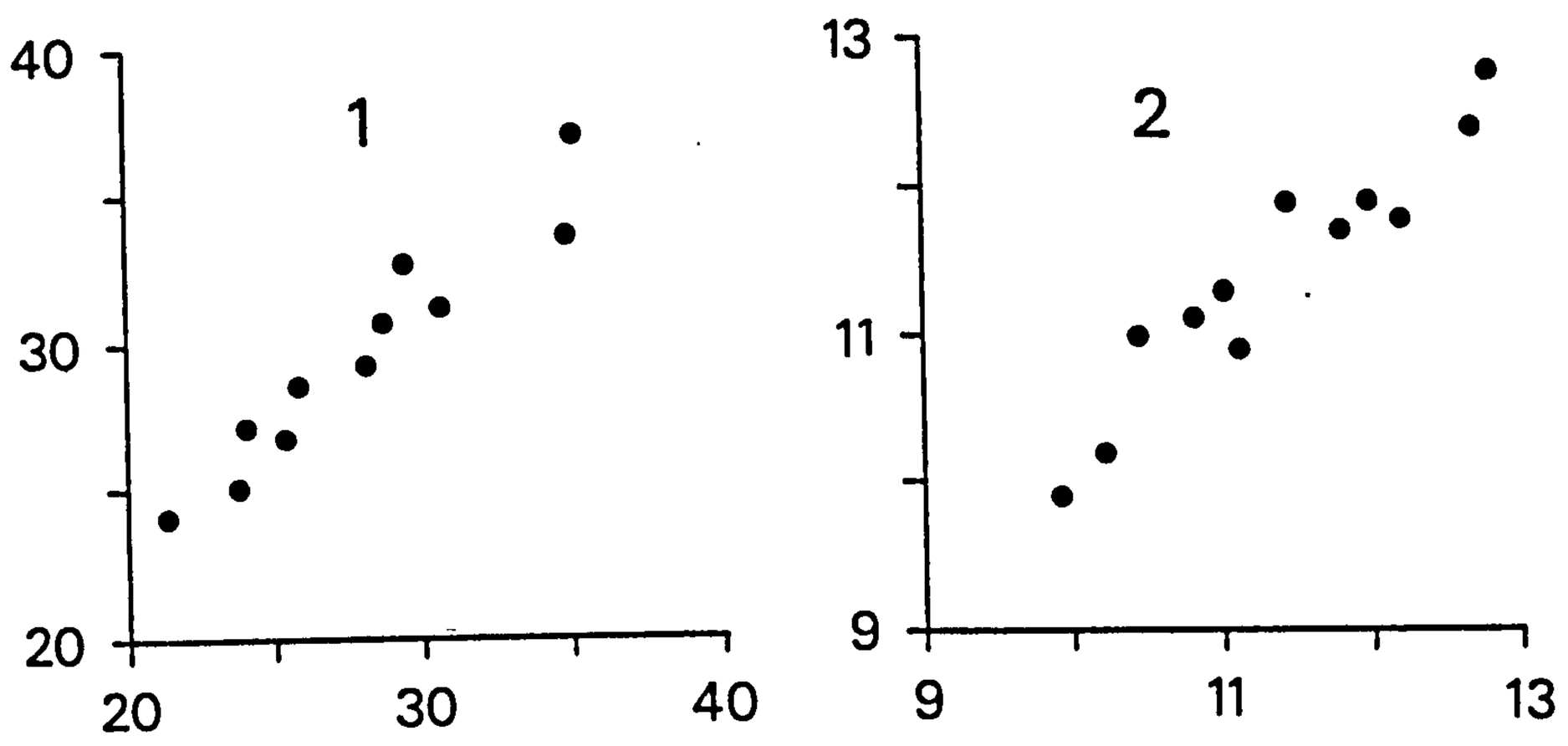


Figure 8.1: Mean clast diameter $\overline{((a+b+c)/3)}$ plotted against mean intermediate axis length ($\bar{b}$). Each dot represents the mean values for a sample of fifty clasts, taken from (1) the Glas Tholl talus cone, (2) the Coire a'Mhuilinn debris slope, (3) the Coire a'Mhuilinn talus sheet and (4) the Glas Mheall Liath blockslope.

on processes affecting talus and modelling of talus slope evolution and characteristics. This work is summarized in the following four sections.

8.2.1.2 Morphology

Talus slopes produced by the fall of individual particles from the free face and unmodified by other processes generally consist of two zones: an upper near-rectilinear slope and a basal concavity (Caine, 1969b, 1974; Howarth and Bones, 1972; Young, 1972; Statham, 1973, 1975, 1976a; Kotarba, 1976). The latter may be absent where rapid basal erosion is operative (Andrews, 1961; Howarth and Bones, 1972; Chandler, 1973). Where basal erosion is effectively inoperative, the upper rectilinear slope generally stands at a maximum angle of 35-36^{o} (Young, 1972; Chandler, 1973). Many authors have considered this angle to represent the "angle of repose" of coarse talus material (Ward, 1945; Andrews, 1961; Kotarba, 1976). However, a slope resting at a critical "angle of repose" must be approximately straight, so adherence to this concept requires the invocation of other processes to produce the widely-reported basal concavity (Statham, 1973). Moreover, the slope of taluses not subject to basal erosion is generally less than that of those that are (Howarth and Bones, 1972; Chandler, 1973), which suggests that the former rest at angles well below the angle at which failure occurs. Significantly, engineering tests have shown that talus material has an angle of residual shear of 39^{o}-40^{o} (Van Burkalow, 1945; Scheidegger, 1970; Chandler, 1973), "angle of residual shear" being the angle at which such material comes to rest after dry avalanching; the "angle of initial yield", which relates to the onset of such avalanching, is even higher (Allen, 1969; Statham, 1975). There is therefore a discrepancy of several degrees between the angles of talus slopes and the angle that might be achieved by similar material before failure occurs. Various authors have attempted to explain this discrepancy (and the

formation of a basal concavity) in terms of the operation of processes other than rockfall. Marr (1909) explained the basal concavity in terms of the "spreading" of talus under its own weight; Carson and Kirkby (1972) considered the discrepancy to result from slopewash; Chandler (1973) held creep and fluvial erosion responsible; and others (e.g. Rapp, 1960a) have explained the difference in terms of slope modification by debris flows and avalanches. Although some of these processes undoubtedly contribute to the lowering of talus slope angles, they do not operate on all taluses; hence the need for a more general explanation.

Statham (1973, 1975, 1976a; Kirkby and Statham, 1975) has pointed out that individual clasts falling on to a talus slope possess a certain amount of kinetic energy and are therefore capable of movement (mainly by bouncing) on slopes with angles less than the angle of residual shear, hence maintaining a lower "angle of repose". The gradient of rockfall talus is therefore related to the input energy of falling clasts, not (as previously thought) to the mechanical properties of the accumulated mass. The basal concavity was considered by Statham to form early in the accumulation of the talus, when the energy of input particles is generally high. He argued that as the straight section increases in length, fewer particles reach the base of the slope, so that the rectilinear slope grows at the expense of the basal concavity. Although aspects of this model have been criticized (Carson, 1977; reply by Statham, 1977), it appears to offer a much more successful explanation of talus slope angles than any of its predecessors (Church et al., 1979).

Although rockfall generally provides the main source of talus material in alpine environments (Gardner, 1969; Gray, 1973), snow avalanches transport significant quantities of material on to talus slopes in some areas (Luckman, 1971, 1972, 1977). Even "clean" snow avalanches can cause substantial modification of talus slopes by transporting detritus downslope and thereby creating slopes with a pronounced basal concavity and maximum gradients lower than those of unaffected talus

(Luckman, 1971, 1972; Gray, 1972, 1973; Caine, 1974; Kotarba, 1976; Church et al., 1979). Snow avalanches produce a number of distinctive features on the surfaces of talus slopes, such as striated and perched boulders, debris trails and crude debris lobes (Gray, 1973). Other small-scale features such as distributary lobes and levées are produced by debris flows that result from slush avalanching or heavy rainfall (Rapp, 1960b; Gray, 1972; chapter 9). In periglacial areas the basal concavity may be modified by solifluction or slopewash (Jahn, 1960; Rapp, 1960b; Wilkinson and Bunting, 1975). Various authors have also considered that creep operates on the steep upper slopes of talus (Rapp, 1960a; Jahn, 1975), and possible mechanisms for talus creep have been suggested by Davison (1888a, b) French (1976) and Statham (1977). It is difficult to envisage slow creep of tightly-wedged, angular blocks, however, and the irregular shift of surficial material observed by several authors is more readily attributable to the impact of falling rocks or snow avalanches (Gardner, 1973).

8.2.1.3 Sedimentology

Although talus slopes in periglacial environments give the impression of great thickness, this is rarely the case. Rapp (1960a) summarized data that show the range of thickness to be 1-35m, but both Young (1972) and French (1976) considered talus to be generally less than 5 m thick. In arctic areas, talus may be underlain by permafrost at shallow depth (Bones, 1972; Church et al., 1979). Stratification is generally absent. Sorting at any point on a talus slope is usually poor, but many authors have recorded "fall sorting", a general increase in clast size with distance from the free face (Andrews, 1961; Luckman, 1972; Kotarba, 1976). This may be reflected in either a linear (Caine, 1969b; Statham, 1973) or logarithmic (Gardner, 1969a; McSaveney, 1971; Church et al., 1979) decrease in particle size from the foot of the slope. Lack of fall-sorting has been considered to result from the modification of talus by processes other than rockfall, such as

dry avalanching, snow avalanching and slush flow (Bones, 1973; Statham, 1973, 1976a; Kotarba, 1976), but Luckman (1972) has shown that snow avalanching may increase the degree of downslope sorting and Caine (1969b) has demonstrated that in some instances slush flow may result in a downslope increase in clast size. Fall sorting has been explained in terms of the greater kinetic energy of large particles falling from a free face (Rapp, 1960a; Gardner, 1968; Bones, 1973), and Statham (1973) has shown theoretically that small particles are liable to high energy loss on impact and has demonstrated that most clasts come to rest amongst material of similar size. He also demonstrated that particle shape may play a secondary role, in that particles approaching sphericity tend to travel farther.

The nature of surficial fabrics on talus has been the subject of some controversy. Caine (1969a) claimed that earlier European literature and his own data indicated that talus fabrics are predominantly isotropic. In contrast, McSaveney (1971) considered that the European literature provided evidence of preferred downslope clast orientation, as did her own samples. Rapp (1960a) also thought that talus material evinced preferred downslope orientation, though Gardner (1969a) found such trends significant on only half of his samples. In attempting to reconcile these differences, Statham (1973, 1976a) concluded that fabrics with preferred downslope orientation are only found on rockfall talus that has not been subject to subsequent modification by rapid mass-movement, an idea that found support in Kotarba's (1976) discovery that snow avalanching destroys preferred trends. Statham and others (e.g. Davison, 1888a; Rapp, 1960b; McSaveney, 1971; Jahn, 1976) have attributed preferred orientation of clasts on talus to creep or sliding. This explanation fails, however, to account for the lack of preferred orientation on talus that has suffered modification by rapid mass-movement.

8.2.1.4 Activity

Many different mechanisms may initiate rockfall, including

"frost-bursting", rainstorms, pressure release, thermal changes, chemical weathering, rockfall impact, wind, snow or ice falls, earthquakes and biological activity (Rapp, 1960a, b; Luckman, 1976). In periglacial environments rockfalls are rare during periods of continuous freezing and most abundant during the spring thaw and, to a lesser extent, autumn freezeback (Rapp, 1960b; Bjerrum and Jørstad, 1968; Church et al., 1979), which suggests that temperature fluctuations about the freezing point initiate the majority of rockfalls. Gardner (1969a) found that the frequency of rockfalls in the Rockies displayed two maxima, the lesser in mid-morning (when solar radiation causes thaw) and the greater in mid-afternoon, when air temperatures are at a maximum. This suggests that thaw rather than freezing initiates rockfall, a proposition supported by Luckman's (1976) finding that the rate of rockfall in the Canadian Rockies is greater for west-facing than for east-facing slopes.

Rates of rockfall in periglacial areas have generally been expressed as rockwall retreat in millimetres per year, and exhibit a great range. A summary table in French (1976, p. 148) gives rates of 0.007 mm y^{-1} to 1.30 mm y^{-1}, and he suggested that 0.3-0.6 mm y^{-1} is typical. Conservative estimates have been given by Rapp (1960a, b) of 0.02-0.2 mm y^{-1} for Spitzbergen and 0.06 mm y^{-1} for northern Sweden, and by Gray (1972) who measured rates of 0.009 mm y^{-1} and 0.044 mm y^{-1} in the Yukon. Caine (1974) cautiously suggested that typical alpine rates range "up to 1.00 mm y^{-1}" and supplied a figure of 0.76 mm y^{-1} for the Colorado Front Range. Higher values (1.0-2.5 mm y^{-1} and 1.0-3.0 mm y^{-1}) have been suggested for the Swiss Alps (Barsch, 1977) and Polish Tatra Mountains (Kotarba, 1972) respectively. Although the overall range may partly reflect local influences (in particular the susceptibility of different lithologies to "frost bursting") it would appear that low-latitude high-altitude areas (which experience a large number of diurnal freeze-thaw cycles) are subject to higher rates of rockfall than high-latitude areas.

The movement of talus material after initial deposition (generally termed "talus shift") appears to result mainly

through dry avalanching of oversteepened slope facets (Rapp, 1960a; Bones, 1973) or readjustment resulting from the impact of a falling rock (Statham, 1973, 1976a; Kotarba, 1976). In periglacial areas the slow downslope movement of surface clasts ("talus creep") is operative on some taluses at rates of up to 250 mm y^{-1} (Rapp, 1960b; Jahn, 1975; Kotarba, 1976) but inoperative on others (Rapp, 1960a;. Gray, 1972). French (1976) considered that such creep resulted from the melting of interstitial ice, and Kotarba saw it primarily as a result of needle-ice growth. Snow avalanches, slush flows and debris flows vary in importance from area to area, having little effect in the Yukon (Gray, 1972) but acting as important transportation agents in the Rockies (Gardner, 1967; Luckman, 1971) and northern Sweden, where Rapp (1960b) reported debris flows and slush flows to be highly effective. Rapp (1960a) also considered the principal consequence of wash on talus to be the removal of material in solution, but in arctic areas where meltwater flows in subsurface channels in the active layer (Bones, 1972) appreciable quantities of fines may also be transported by wash (Wilkinson and Bunting, 1975). The deposition of these fines at the talus foot may create solifluction-prone slopes and result in formation of a low-angle wash zone (Jahn, 1960).

8.2.1.5 Models

The apparent geometric simplicity of talus has tempted many workers to attempt the construction of mathematical and graphical models of talus evolution (Fisher, 1866; Lehmann, 1933; King, 1942; Wood, 1942; Ward, 1945; Bakker and Le Heux, 1946, 1947, 1950, 1952; Bakker and Strahler, 1956; Scheidegger, 1961, 1970). However, all such models assume a straight talus slope with material accumulating at a constant angle of repose (Carson and Kirkby, 1972) and fail to explain several widely-reported characteristics of talus, namely a straight slope at an angle below that of residual shear, a basal concavity and downslope size sorting. A much more successful model has been constructed by Kirkby and

Statham (1975; Statham, 1976a) based on the energy of clasts falling from a free face. The predictions of this model were tested in the field and on a scale model, and the results indicate that it provides a reasonable approximation to the manner in which "real" talus accumulates. In particular, the model explains the origin of the basal concavity, the discrepancy between the "angle of repose" and the angle of residual shear and the cause of size sorting.

8.2.2 Talus in upland Britain

In upland Britain, the majority of talus slopes that have been studied display characteristics typical of unmodified rockfall talus: an upper straight section and basal concavity, maximum slope angles of around 36° and a degree of fall-sorting and downslope orientation of surface clasts (Davison, 1888a; Andrews, 1961; Ball and Goodier, 1970; Statham, 1973, 1976a; Mathieson, 1977; Shaw, 1977). The results of some researchers also indicate that shape sorting has operated in some upland areas, such as Lochnagar (Shaw, 1977) and Skye (Statham, 1976a). Whyte (1970) demonstrated that quartzite talus in the Mamores rests at a slightly lower mean gradient than that composed of granite, which is in turn less steep than that of mica-schist, such variations in gradient being attributable to differences in the sphericity of the talus material. Many talus slopes in Britain are, however, partly or entirely vegetation-covered (Ball, 1966; Ball and Goodier, 1970; Whyte, 1970; Statham, 1976b; Mathieson, 1977) and this has led several authors to conclude that such slopes were largely formed under periglacial conditions during the Lateglacial period (Galloway, 1958; Andrews, 1961; Ball, 1966; Ryder, 1968; Tufnell, 1969; Ryder and McCann, 1971) when macrogelivation was extremely active (chapter 6). Moreover, the formation of protalus ramparts and rock glaciers during the Lateglacial (Watson, 1966; Sissons, 1975, 1976b, 1979b, 1980) provides evidence of considerable rockfall activity at this time. From calculations based on the volume of a fossil rock glacier in Jura, Dawson (1977) suggested a

rockwall retreat rate of between 2.6 and 9.2 mm y^{-1} during the Loch Lomond Stadial, a rate well in excess of those cited for rockwall retreat in many periglacial areas at present (see above).

Well-developed talus slopes nevertheless occur in areas that were under glacier ice during the Loch Lomond Stadial (e.g. Whyte, 1970; Statham, 1976a) and these must have formed since the disappearance of the last glaciers. Several authors have expressed the opinion that limited talus accumulation is active on high ground at present (Thompson, 1950; Galloway, 1958; Godard, 1958; Ryder, 1968; Whyte, 1970; Ryder and McCann, 1971; Mathieson, 1977), but the only relevant data are observations made by Shaw (1977) in a corrie on Lochnagar. He observed nine rockfalls on occasional visits to the area but concluded that the present rate of rockwall recession is "negligible".

8.2.3 Talus in the Northern Highlands

With regard to investigation of the characteristics of talus slopes in the Northern Highlands, the concern of the present study was threefold:

(i) to determine the morphological and sedimentological characteristics of sample talus slopes;

(ii) to derive thereby some indication of the processes responsible for the formation and modification of talus in northern Scotland; and

(iii) to assess rates of rockfall activity.

Well-developed talus slopes are a common feature of the Northern Highlands, a reflection of the abundance of glacially-steepened free faces. Many of these taluses are partly or completely vegetation-covered, as in the Fannich Mountains (maps 3 and 4); others, such as the impressive taluses of Skye and Rhum (maps 7 and 8) are largely unvegetated. Examples of both types are found on An Teallach, where one vegetation-free talus and one vegetation-covered talus were investigated in detail. The vegetation-free slope is a well-

developed talus cone in Glas Tholl (NH 075844; figures 8.2 and 8.3), fed by a gulley that is incised in the steep rockwall above. It is approximately 190 m from apex to foot, faces N.N.E., has a mean altitude of 730 m and lies inside the Loch Lomond Advance limit. The vegetated talus that was investigated is located in Coire a'Mhuilinn below a 20-50 m high cliff above which there is an unvegetated debris slope. This talus ranges in length (downslope) from 85 m to 130 m, faces north, has a mean altitude of 640 m and lies just outside the limit of Loch Lomond glaciation. It is totally vegetated (mainly Festuca, Vaccinium and Empetrum spp.).

8.2.4 Characteristics of the Glas Tholl talus cone

A single slope profile was measured down the Glas Tholl talus cone by taking abney level readings at 5 m intervals (figure 8.4). This profile displays elements typical of rockfall talus, with a 130 m long upper straight section and a 60 m long basal concavity. However, 55 m from the apex of the cone there is a "step" in the profile where the slope angle increases to over 40°. This step represents the lower edge of a belt of large (up to 1.0 m^3) boulders that mark the lower limit of a zone that is entirely buried by freshly-shattered, angular, lichen-free blocks. Although lichen-free boulders are abundant downslope of this belt, these are scattered amongst lichen-covered boulders that have apparently not moved for many years (figure 8.3). Many of the blocks upslope of the belt of large boulders have clearly broken on impact, as matching fragments are found within a few centimetres of each other. This evidence suggests that the freshly-shattered material was deposited on the talus by at least one massive rockfall in the fairly recent past, a conclusion supported by the existence of an extensive lichen-free zone on the cliff above the talus cone (figure 8.3). If this interpretation is correct, then the behaviour of large-scale rockfalls on the surface of talus apparently differs from that of individual particles in that most of the rockfall material (including some very large boulders) halted no more than one

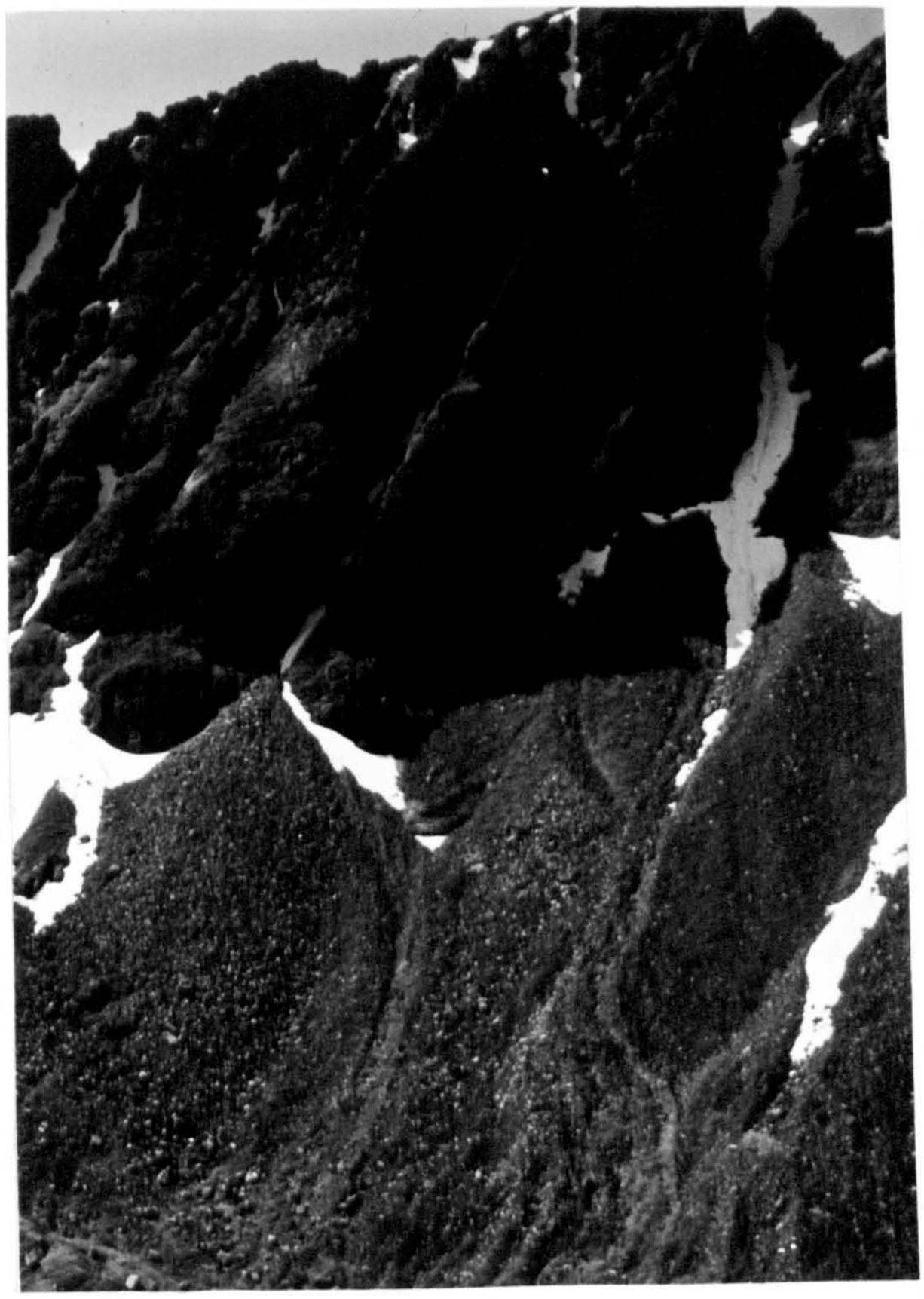


Figure 8.2: The active talus cone in Glas Tholl (left) and two less active, partly-vegetated cones, partly modified by debris-flow activity.



Figure 8.3: Surface of the Glas Tholl talus cone, looking towards the apex. Lichen-free source of recent massive rockfall is arrowed.

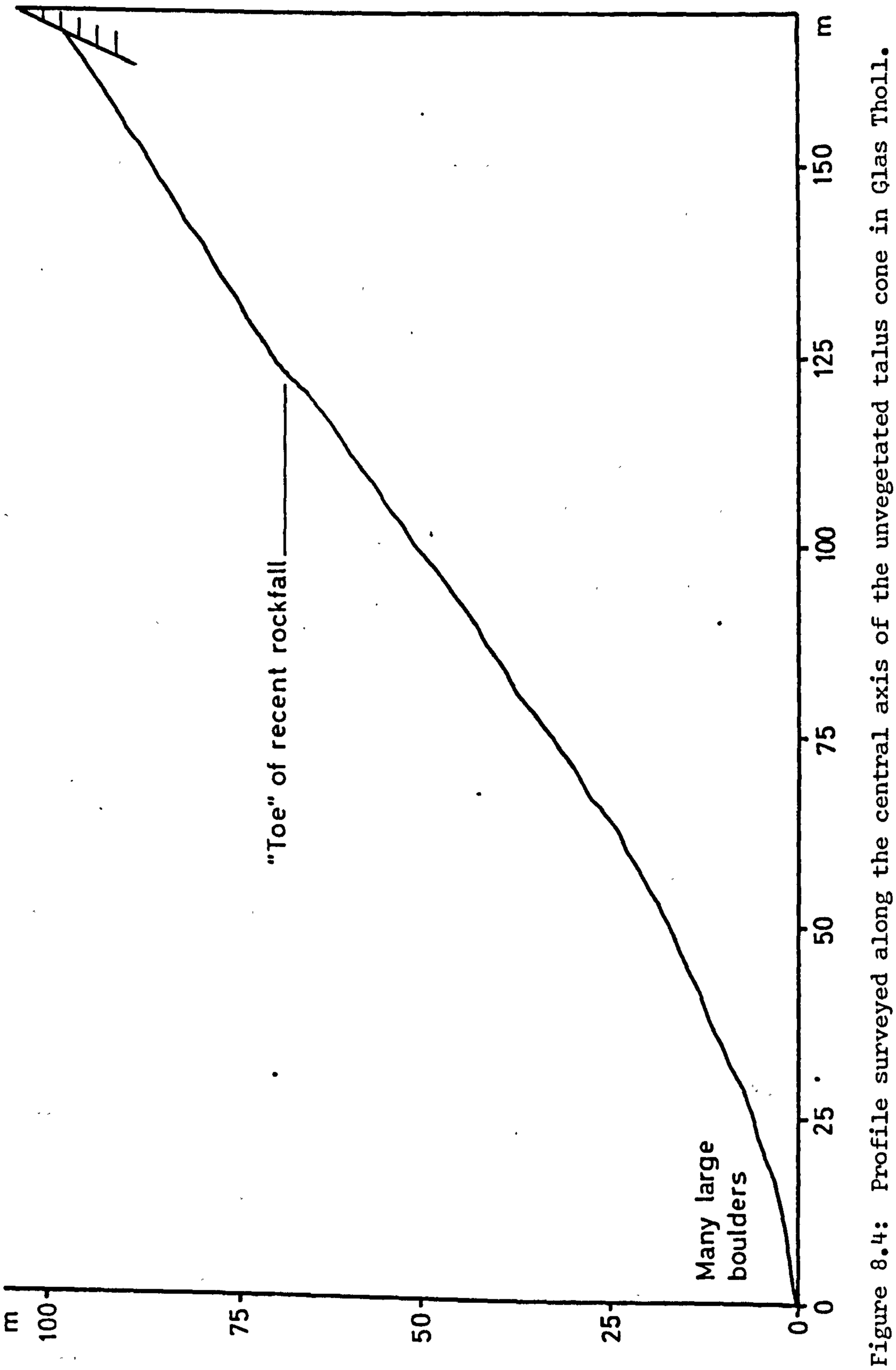


Figure 8.4: Profile surveyed along the central axis of the unvegetated talus cone in Glas Tholl.

third of the way down the cone, thereby creating an over-steepened step that is perched at an angle above the typical 36^{o} "angle of repose". A similar phenomenon was recorded by Rapp (1960b), who noted that although fall-sorting is normally evident on the slopes of Karkevagge, northern Sweden, a large rockfall stopped near the top of a talus slope. A possible explanation for the early arrest of massive rockfalls is offered by Kirkby and Statham's (1975) observation that when large particles are dropped on to a surface composed of much smaller particles a great deal of kinetic energy is absorbed on impact so that the falling mass may have insufficient energy for further travel.

Within the zone of recent rockfall debris upslope of the step on figure 8.4 fall-sorting is pronounced: mean intermediate axis length increases from 12.8 cm at a site 15 m from the apex of the cone to 23.2 cm at a point 35 m from the apex and 31.4 cm at the boulder belt 55 m from the apex. Fall-sorting is also apparent from the presence of very large boulders with intermediate axis lengths around 100 cm at the foot of the cone. However, sampling carried out on a grid of 31 points set up between the boulder belt and the foot of the slope revealed no evidence of fall-sorting (figure 8.5). The lack of fall-sorting in this zone cannot be attributed to modification of the talus by snow avalanching, as observed elsewhere (Kotarba, 1976; Statham, 1976a) as such avalanches would presumably have destroyed the fall-sorting in the zone of recent rockfall debris. Moreover, no evidence of snow avalanching has been observed in this corrie, where the headwall is too steep to allow the accumulation of large masses of snow. The existence of preferred downslope orientation in sampled talus fabrics (see below) also indicates an absence of surface modification by snow avalanching (Kotarba, 1976). There is also no evidence for disturbance by debris flow or dry avalanche activity except on the western margin of the cone, where there is a single active debris chute, but none of the sample sites lay near this chute. Explanation of the lack of fall-sorting in the central part of the talus slope probably lies in the fact that the samples included both material that

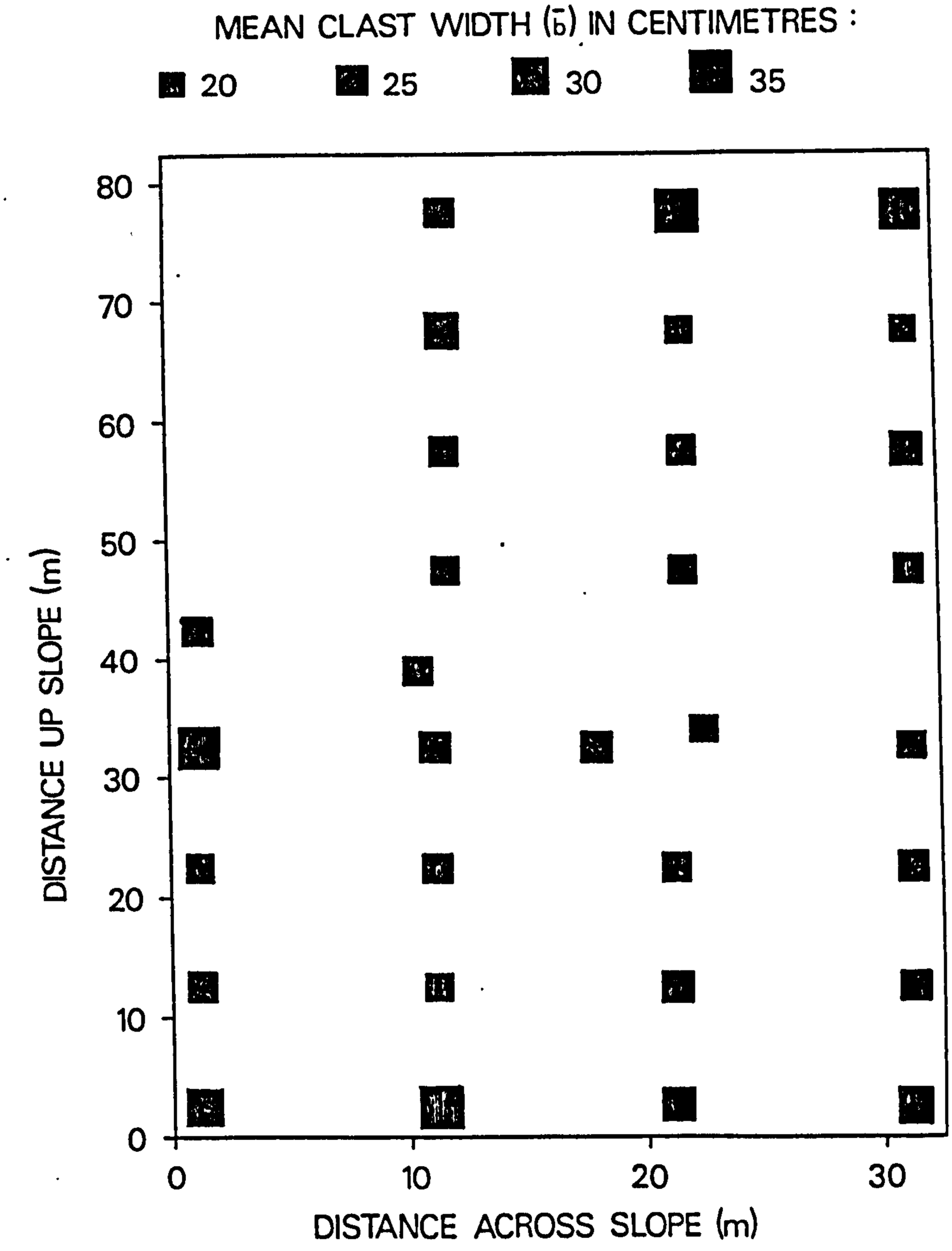


Figure 8.5: Mean clast widths (intermediate axis lengths) of 31 samples of 50+ clasts on the central section of the Glas Tholl talus cone.

has been long in situ and detritus of highly variable size produced by the recent massive rockfall(s).

Surficial fabrics were measured at eleven sites on the talus cone, all between the boulder belt and the foot of the slope (figure 8.6). The statistical significance of both dip direction and orientation was calculated using the A_n360 and A_n180 statistics (Dale and Ballantyne, 1980). All fabrics except T1 and T2 possess a preferred downslope dip direction significant at the 0.05 level, and seven of the eleven fabrics display a preferred orientation significant at the same level. In all cases preferred dip direction follows the line of maximum slope giving an overall diverging pattern that reflects the convex cross-profile of the cone. The strengths of preferred orientation and dip direction both tend to diminish with distance from the apex of the cone and away from its central axis.

Previous explanations of preferred downslope orientation of talus fabrics have involved the operation of sliding or creep at or after the time of deposition (Davison, 1888a; Rapp, 1960b; McSaveney, 1971; Jahn, 1976; Statham, 1976a). It is difficult, however, to envisage the operation of such processes on a "blocky" talus such as that in Glas Tholl where most of the clasts are tightly wedged together. These explanations also fail to account for the observed decline in preferred trends away from the central axis of the cone. A more likely mechanism lies in the adjustment of particle orientation and dip direction on final impact. In the air, falling particles may adopt any position, spinning end over end or about their principal axes (McSaveney, 1971). Those arriving with a transverse orientation are likely to be swung into a downslope orientation on final impact with protruding stones, thereby settling with an orientation approaching that of the local slope. Those spinning end over end are less likely to be spun round on final impact, as the chances of their hitting a protrusion are less. This proposition was tested using an artificial "talus" of pebbles, on to which an elongate stone of similar size was dropped; this tended to

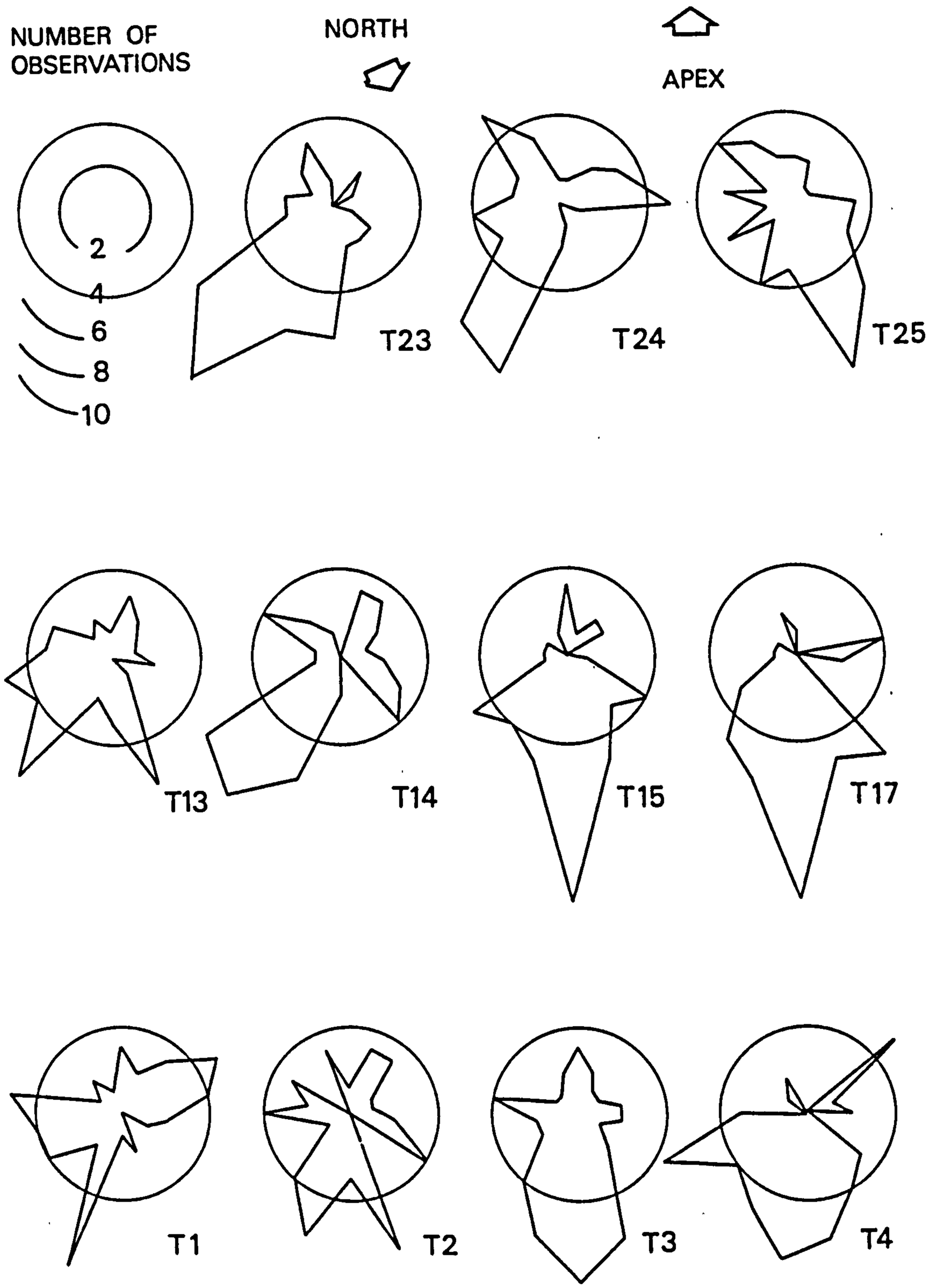


Figure 8.6: Surficial fabrics measured on the Glas Tholl talus cone. Each fabric is plotted in the correct position relative to the others and to the central axis of the cone.

adopt a downslope orientation on final impact. Cross-slope orientations (such as those that form secondary modes on fabrics T2, T3, T14 and T23) can be explained by particles being brought to rest against a large protrusion (or possibly two protrusions) and therefore not completing the swing towards a downslope orientation. This explanation also accounts for the lack of preferred trends on avalanche-modified slopes (Statham, 1973, 1976a; Kotarba, 1976). The apparent decline in the strength of preferred trends away from the central axis of the Glas Tholl cone can be related to an apparent decline in sorting at any given point. As the material at the cone margin is more variable in size than that near the central axis of the cone, the spacing of effective protrusions may be too large to cause the reorientation of relatively small blocks, or too small to allow relatively large blocks to complete their swing towards a downslope orientation.

The Glas Tholl talus cone is covered by predominantly blocky angular clasts (figure 8.3) that are very different in shape from the slabs that are produced by macrogelivation of Torridon Sandstone outcrops (chapter 6). Many of the blocks on the surface of the talus were apparently produced by the break up of larger fragments on impact during major rockfalls. Significantly, the "old" lichen-covered talus material downslope of the "boulder belt" is similar in shape to that produced by the recent large scale rockfall(s), which suggests that it may have had similar origins. This in turn implies that the major source of nourishment of the talus may take the form of infrequent large scale rockfalls, rather than the fall of individual rocks.

Sections through the surficial part of the talus (figures 8.7 and 8.9) reveal downward diminution in the average size of detritus. The basic structural framework of the talus in these sections is, however, composed of large blocks of similar size, but below a depth of about 30 cm the voids between these are infilled with smaller clasts. This pattern suggests that small particles fall down amongst the framework of larger blocks on final impact.



Figure 8.7:

Section through the uppermost 0.8 m of the Glas Tholl talus cone.



Figure 8.8:

Section through the uppermost 1.0 m of the Coire a'Mhuilinn vegetated talus sheet.

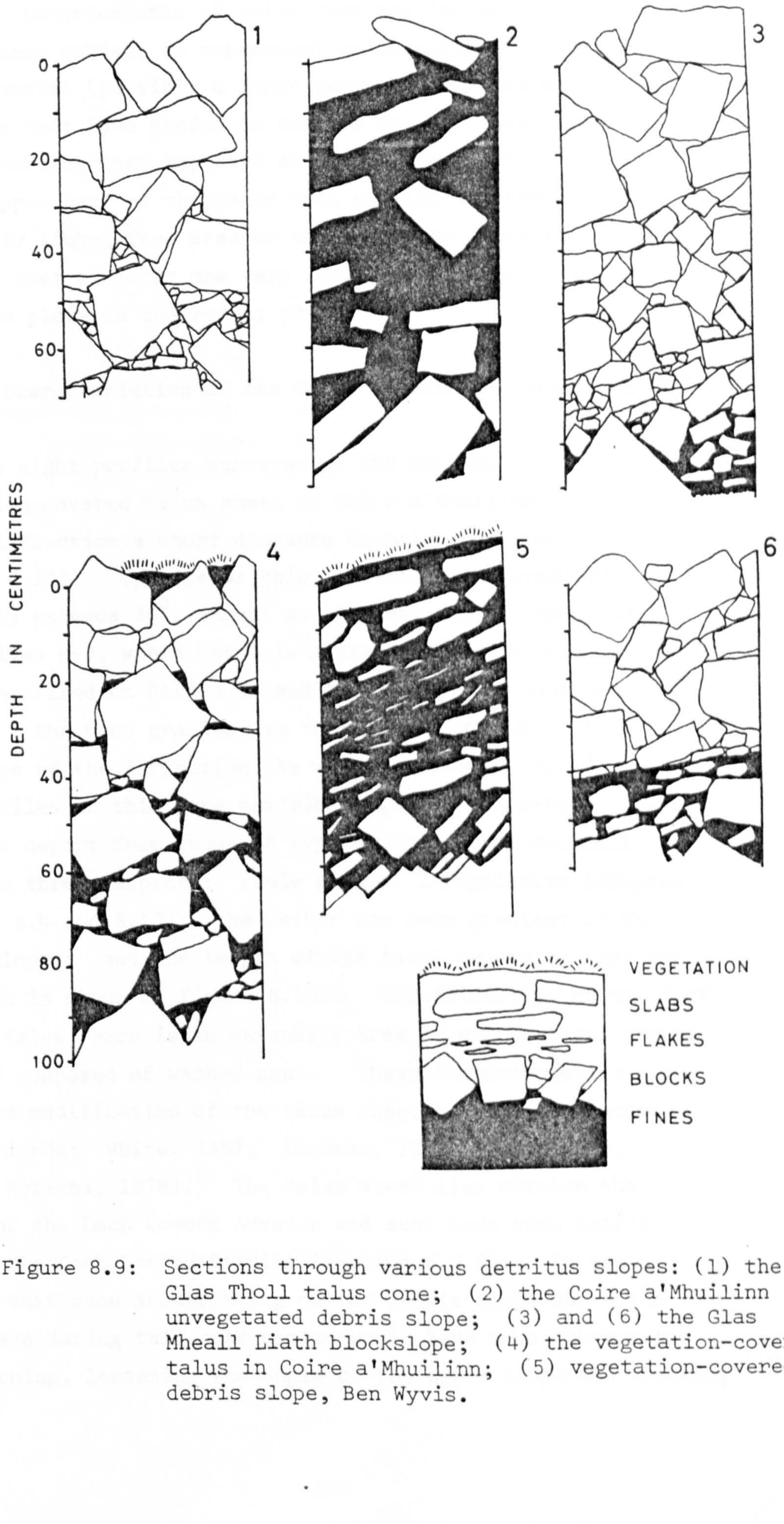


Figure 8.9: Sections through various detritus slopes: (1) the Glas Tholl talus cone; (2) the Coire a'Mhuilinn unvegetated debris slope; (3) and (6) the Glas Mheall Liath blockslope; (4) the vegetation-covered talus in Coire a'Mhuilinn; (5) vegetation-covered debris slope, Ben Wyvis.

In summary, the profile and surficial fabric of this talus cone are characteristic of talus that has formed by rockfall and not been subject to subsequent modification. Part of the talus material (possibly a large part) accumulated during major rockfalls that have tended to destroy any continuous pattern of fall-sorting that may have existed. The lack of lichens on the upper part of the talus cone and the existence of an apparently lichen-free area on the rock face above the talus indicate that at least one very large rockfall on to the cone has taken place in the recent past.

8.2.5 Characteristics of the Coire a'Mhuilinn talus sheet

The eight profiles surveyed on the surface of the vegetation-covered talus sheet in Coire a'Mhuilinn all show a marked inflection a short distance downslope of the midpoint (figure 8.10). Upslope of this inflection the gradient generally exceeds 30°, except at the very top of the talus on profiles 4-8, where there is a small convexity similar to those described on Baffin Island talus slopes by Church et al., (1979). The mean gradient in this upper zone is 33.3°. Downslope of the inflection the mean gradient is 20.9°, and the profiles in this zone describe a gentle concavity. The profiles depart from those of typical unmodified rockfall talus in three respects: their general irregularity (compare figures 8.4 and 8.10); the rather low mean gradient of the upper slopes; and the length of the basal concavity (not all of which is shown on figure 8.10). Significantly, at the foot of the talus there is an extensive area of almost level ground largely composed of washed sand. These characteristics all point to modification of the talus sheet by snow avalanches (Rapp, 1960a; White, 1967; Luckman, 1971, 1977; Gray, 1973; Kotarba, 1976). The talus sheet lies outside the limit of the Loch Lomond Advance and must have been active during the Loch Lomond Stadial (section 8.2.2). It seems likely that snow accumulating on the debris slope above the free face during this cold period would have been subject to avalanching, lessening the angle of the upper slope and creating

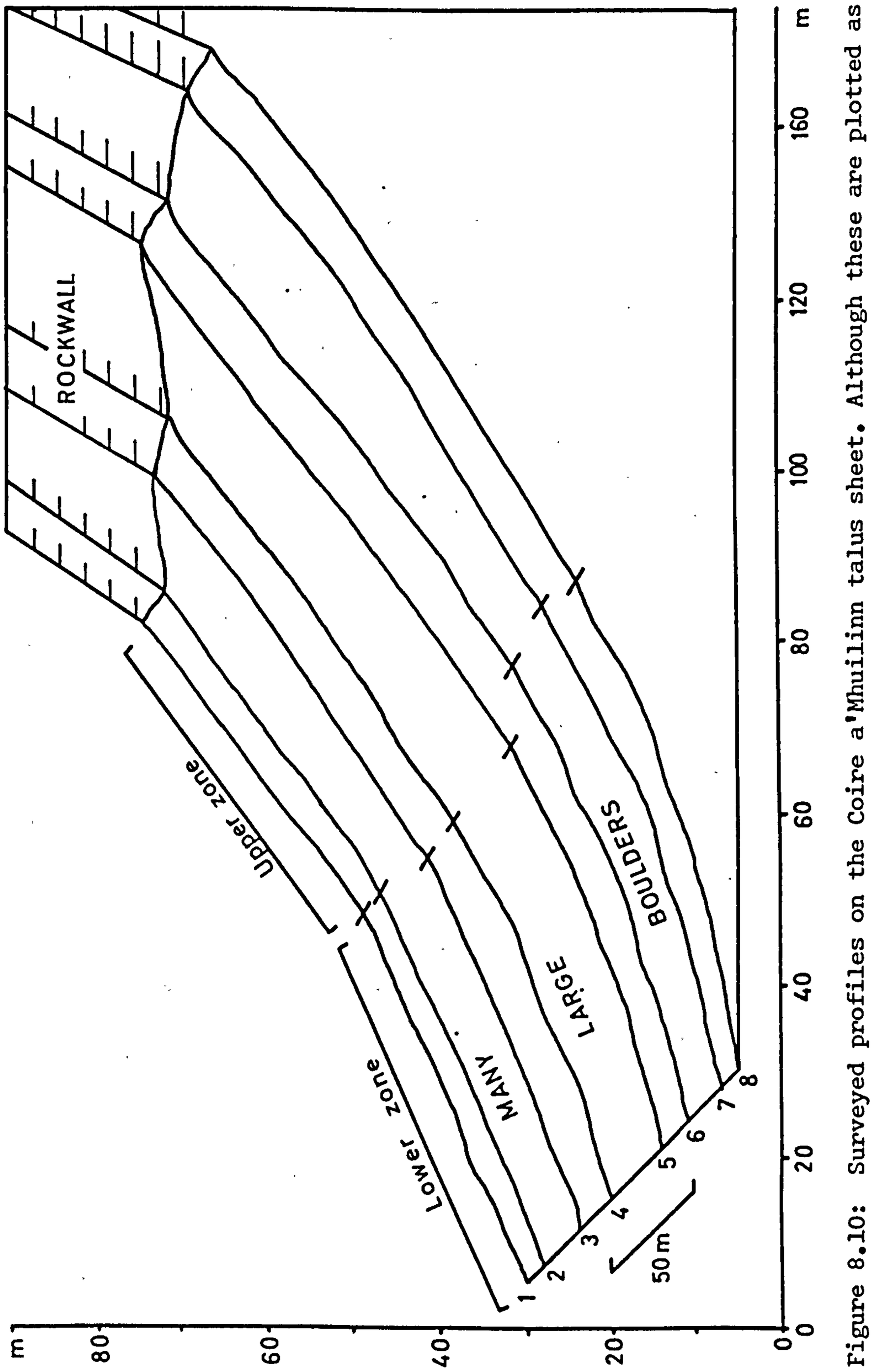


Figure 8.10: Surveyed profiles on the Coire a'Mhuilinn talus sheet. Although these are plotted as a block diagram, the profiles are undistorted.

a long trailing basal concavity with an associated wash zone. The convexity at the top of the slope probably represents the site of a perennial snowpatch (the slope faces north) during the main period of talus accumulation, probably the Loch Lomond Stadial. The presence of a snowpatch at this location would prevent the accumulation of debris at the top of the slope and thus, over time, give rise to the convexity.

Although the slope profiles of the Coire a'Mhuilinn talus sheet appear to be those of an avalanche-modified Late-glacial relict, clasts lying on top of the vegetated surface of the talus (and therefore presumably the products of relatively recent rockfalls) display the marked fall-sorting typical of unmodified rockfall talus (figure 8.11). In this case fall-sorting cannot be explained by falling clasts becoming trapped in depressions amongst material of similar size, as suggested by Statham (1973, 1976a), as the vegetated surface of the talus contains no such depressions. The fall-sorting evident in figure 8.11 can only be explained in terms of the kinetic energy of falling clasts, as suggested by Rapp (1960a): the greater the kinetic energy a falling body possesses, the farther downslope it is likely to travel.

Kinetic energy is a function of mass times the square of velocity, so the distance travelled down the slope by a falling clast will be dependent jointly on the size of the clast and the height of fall. It would therefore be expected that a plot of particle mass against distance from the top of the talus would show an envelope-type distribution: at short distances from the top of the slope particles with a small mass and low velocity (i.e. those that have fallen from a low height) would be found, whereas both large-mass low-velocity particles and small-mass high-velocity particles (and, of course, large-mass high-velocity particles) would reach the bottom of the slope.

This proposition was tested by measuring the a, b and c axes of 183 stones believed to have fallen on to the talus in a two-year period (see section 8.2.6.3 below) and the

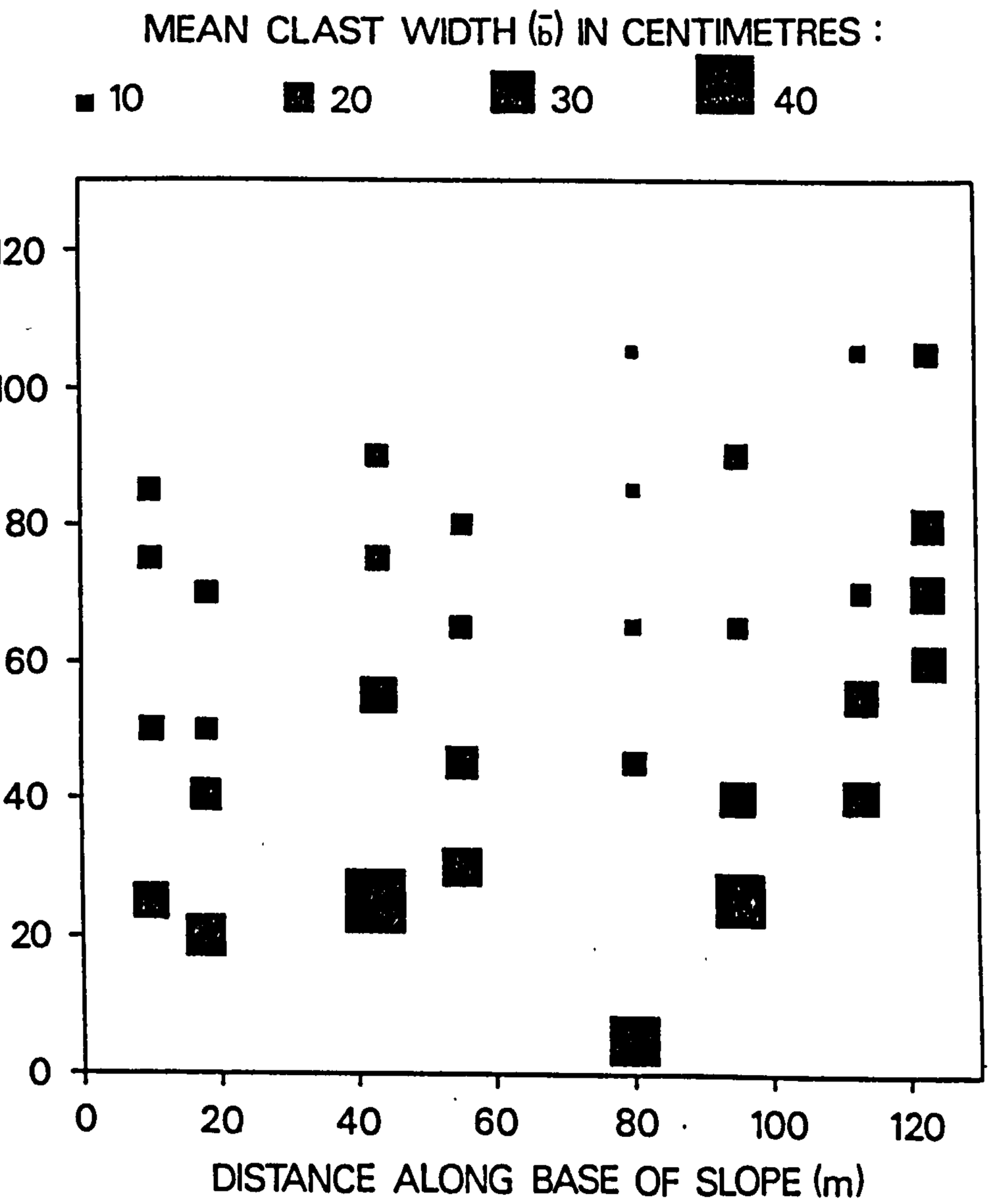


Figure 8.11: Mean clast widths (intermediate axis lengths) for 33 samples of 50+ clasts on the surface of the Coire a'Mhuilinn talus sheet.

distance from each stone to the top of the talus (to the nearest 5 m). The mass (m_i) of each stone in kilograms was calculated as

$$m_i = \frac{\rho \cdot k\,(a_i \cdot b_i \cdot c_i)}{1000} \qquad (8.1)$$

where ρ is clast density (assumed to be 2.65 g cm^{-3}) and k is an empirically-derived constant relating actual clast volume to axial measurements (a sample of 20 clasts gave a mean value of 0.696 for k; 0.7 was assumed). The results are plotted in figure 8.12. These bear out the proposition described above, in that only small clasts are found at the top of the slope, but clasts of all sizes are found at the foot.

The relationship between the mass of individual particles and the distance downslope can be approximated by the exponential relationship

$$m = 0.090\ e^{0.0126\ d} \qquad (r^2 = 0.24,\ p<0.00001) \quad (8.2)$$

Curves of a similar form may be used to approximate the relationship between the length of each axis and distance:

$$a = 6.405\ e^{0.00283\ d} \qquad (r^2 = 0.11,\ p<0.00001) \quad (8.3)$$

$$b = 4.297\ e^{0.00338\ d} \qquad (r^2 = 0.14,\ p<0.00001) \quad (8.4)$$

$$c = 1.770\ e^{0.00636\ d} \qquad (r^2 = 0.36,\ p<0.00001) \quad (8.5)$$

However, the correlation coefficients for these regressions show that a much better "fit" is achieved for the c axis than for the other two. This suggests the operation of shape sorting: particles approaching sphericity (high c:a axis ratio) have a lower coefficient of sliding friction and hence tend to travel farther (Statham, 1973, 1976a). Consideration of c-axis length alone therefore offers a higher level of explanation of the observed pattern (36%) than consideration of mass (24%), as the former effectively takes shape as well as mass into account.

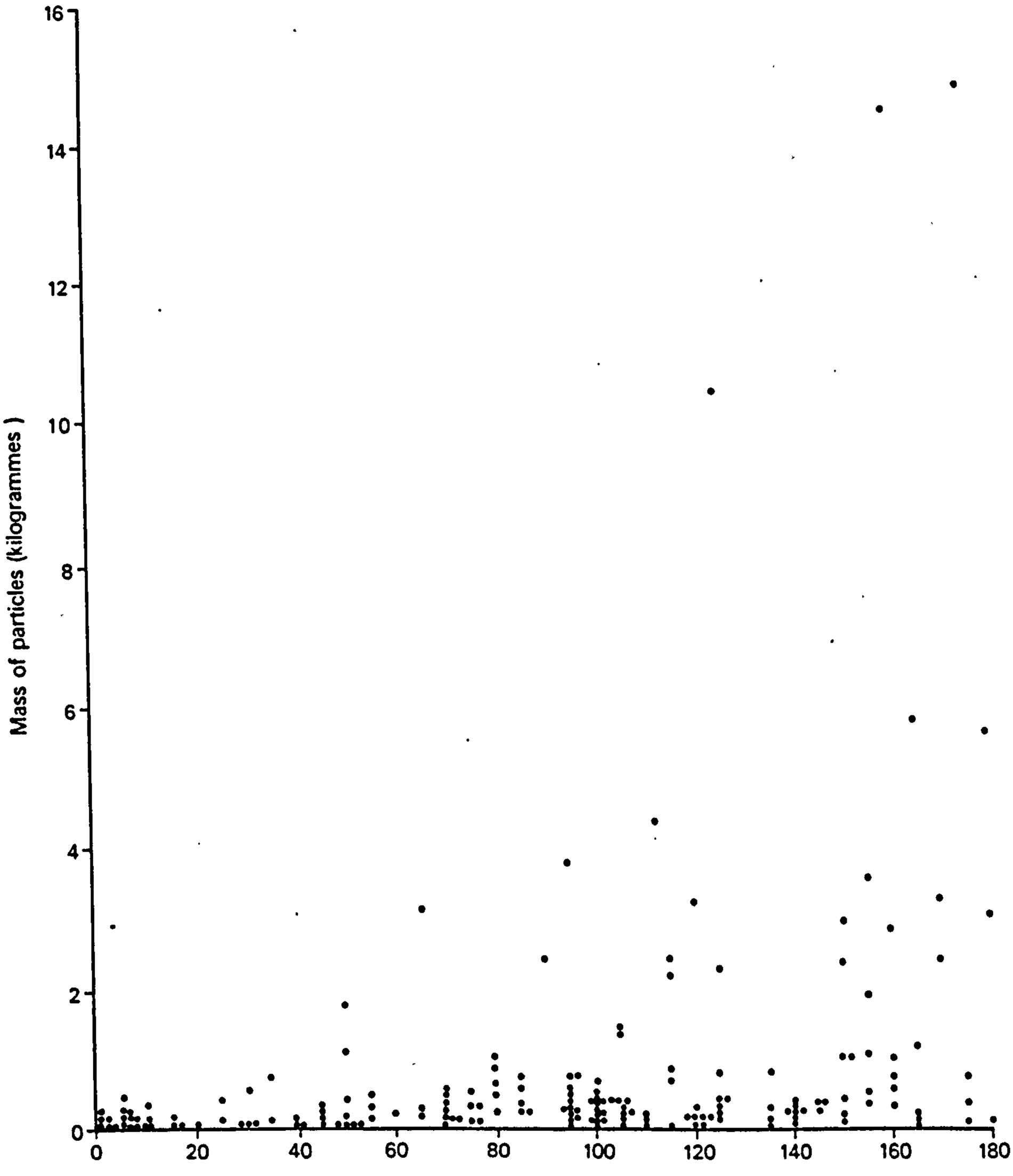


Figure 8.12: Plot of the masses of individual clasts believed to have fallen recently on to the Coire a'Mhuilinn talus against the distance from each clast to the top of the talus.

All but two of the surficial fabrics measured on this slope (figure 8.13) show a preferred dip direction significant at the 0.05 level (A_n360 statistic). A preferred downslope dip is to be expected for clasts resting on a slope; in general, upslope imbrication only occurs where clasts are resting against each other. The lack of preferred orientations significant at even the 0.1 level for five of the seven fabrics (A_n 180 statistic) provides support for the argument that preferred downslope orientation is adopted as a result of falling particles hitting a protrusion on final impact (section 8.2.4).

There are marked differences in the nature of detritus lying on top of the vegetation cover (presumably the result of relatively recent rockfalls) and that underlying the vegetation mat. The latter is blocky and angular, and very similar in appearance to material at the surface of the Glas Tholl talus cone, although on the vegetated talus the voids between blocks are largely infilled with sand (compare figures 8.8 and 8.9). However, 31% of the clasts sampled on the surface of the talus were moderately- or well-rounded by microgelivation and similar to the material on the debris slope above the free face. A further 7% were composed of Cambrian Quartzite, which crops out at the summit of the same debris slope. These observations strongly suggest that a high percentage (at least 40%) of recent rockfall detritus has fallen from the debris slope above the free face and is therefore not "primary" rockfall material in the sense defined by Rapp (1960b). No quartzite blocks and few rounded blocks were found under the vegetation cover, however, which suggests that "primary" rockfall from the free face provided most of the talus material under the vegetation mat.

In summary, it is considered that the Coire a'Mhuilinn talus slope was largely formed during the Lateglacial period, when it was fed by rockfall from the free face and modified by avalanching of snow that accumulated on the debris slope above. The formation of a complete vegetation cover on this slope indicates that activity is now much reduced. A substantial proportion of recent rockfalls has originated not

on the free face but on the debris slope above. Recent rockfall material displays pronounced fall-sorting but a general lack of preferred orientation.

8.2.6 Rates of activity

Present rates of rockfall activity on An Teallach were assessed in three ways:

(i) by making an inventory of all the rockfalls that occurred in Glas Tholl (the corrie in which the active talus cone is located) whenever this corrie was visited;

(ii) by recording the movement of painted boulders on the Glas Tholl talus cone; and

(iii) by measuring the total amount of rockfall detritus that accumulated on sections of the Coire a'Mhuillin talus sheet within a known time period.

Together, these measurements give some indication of the frequency of rockfalls (i), the magnitude of individual events (ii) and absolute rates of talus accumulation (iii).

8.2.6.1 Rockfall inventory

The results of the rockfall inventory in Glas Tholl are plotted in figure 8.14. A total of 39 falls occurred during the survey period of 58 hours, a mean rate of 0.67 rockfalls per hour. This figure falls between the values of 0.20 and 0.94 rockfalls per hour obtained by Luckman (1976) for two sites of different aspect and morphology in the Canadian Rockies. The probability of rockfall in any given hour of observation is, however, greater (0.46, as opposed to Luckman's 0.16 and 0.40 for the same two sites). Luckman found that the number of rockfalls per hour follows a Poisson distribution; in the present case the distribution of rockfalls per hour closely approximates the upper half of a normal frequency distribution (table 8.1). Caine (cited in Luckman, 1976) suggested that temporal clustering of rockfall events (as evident in figure 8.14) may be due to non-stationarity in the

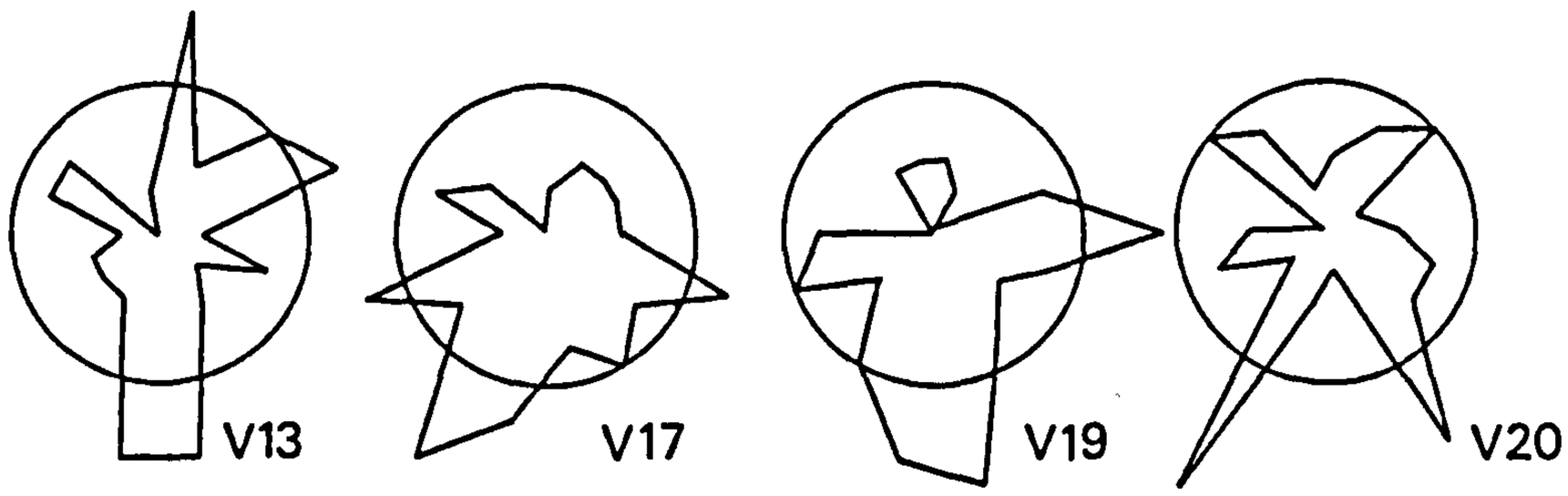


Figure 8.13: Fabrics measured for samples of 50 clasts resting on top of the vegetation cover on the Coire a'Mhuilinn talus. This position of the fabrics on this diagram is not representative of the position of the samples on the slope. The line of maximum gradient is from the top to the foot of the page. Each circle represents four observations.

Table 8.1

Probability distribution of rockfalls per hour

Rockfalls per hour	No. of observations	% of observations	Predicted percentage*
0	31	53.4	50.0
⩽ 1	50	86.2	84.1
⩽ 2	56	96.6	97.7
⩽ 3	58	100.0	99.8

* Predicted percentages were derived from the normal distribution function values given in Lindley and Miller (1971), using the upper half of the normal frequency distribution curve as a model for the probability distribution of rockfalls per hour.

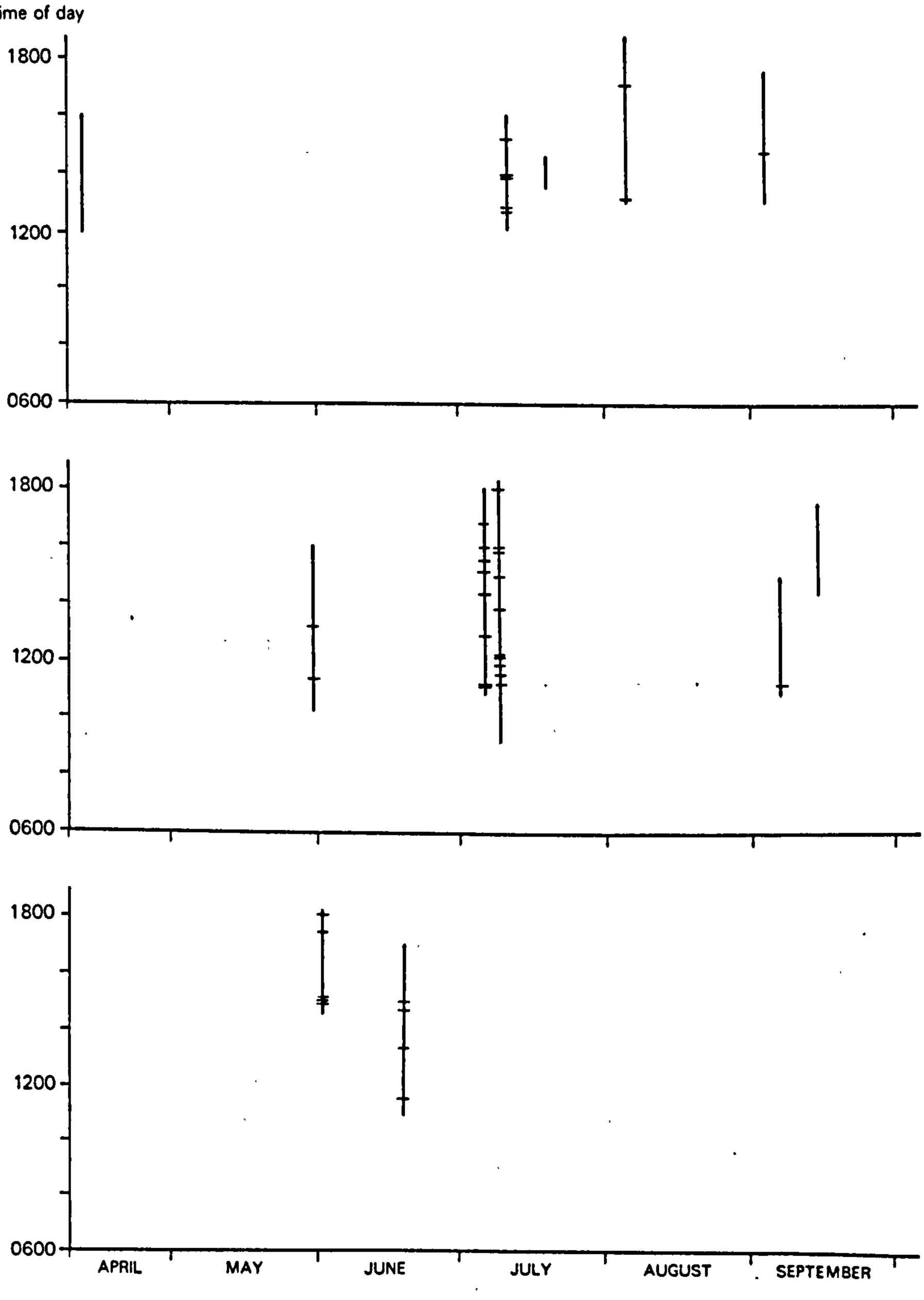


Figure 8.14: Results of the Glas Tholl rockfall inventory. The vertical lines represent periods of observation and the horizontal lines represent rockfall events that took place during these periods.

data, in that the timing of any given rockfall is not independent of that of the previous fall. In Glas Tholl, successive rockfalls were observed to take place from the same area of the corrie, which suggests that some rockfalls were triggered by the occurrence of earlier falls in their vicinity.

The inventory covers too short a period to permit the identification of seasonal or diurnal trends, but it may be significant that the three periods of maximum observed activity (050777, 080777 and 020678) were warm (17° to 20°C) and sunny. This suggests that expansion of rock on heating may have initiated rockfall. By contrast, on cool and windy days (110476, 010976, 050977 and 130977) few or no rockfalls were observed. A large proportion of the observed rockfalls involved the fall of only one clast, which sometimes initiated secondary falls on impact with loose material on steep slopes. No large-scale rockfalls were observed.

8.2.6.2 Observations on the Glas Tholl talus cone

Two lines (one red; one green) were painted across the surface of the Glas Tholl talus cone approximately 15 m from the apex on 10th July 1976. On the same date twelve large boulders on the upper third of the talus cone were marked with identifying symbols, and their positions measured with reference to fixed points on the rockwall using a 30 m tape. Twenty-five days later (040876) the lines were largely intact, but four stones at different points on the lines had been displaced between 56 cm and 210 cm downslope. Such irregular displacement suggests that movement resulted from the impact of falling particles rather than any form of slow creep. The marked boulders farther downslope had not moved. Further displacement of the painted lines was observed 29 days later (020976), when both lines showed a downslope "bulge", indicating overall displacement of about 150 cm in the centre of each line. Again, however, displacement was irregular: the maximum recorded was 420 cm for a small boulder (25 cm x 15 cm x 12 cm) but stones originally adjacent to this

boulder had apparently not moved at all. The evidence for this period (100776 to 020976) therefore suggests selective shift of surficial material at the talus apex as a result of several minor rockfalls, with maximal displacement down the central axis of the cone.

When this site was revisited after ten months, however (on 050777), both painted lines had been completely destroyed and the apex of the cone had been removed revealing the underlying bedrock. Only three of the stones from the painted lines were recovered, all from the levées of the debris chute that runs down the west margin of the cone and all more than 30 m from their original positions. The destruction of the apex of the cone and the recovery of these stones on the levées of the debris chute suggests that a major debris flow had taken place (presumably following heavy rain) sometime during the previous ten months. In addition, the downslope displacement of nine of the twelve marked boulders (figure 8.15) provides impressive evidence for a major readjustment of the upper section of the cone. Such widespread disruption cannot be attributed to debris flow activity but indicates reworking of the top of the cone by a major rockfall or snow avalanche. The addition of several large boulders near the top of the cone between 020976 and 050777 suggests that a large rockfall rather than a snow avalanche had occurred.

On 5th September 1977 replacement lines were painted across the talus at distances of 25 m and 35 m from the former apex. When these were examined after nine months (020678) the eastern and central parts of both lines were largely intact, although individual stones had moved up to 100 cm downslope. This suggests that no major rockfall or snow avalanche had occurred during the preceding winter. The marked boulders farther downslope had also not moved. The western portions of both lines had been completely disrupted, however, by the formation of a gulley 4 m wide and up to 1.3 m deep (reaching bedrock at the upper line). This had apparently been produced by debris flow: painted stones from both lines were recovered from the floor of the gulley, 4 m to 5 m downslope

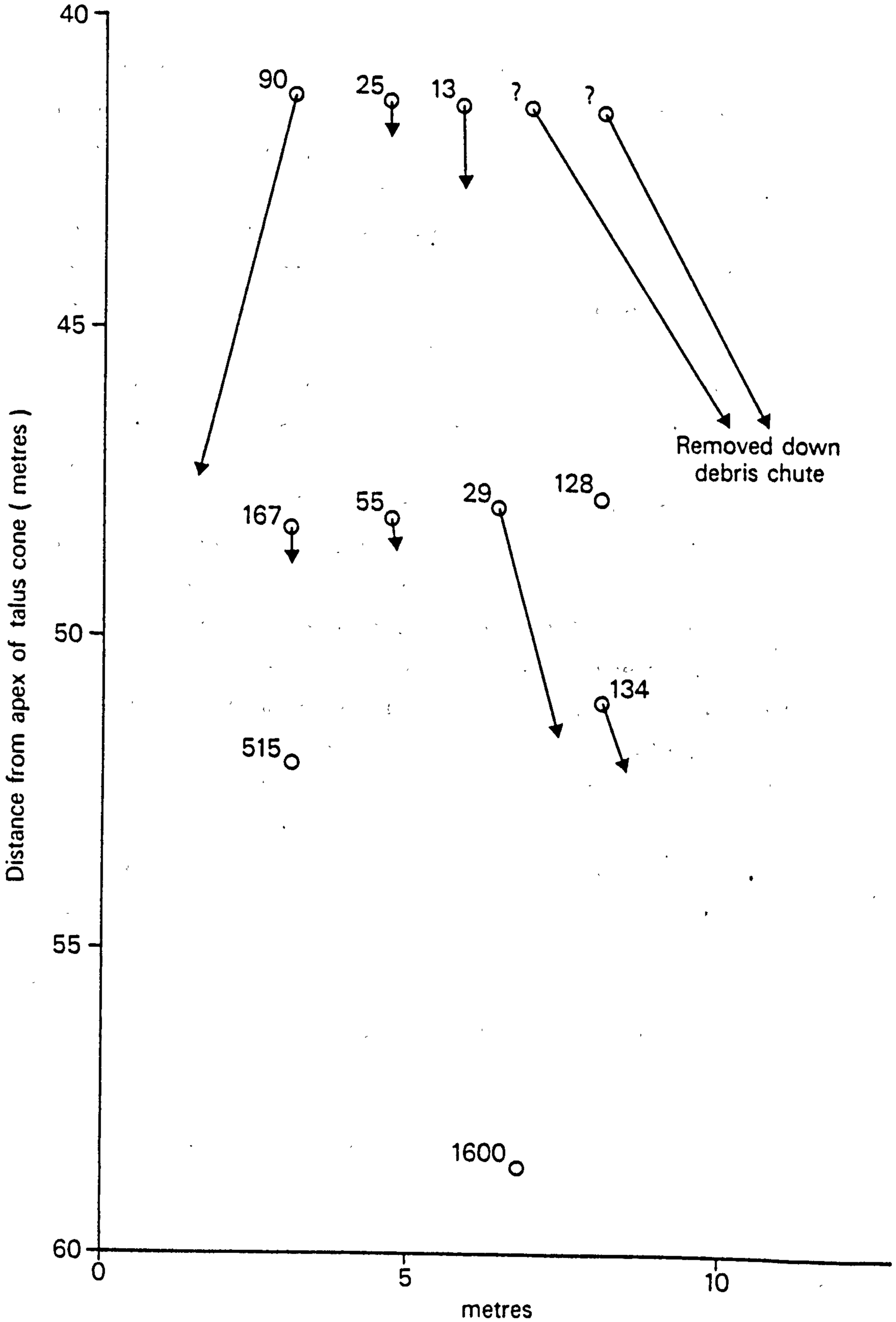


Figure 8.15: Downslope displacement of large boulders on the Glas Tholl talus cone over the period 020976 to 050777. The figures represent the masses of the boulders (kg) estimated from measurements of their principal axes.

from their original positions. A small stone that fell on to the upper part of the cone when these measurements were being recorded was observed to follow the gulley downslope.

8.2.6.3 Observations in Coire a'Mhuilinn

Rapp (1960b) estimated current rates of rockwall retreat in northern Sweden by measuring the volume of detritus deposited on snow below free faces. In Scotland, however, lack of a persistent winter snow cover precludes the use of this technique. As an alternative, the volume of rockfall detritus resting on live vegetation on talus slopes below two areas of rockwall in Coire a'Mhuilinn was calculated through measurements of all three axes of individual clasts, carried out in May 1977. It was assumed that as the clasts measured were not embedded in the vegetation cover and as the underlying vegetation had not died, then these particles must have fallen relatively recently. Some idea of the time required for vegetation covered by a stone to die off was obtained by placing marked slabs of different size on top of vegetation in September 1976. The vegetation (mainly grasses and Vaccinium) survived the summer of 1977 and the ensuing winter, but by the autumn of 1978 the vegetation under all but the smallest clasts was dead or dying. These experiments indicate that most clasts resting on live vegetation below a free face must have fallen within the previous two years.

From the axis measurements and estimates of the area (in square metres) of contributing rockwall (A_r) and of talus accumulation (A_t) at both sites it is possible to calculate rockwall retreat rate (R_r) and talus accumulation rate (R_t) in millimetres per year, assuming that all measured clasts fell within the previous two years, as

$$R_r = \frac{0.7 \sum_{i=1}^{n} (a \cdot b \cdot c)_i}{2000 \; A_r} \qquad (8.6)$$

and

$$R_t = \frac{0.7 \sum_{i=1}^{n} (a \cdot b \cdot c)_i}{1500 \; A_t} \qquad (8.7)$$

where a, b and c are the axis lengths of measured clasts in centimetres and 0.7 is an empirically-derived conversion factor relating real clast volume to axial measurements (see section 8.2.5 above). The difference in the coefficient in the denominator results from assuming that one third of the volume of accumulated talus is occupied by voids.

The results of these calculations are given in table 8.2, and indicate a present-day rockwall retreat rate of around 0.015 mm y^{-1}, one quarter of the rate estimated by Rapp (1960b) for northern Sweden (which Shaw (1977) considered representative for the corries of Lochnagar) and similar to Rapp's (1960a) lowest estimate for Spitzbergen. The representativeness of the figure obtained here may be questioned on the grounds that an unknown but substantial amount of recent material on the talus had fallen from the debris slopes above the rockwall source areas (implying overestimation), and that the figure does not take into account infrequent large-scale rockfalls (leading to underestimation). Even treated as an order-of-magnitude estimate, however, the figure vindicates Shaw's (1977) conclusion that present rates of retreat are "negligible". If present rates of activity are reasonably representative of activity throughout the 10,000 years since the end of the Loch Lomond Stadial, then the total average rockwall retreat during this period would be only about 15 cm, and the average total talus accumulation would be a mere 3.5 cm.

These values are clearly at variance with the evidence from Glas Tholl where talus slopes up to 180 m in length have developed since the disappearance of the Loch Lomond Advance glacier and where rockfalls appear to be much more frequent. The greater rockfall activity in Glas Tholl is relatable to (i) steeper and therefore less stable rockwalls and (ii) the much greater height of the free face, which is between six and ten times higher than that in Coire a'Mhuilinn. These features reflect the relative maturity of the two taluses. As Towler (cited in Chorley and Kennedy, 1971) pointed out, the relationship between cliff height and talus height on any cliff-talus system takes the form of a negative feedback loop:

Table 8.2

Rockfall measurements in Coire a'Mhuilinn, May 1977

Site	Rockwall source area (m^2)	Talus accumulation area (m^2)	No. of clasts	Volume of clasts (m^3)	Rockwall retreat rate (mm y^{-1})	Talus accumulation rate (mm y^{-1})
1	1035	7200	91	0.0337	0.0163	0.0031
2	782	3060	92	0.0209	0.0134	0.0046
1 + 2	1817	10260	183	0.0546	0.0149	0.0035

Notes: (i) "Rockwall retreat rate" and "talus accumulation rate" are calculated on the assumption that all measured clasts fell within a two-year period (see text).

(ii) The calculation of "talus accumulation rate" incorporates the assumption that one third of the talus is void space.

as the talus slope grows, the rate of growth decreases because the rockfall source area is diminished. In Coire a'Mhuilinn, the talus sheet has an average height (h_t) of about 60 m, whereas that of the rockwall (h_r) is only about 30 m, a ratio (h_t:h_r) of approximately 2:1. In Glas Tholl, however, the talus cone has a height of approximately 100 m but the rockwalls in this corrie range in height from 200-300 m, giving an average (h_t:h_r) ratio of 0.4:1. Thus although the Glas Tholl talus cone is higher than the Coire a'Mhuilinn talus, it is less "mature" in its development, presumably because talus accumulation did not take place in Glas Tholl during the Loch Lomond Stadial. The exposed rockwall in this corrie is consequently much higher and steeper than that in Coire a'Mhuilinn. It follows that the present rate of rockfall is also much greater. This comparison also suggests that although taluses outside the Loch Lomond Advance limits are not necessarily larger, the ratio (h_t:h_r) for such taluses may prove to be consistently greater than for taluses that have developed in areas that were occupied by glaciers at this time.

8.3 Blockfields and blockslopes

8.3.1 Previous research

8.3.1.1 Introduction

The term "blockfield" has been defined by White (1976b) as " ... a thin cover of blocks ... with no fine material in their upper part" and this definition is adopted here. As noted above (section 8.1), differentiation of blockslopes from blockfields on the basis of gradient is arbitrary, but for convenience the former term is here considered to refer to blockfields on slopes of 5° or more (Washburn, 1973). Following Caine (1968), "blockstream" is considered to represent a block deposit that is elongate downslope, usually concentrated in a shallow valley or gulley. Glossaries of equivalent or near-equivalent terms (e.g. felsenmeere, blockmeere, boulder field, rubble sheet, boulder run, stone run) are given in

Washburn (1973) and White (1976b).

Early hypotheses of blockfield formation invoked various non-periglacial origins (summarized in Caine, 1968), but it is now generally accepted that most blockfields in past or present periglacial areas achieved their present form through the operation of frost action and/or solifluction. Two broad categories have been recognised (Högbom, 1914; Wilhelmy, 1958; Dahl, 1966; Hamelin and Cook, 1967; Embleton and King, 1975; White, 1976b): autochthonous blockfields, in which the block deposit consists largely or entirely of the products of in situ weathering of bedrock (e.g. Svenonius, 1909; Högbom, 1914; Foslie, 1941; Lundqvist, 1948) and allochthonous blockfields, in which the blocks were originally produced by some other agency,such as deep chemical weathering (e.g. Linton, 1955; Wilhelmy, 1958; Caine, 1968) or glacial deposition (e.g. Högbom, 1914; Rapp and Rudberg, 1960; Lundqvist, 1962; Dahl, 1966).

The following review describes previous work on blockfields under four headings: sedimentology; formation and activity; distribution and significance; and previous research on blockfields in upland Britain.

8.3.1.2 Sedimentological characteristics of blockfields

The structure of blockfields appears to vary little from area to area, and to be similar for both active and inactive blockfields. One widely-reported characteristic is the appearance of fine material at variable depth. In some blockfields this takes the form of a matrix, infilling the voids between tightly-wedged blocks of similar size to those at the surface (Caine, 1968; Caine and Jennings, 1968), but in others there is a distinct downward-fining sequence (Klatka, 1961, 1962; Rudberg, 1962; Dahl, 1966; Potter and Moss, 1968; Stromquist, 1973). Similar sequences have been reported under blockstreams (Brockie, 1965; Clark, 1972; Clapperton, 1975). The lack of interstitial fines at the surface may simply reflect downwashing (Svensson, 1967), particularly

where the upper limit of fines coincides with the water table (Smith, 1953; Caine, 1968), but a downward-fining sequence implies the operation of vertical frost-sorting (Dahl, 1966; Stromquist, 1973) capable of destroying any recognizable transition from unweathered bedrock to frost-shattered blocks (Løken, 1962). Nonetheless, the size and shape of the largest blocks in autochthonous blockfields usually reflects the joint density of the underlying bedrock (Markgren, 1963; Klatka, 1962). Blockfield boulders are usually angular; roundness has been considered symptomatic of an allochthonous origin (Wilhelmy, 1958), although in some cases rounding may simply reflect susceptibility to microgelivation (Bout and Godard, 1973; Clapperton, 1975), as may the abundance of fines at depth (Rudberg, 1962). These sometimes include clays that may be residual from deep weathering (Caine, 1968), although the preponderance of illite in most blockfield clays may indicate that these were produced under cold climate conditions (Klatka, 1962; Caine, 1968).

Various surficial structures on blockfields and blockslopes indicate frost sorting and/or downslope movement. These include large sorted circles (Dahl, 1966), "islands" of fine material (Lundqvist, 1962; Rudberg, 1962), sorted stripes (Rudberg, 1962; Dahl, 1966) and boulder lobes and terraces (Rudberg, 1962, 1964; Caine, 1968; Clark, 1972; Clapperton, 1975). Downslope movement of surficial material is also indicated by blockslope fabrics. Although these may be isotropic in the upper parts of blockslopes, a preferred trend is evident over most areas, and tends to increase in strength downslope (Cailleux, 1947; Wahrhaftig, 1949; Denny, 1951; Klatka, 1961, 1962; Caine, 1968; Potter and Moss, 1968). The typical surficial fabric has been described as "a semi-circular normal distribution displaced from local slope direction by up to 20°" (Caine, 1972) although radial trends may be found associated with lobes and terraces (Rudberg, 1964). A secondary characteristic of blockfield fabrics is that the majority of surface particles are imbricate upslope (Cailleux, 1947; Klatka, 1961, 1962).

The thickness of blockfields is variable: in Scandinavia, Stromquist (1973) found a fairly uniform mantle averaging 0.7 m in depth, and Dahl (1966) described an average depth of 1.0 m (maximum 1.7 m). Clapperton (1975), however, reported depths of over 3.0 m in the Falkland Islands.

8.3.1.3 Blockfield formation and activity

In his classic paper of 1906, Andersson ascribed the formation of the Falkland Island stone runs (blockstreams) to macrogelivation of the underlying quartzite, solifluction of the weathered mass and immobilization as a result of the eluviation of fines. Subsequent investigation of the same features (Joyce, 1950; Clark, 1972; Clapperton, 1975) have added little to this initial explanation. The three themes involved have been repeatedly invoked, with relatively minor variations, to account for blockfield formation elsewhere, and a fourth process, the upfreezing of blocks (vertical frost sorting) has been added to this list to account for the downward fining characteristic of many blockfields (Jahn, 1961; Richmond, 1962; Dahl, 1966; Caine, 1968; Stromquist, 1973).

Variations on this model have been mainly concerned with the origins of the blockfield material and the nature of downslope movement. The majority of blockfields studied are autochthonous, and the coarse material in these has apparently been produced by mechanical disintegration of chemically fresh bedrock (Joyce, 1950; Klatka, 1961, 1962; Dahl, 1966; Bout and Godard, 1973; Clapperton, 1975). Caine (1968), however, considered that the block deposit in Tasmanian blockfields may have been produced by deep chemical weathering, and various authors have mentioned blockfields formed from frost-sorted till and other initially unsorted detritus (Richmond, 1962; Dahl, 1966). The present author has observed a well-developed blockfield of rounded granite erratics covering sedimentary bedrock on Ellesmere Island. Microgelivation is generally considered to be responsible for the fines that are found at depth in blockfields, although Klatka (1961, 1962) considered that those in the Polish Carpathians contained fines derived from till and loess as well.

Troll (1947) observed that blockstreams are capable of moving several kilometres over fairly low angle slopes, and Williams (1968) described several examples of such blockstreams from Southern England. Caine (1968) deduced from the surficial fabrics of Tasmanian blockfields that movement must have taken place through "plastic deformation of a flowing mass" and concluded, following Andersson (1906) and Klatka (1961, 1962) that this took the form of solifluction at a time when fines were more abundant. Like Brockie (1965) and Bout and Godard (1973) Caine also entertained the possibility that flow took place when ice filled the voids between the blocks, as it does in rock glaciers. He rejected this explanation, however, on the grounds that thermokarst features were absent and that the blockfields studied were generally dissimilar to rock glaciers. Richmond (1962) and Dahl (1966) considered that blockfield movement occurred as some form of creep, and Rudberg (1964) described a possible mechanism for creep in the frost heave of underlying fines. The validity of these conflicting explanations is difficult to assess.

Most writers are agreed, however, as to the effectiveness of subsurface wash in the eluviation of fines from blockfields (e.g. Jahn, 1961; Klatka, 1961; Kachurin, 1964; Dahl, 1966; Caine, 1968, 1972; White, 1976b) aided possibly by piping (Smith, 1968) and subsurface mudflows (Stromquist, 1973). Eluviation of fines has been considered by authors advocating movement by solifluction as the major cause of blockfield immobilization.

There are few accounts of rates of blockfield formation or activity. King and Hirst (1964) concluded that a blockfield on the Åland Islands (S.W. Finland) had formed in the last 6,000 years, and some in Norway appear to be postglacial in age and still active (Dahl, 1966). The only direct measurements of downslope movement, however, are those of Rudberg (1964), who found displacement to be irregular and intermittent, with mean rates of 1.0-5.0 mm y^{-1}. Jahn (1961) measured wash yields of 1.0-16.3 g $m^{-2}y^{-1}$ from "rubble" in Spitzbergen, and attributed the efficacy of this process to the lack of protective vegetation, high precipitation and rapid meltwater runoff.

8.3.1.4 Distribution and significance

Blockfields may occur in both lowland and highland areas, but are most common in the latter (White, 1976b). In Scandinavia, where they have been more exhaustively studied than anywhere else, the lower limit of blockfields exhibits considerable local variation, due partly to lithological differences and partly to the presence of allochthonous blockfields derived from till at lower levels (Rapp and Rudberg, 1960; Rudberg, 1962; Dahl, 1966; Bout and Godard, 1973). The lower limit of autochthonous blockfields (broadly equivalent to the "frost shatter zone" described by various Scandinavian authors) nevertheless declines to the north and west, being at sea level in Spitzbergen, 500 m in Lapland and 1,600 m in the Dovrefjell area (Högbom, 1914; Rudberg, 1962). The climatic significance of this trend has, however, been questioned by Ives (1966) who considered annual rather than diurnal freeze-thaw cycles to be of primary importance in blockfield genesis.

Although many Scandinavian blockfields show evidence of movement at present (Rudberg, 1962, 1964; Dahl, 1966), Stromquist (1973) considered that they were formed under permafrost conditions, and that blockfield depth (on fairly level ground) provides an indication of the depth of the former active layer. The presence of relict ice-wedge polygons (Svensson, 1967) and large scale sorted patterned ground (Dahl, 1966) on Scandinavian blockfields indicates formation under permafrost conditions (Washburn, 1979), and it is notable that cryoplanation terraces, forms believed to undergo a genesis similar to that of blockfields, are now thought to form under permafrost conditions (Reger and Péwé, 1976). In Scandinavia, therefore, blockfield formation is generally considered to have taken place in Lateglacial or early postglacial times (Dahl, 1966; Svensson, 1967). Earlier authors (e.g. Dahl, 1956) proposed that blockfields are indicative of areas that were not ice-covered during the last glaciation, and more recently it has been proposed that blockfields may survive glaciation (Ives, 1966; Hamelin and

Cook, 1967). Although this may be the case in some areas, it seems unlikely to be so in Scandinavia, where blockfields appear to be of fairly uniform shallow depth. In areas beyond the margins of the last ice sheet, blockfields may be of much greater age: Klatka (1962) considered that those of the Carpathians may have formed in Riss-Saale (Wolstonian) times.

8.3.1.5 Blockfields in upland Britain

Although mountain-top detritus is widespread in Scotland, Wales and the Lake District, true blockfields and blockslopes (with an absence of surficial fines) are relatively rare, being confined to rocks that have proved more susceptible to macrogelivation than ~~microgelivation~~ granular disintegration (chapter 6), particularly quartzite but also microgranite, felsite, porphyrite and (in some cases) granite (Crampton, 1911; Crampton and Carruthers, 1914; King, 1968; Ryder, 1968; Kelletat, 1970a, b; Whyte, 1970; Ryder and McCann, 1971; Ballantyne, 1977; Ballantyne and Wain-Hobson, 1980). Sedimentological information on these blockfields is sparse. Crampton (1911; Crampton and Carruthers, 1914) found a white silt layer below quartzite blockfields in Caithness, and King (1968) described a downward-fining sequence under massive blocks (mean diameter 1.0 m) in the Cairngorms, indicating similarities with structures described from elsewhere. Most blockfields have been described as the products of large-scale frost shattering (Crampton, 1911; Crampton and Carruthers, 1914; Bailey and Maufe, 1916; Marr, 1916; Galloway, 1958; King, 1968; Ryder, 1968; Ryder and McCann, 1971; Ballantyne and Wain-Hobson, 1980) but several authors have considered that some may be derived from till (Tufnell, 1969; Ball and Goodier, 1970), although no evidence has been given. Galloway (1958) also entertained the notion that blocks on the Cairngorms may have been produced by deep weathering, along the lines suggested by Dahl (1956). This idea has also been proposed by Sugden (1970b, 1971) but it seems unlikely that interglacially- or preglacially-weathered rock would have survived <u>in situ</u>

during glaciation. The rounding of surficial blocks, and indeed tors (Linton, 1949, 1955) on the Cairngorms is more readily attributable to microgelivation than deep chemical rotting.

Although claims of recent downslope movement have been made for several blockslopes (Crampton, 1911; Crampton and Carruthers, 1914; Clark, 1962; Whyte, 1970) no supporting evidence has been provided. Removal of fines by wash has also been widely postulated (Crampton, 1911; Ryder, 1968; Ryder and McCann, 1971) and cited as a cause of immobilization by Clark (1962), but again evidence is lacking. In general, a Lateglacial age is favoured for blockfield formation (Galloway, 1958; Ryder, 1968; Potts, 1971; Ryder and McCann, 1971). Ballantyne and Wain-Hobson (1980) described large-scale sorted circles and stripes on the blockfield of Sron an t-Saighdeir, Rhum (figures 8.16 and 10.9) and concluded that blockfield formation took place under permafrost conditions during the Loch Lomond Stadial.

8.3.2 Blockfields and Blockslopes in the Northern Highlands

The distribution of blockfields and blockslopes within the study area is limited to certain lithologies. True blockfields and blockslopes are absent from the Torridon Sandstone and basal quartzite of An Teallach, where sand fills the interstices between clasts at the ground surface, but are superbly developed on fine-grained Cambrian Quartzite (map 2). Indeed, well-developed blockfields appear to characterize all areas of fine-grained quartzite above 500-600 m in Scotland, other examples being the quartzite hills of Caithness (Crampton, 1911; Crampton and Carruthers, 1914), the summit cap of Scùrr a'Mhaim in the Mamores (Whyte, 1970), the upper parts of Arkle, Foinaven and Ben More Assynt (Sutherland) and the summit ridges on Liathach and Beinn Eighe (Torridon). This characteristic of quartzite is not confined to Scotland (Bout and Godard, 1973).

In the Fannichs blockfields and blockslopes are poorly developed and restricted in distribution to the siliceous rocks of Mheall a'Chraisgaidh and Beinn Liath Mhór (map 4);

on Ben Wyvis they are entirely absent from the main ridge but strikingly well-developed on the granulite of Càrn Gorm (map 5). On Rhum (maps 7 and 8) blockfields and blockslopes are associated with microgranite on the Western Hills (figure 8.16) and felsite in the Rhum Cuillin (Ryder, 1968; Ryder and McCann, 1971; Ballantyne and Wain-Hobson, 1980). Rock type is clearly the primary determinant of blockfield distribution in the study areas and lack of suitable lithologies at low altitudes precludes detailed analysis of the lower limits of blockfield development. On Rhum blockslopes descend to 230 m, where they are truncated by sea cliffs, which suggests that blockslope formation may have been possible at even lower altitudes here. On An Teallach the lower limit lies between 460 m (the highest point on the quartzite escarpment east of the massif, on which blockfields are absent) and 670 m (the lowest extent of blockfield on Glas Mheall Liath; at lower altitudes the bedrock is Torridon Sandstone). On Càrn Gorm, Ben Wyvis, the lower blockslope limit is not lithologically controlled and reaches 590 m. In the author's experience blockfields and blockslopes are rarely found below 500 m in highland Scotland. The low altitude of the Rhum feature may relate to the exposed nature of the site or may reflect the susceptibility of microgranite to macrogelivation.

8.3.3 Characteristics of certain blockfields

The sedimentological characteristics of blockfields in the study area were investigated through the excavation of pits in level areas of the blockfields on Càrn Gorm (Ben Wyvis), Glas Mheall Liath (An Teallach) and Mheall a'Chrasgaidh (Fannichs) at 730 m, 950 m and 930 m respectively. Representative sections are depicted in figure 8.18.

The Càrn Gorm blockfield is the most extensive in the study area (though less extensive than that on Sron an t-Saighdeir, Rhum) and is underlain by steeply-dipping granulite. It is of very variable depth: the path cut through the blockfield reveals bedrock at a depth of 30 cm in places, but excavation was stopped at 160 cm with no sign of bedrock below. Although the



Figure 8.16: The microgranite blockfield at 510 m altitude on Sron an t-Saighdeir, Rhum.



Figure 8.17: Pit excavated in granulite blockfield at 700 m altitude on Càrn Gorm, Ben Wyvis, showing diminution in clast size with depth.

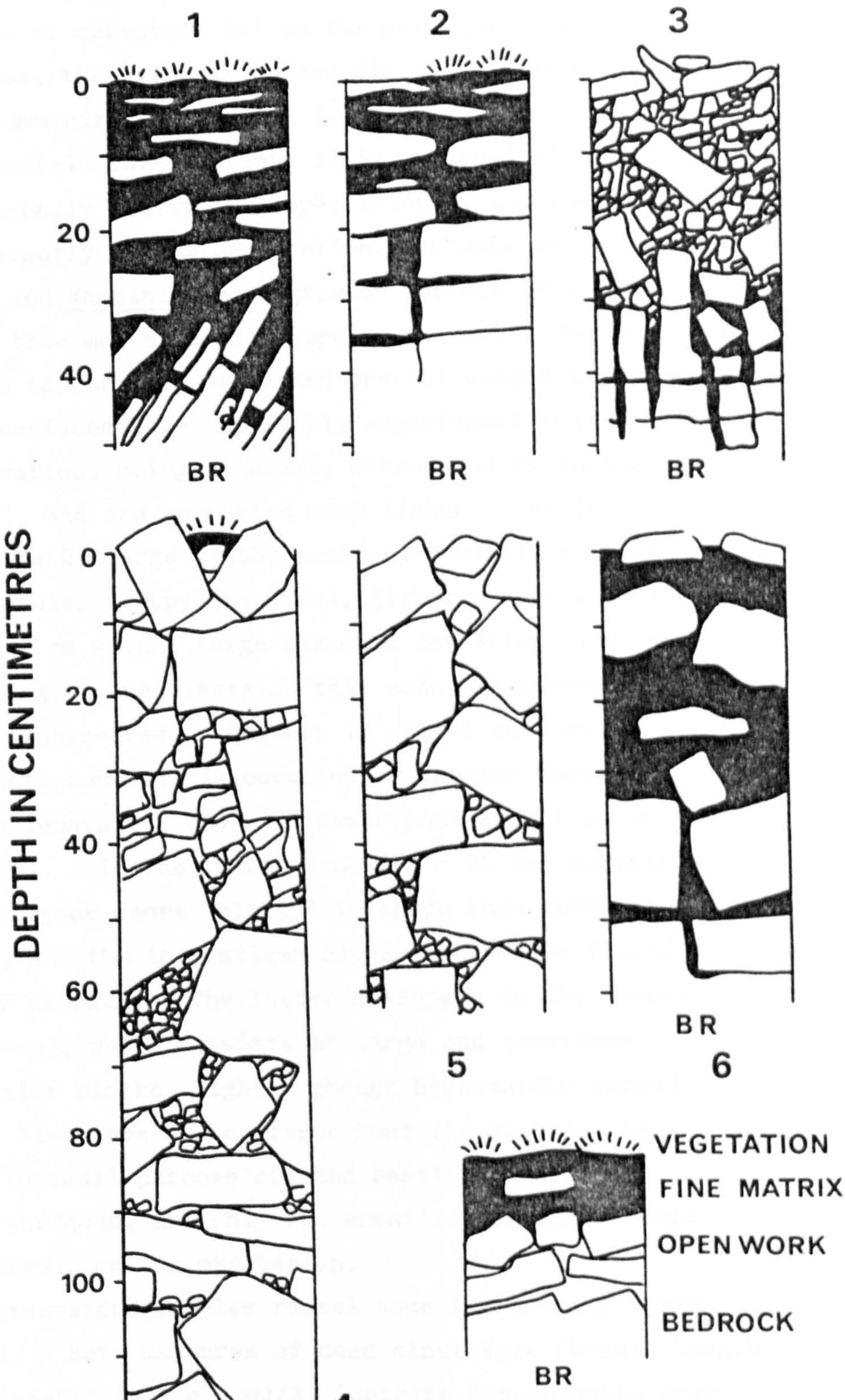


Figure 8.18: Sections trenched in blockfields and debris surfaces. 1, 2 and 6 are debris surfaces on Ben Wyvis, Scurr Breac (Fannichs) and An Teallach respectively. 3, 4 and 5 are blockfields on Mheall a'Chrasgaidh (Fannichs), Carn Gorm (Ben Wyvis) and Glas Mheall Liath (An Teallach) respectively.

section revealed no well-defined stratigraphy, it is convenient to describe its structure by zones. The sedimentological characteristics of three of these zones have been quantified by measuring the axes of a sample of stones, together with minimum radius of curvature (r) on the principal plane and, for two samples, the orientation and dip of 50 clasts. The results are summarized in table 8.3.

The blockfield surface (zone 1) is composed of large boulders, generally slabby in shape; lying at low angles of dip with apparently random orientation. Clumps of Rhacomitrium and Vaccinium spp. grow on pockets of mineral soil, though fine material is generally scarce. The boulders simply lie on top of each other and are not wedged together. Their upper surfaces have apparently experienced rounding due to microgelivation, being generally subrounded or subangular (figure 8.17), and are encrusted with lichen. At depths of 0 - 15 cm (zone 2) large slabby boulders persist, although occasional cobbles occupy the voids, giving a mean b-axis length of 21.1 cm with a large standard deviation, indicative of poor sorting. The clasts in this zone are generally angular and lichen-free. Between 15 and 45 cm sorting improves; this zone (3) is occupied by angular cobbles (mean b-axis length 8.0 cm) with fewer boulders (figures 8.17 and 8.18). In the fourth zone (45 - 85 cm) angular boulders reappear, more "blocky" in shape than those near the surface, and the interstices between these are filled with angular pebbles. The latter disappear in the lowest zone (> 85 cm), which consists of large and sometimes massive angular blocks, tightly though haphazardly wedged together. Fines are scarce throughout the section, being restricted to small patches of sand resting on near-horizontal surfaces, and only two erratics (of Moine Schist) were recovered from the excavation.

The three stone samples reveal some interesting trends (table 8.3). Both measures of mean clast size (b-axis length and mean diameter (a + b + c)/3) indicate a pronounced drop in the mean size of material between 10 cm and 40 cm depth, with little further change at 80 cm. However, the sample

Table 8.3

Variations in clast size, shape and fabric with depth in the Càrn Gorm blockfield, Ben Wyvis

Characteristic	Parameter	Sample 1	Sample 2	Sample 3
Depth (cm)		10	40	80
Sample size		68	67	50
Intermediate Axis Length (cm)	Mean	21.1	8.0	7.0
	Std. Dev.	13.4	3.2	7.2
$\frac{(a + b + c)}{3}$ (cm)	Mean	19.8	8.2	7.1
	Std. Dev.	11.1	2.8	6.8
(a + b) / 2c	Mean	4.9	3.2	2.9
b / a	Mean	0.68	0.65	0.67
c / a	Mean	0.23	0.34	0.39
1000 (2r / a)	Mean	59.4	73.8	68.9
Dip (degrees)	Mean	15.8	27.6	N.M.
A_n 180	Score	0.50	0.40	N.M.
	Significance	N.S.	N.S.	N.M.
A_n 360	Score	0.11	0.49	N.M.
	Significance	N.S.	N.S.	N.M.

N.M. - Not Measured. N.S. - Not Significant.

from 10 cm depth is poorly sorted (coefficient of variation = 56%), reflecting the presence of both large slabs and interstitial cobbles, the 40 cm sample is moderately sorted (34%) and the 80 cm sample is very poorly sorted (96%) as it includes both small pebbles and large blocks. Mean elongation (b/a) varies little amongst the three samples, but mean c/a and (a + b)/2c values show consistent change with depth, indicating an increase in "blockiness" and sphericity, and suggesting vertical shape sorting. Mean roundness values (1000 (2r / a)) vary little and are rather low, demonstrating angularity throughout the section. Mean dip increases from 10 cm to 40 cm, and confirms the impression that surface and near-surface slabs tend to lie almost flat, whereas those at depth are more disordered, often showing signs of upthrusting (dips of 70^{o} were recorded at 40 cm). Neither orientation (A_n 180) nor dip direction (A_n 360) show a statistically significant departure from circular uniformity; both appear to be entirely random.

Both of the other blockfields excavated revealed similar characteristics. Indeed, the uppermost 60 cm of the quartzite blockfield on Glas Mheall Liath closely resembles the equivalent part of the Carn Gorm blockfield in section (figure 8.18). Here large boulders with intermediate axes greater than 20 cm are found throughout the section, and the interstices between these are void near the surface, but filled with progressively smaller stones as depth increases. Again, the boulders are more tightly wedged together at depth. Exposed surfaces are much more rounded and weathered than those at even shallow depth and tend to be lichen-covered; erratics are absent and fines rare. On the felspathic gneiss of Mheall a'Chrasgaidh the "block" size is much smaller, with intermediate axes typically less than 10 cm in length. Again, however, the surface clasts are less angular and generally larger than those in the underlying layer. Depth to unweathered bedrock (50 cm) is less than in the previous examples, and the lowermost 20 cm consists of relatively large angular blocks, apparently in situ and surrounded by coarse sand. Erratics are again absent.

Various inferences can be drawn from the structural characteristics of these three sections.

(i) All three blockfields are apparently autochthonous, as indicated by the almost total absence of erratics. The angularity of subsurface clasts and the apparent lack of preferred stone orientations suggest that the block deposit has undergone little (if any) lateral displacement.

(ii) The block deposit is the product of the macrogelivation of physically and chemically sound though well-jointed bedrock. The absence of effective chemical weathering is indicated by the angularity of the blocks and the coarseness of such fine material as is present. The tetragonal regularity of many clasts and the presence of in situ joint-bound blocks at the base of the Mheall a'Chrasgaidh section is symptomatic of the operation of macrogelivation along joint planes.

(ii) The block deposit has undergone vertical frost sorting. Movement of individual clasts is evident in that only those at the base of the Mheall a'Chrasgaidh section reflect the dip of the original strata; all other clasts display apparently random orientation. Upward movement of larger clasts is indicated by the shape sorting on Càrn Gorm, in that steeply-dipping slabs are more susceptible to upfreezing than blocks (Chambers, 1967). The changes of particle size with depth in all three sections indicate repeated heaving and resettling of the deposit. In the course of such movement the framework of larger blocks would act as a sieve, with the result that the smaller stones that occupy the voids between boulders diminish in size downwards. Conclusive evidence of vertical sorting on the Mheall a'Chrasgaidh blockfield is provided by the presence of relict sorted polygons up to 0.9 m in diameter near to the site of the section (section 10.2; figure 10.6).

(iv) The blockfields are effectively inactive at present. This is implied by the lack of erected stones and lichen-free clasts at their surfaces, and by the rounding of exposed

clasts by microgelivation. The last-mentioned is particularly significant, in that the contrast between subrounded and subangular clasts on the blockfield surface and angular clasts below the surface suggests that no subsurface clasts have been subject to ~~microgelivation~~ granular disintegration. Since the rocks involved are apparently resistant to granular disintegration by freezing and thawing (chapter 6), the contrast in roundness between surface and subsurface clasts implies a long period of stability.

All of the blockfield areas studied were apparently glaciated during the Late Devensian glaciation (Chapter 4), and it seems very unlikely that such incohesive material could survive glaciation, despite the assertions of Ives (1966), Sugden (1970b, 1971) and others. Yet these features are apparently inactive at present, and have been so over a period of sufficient length to allow rounding by microgelivation of rocks that are not susceptible to granular disintegration. In this context it is noteworthy that the writer knows of no locality in Scotland where autochthonous blockfields have formed within the limits of the Loch Lomond Advance glacier. Indeed, in Rhum the areas occupied by these glaciers and the areas covered by blockfields are complementary and mutually exclusive (Ballantyne and Wain-Hobson, 1980). It is difficult to avoid the conclusion that the blockfields in the study area (and probably all the autochthonous blockfields in Scotland) were produced by large-scale frost-wedging and concomitant vertical frost sorting acting on mechanically sound though well-jointed bedrock during the Lateglacial period.

8.3.4 Characteristics of the Glas Mheall Liath blockslope

The blockslope east of Glas Mheall Liath, An Teallach (NH 076854; map 2) was selected for detailed study as it is one of the longest, steepest and best-developed features of its type and is representative of the quartzite blockslopes that cap many Scottish mountains. This blockslope descends from 950 m to 670 m, faces east and although largely

vegetation-free supports "islands" of Calluna, Rhacomitrium and various mosses.

The surveyed profiles of this blockslope (figure 8.19), though steep, are less so that the upper portions of talus slopes. Gradients in excess of 35° occur only where the blockslope crosses the Torridonian-Cambrian unconformity. Above this line there is a fairly straight slope with a modal gradient of 31.5° and a range of 25°-35°. Upslope of the surveyed profiles the gradient lessens. The shape of these profiles may be regarded as essentially that inherited from glaciation, yet no trace of glacial activity remains on this slope: no rock outcrops interrupt the cover of quartzite boulders and no erratics were found here, although both quartzite outcrops and erratics of "thrust" Torridon Sandstone are fairly common on the neighbouring blockslope on Sàil Liath.

The mean size of surface blocks apparently bears no relation to position on the slope (figure 8.20). On the Sàil Liath blockslope, however, the largest boulders often occur immediately downslope of quartzite outcrops, as reported by Crampton and Carruthers (1914) for blockslopes in Caithness. All of the six fabrics measured from surficial clasts (figure 8.21) display preferred downslope inclinations significant at the .05 level (A_n 360 statistic) and a degree of preferred downslope orientation (significant at the .05 level (A_n 180 statistic) for four samples) that is indicative of mass movement. In section (figure 8.9) the blockfield structure appears similar to that of the Glas Tholl talus cone (figures 8.7 and 8.9), but there are important differences. Large blocks do not occur at all depths, and there is a "true" diminution in clast size from the surface downwards. Bedrock is encountered as little as 0.4 m below the surface in places (and on the Sàil Liath blockslope crops out above the blockslope surface), which suggests that the block deposit is much shallower than the talus. Lastly, the voids between clasts in the lower parts of the sections are infilled with fines, comprising mainly medium and coarse sand.

The similarity between the structural and sedimentological characteristics of this blockslope and those of the blockfields

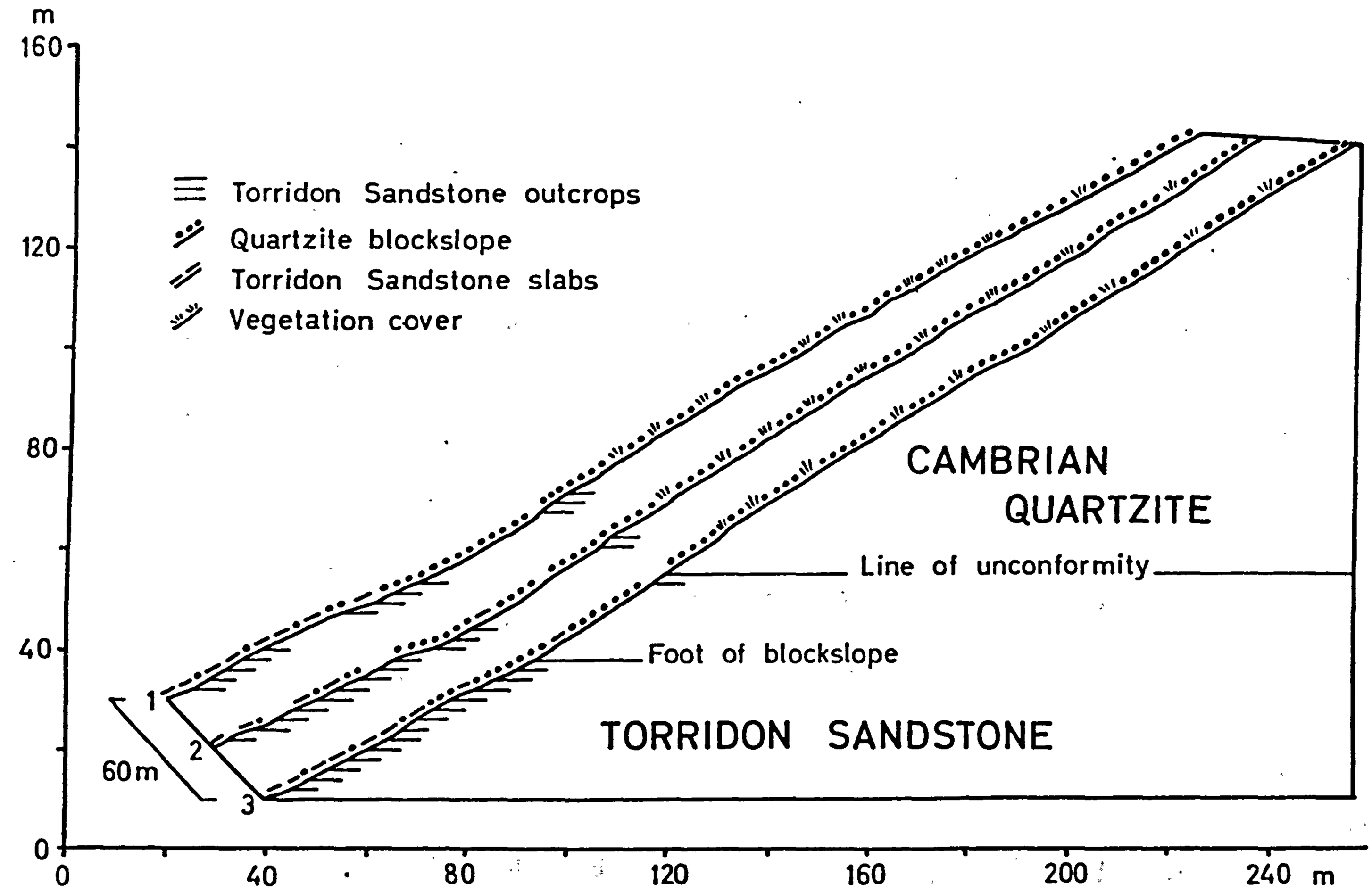


Figure 8.19: Surveyed profiles on the Glas Mheall Liath blockslope, showing the extension of the blockslope detritus downslope of the Torridonian-Cambrian unconformity. Although plotted as a block diagram, the individual profiles are undistorted

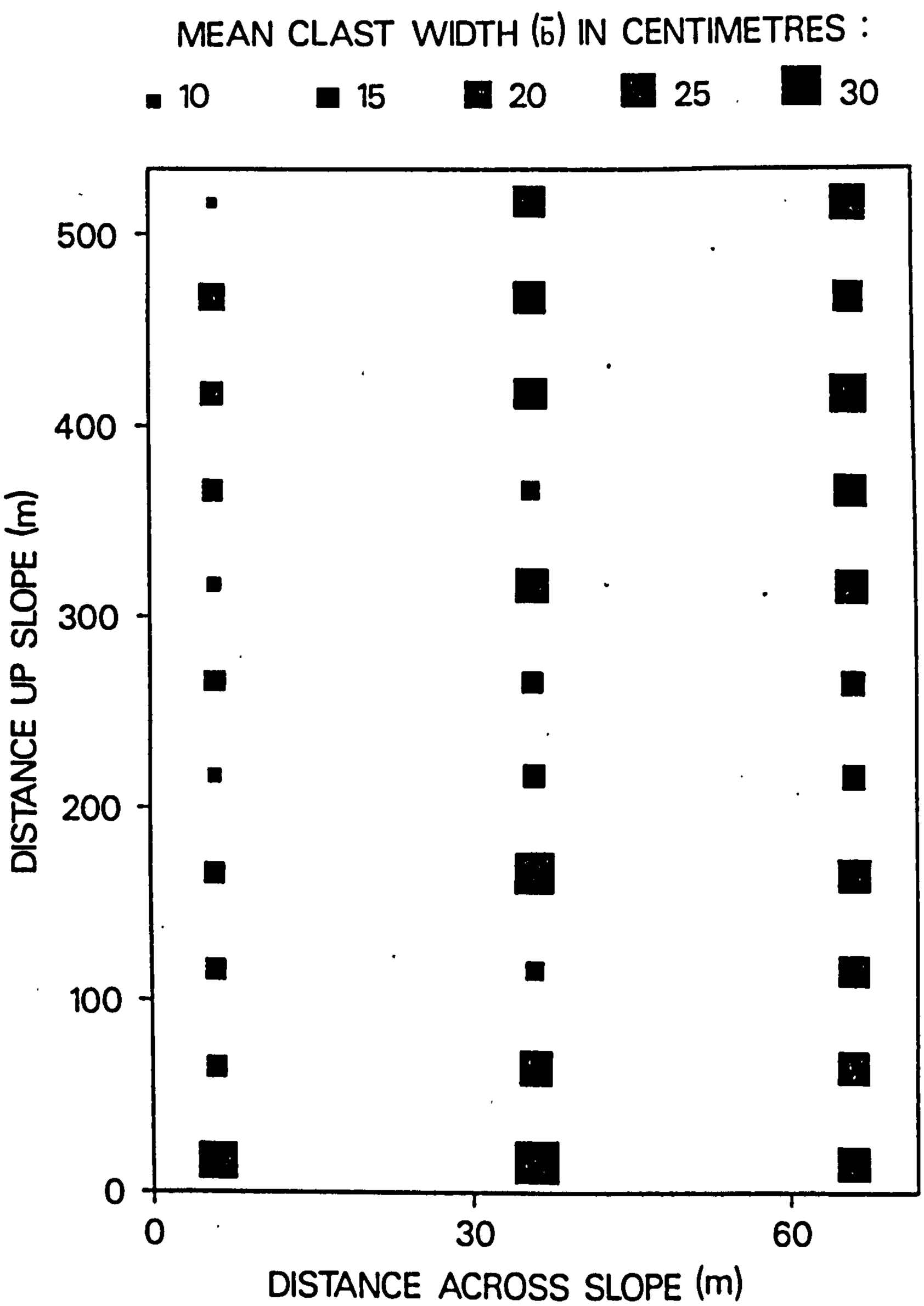


Figure 8.20: Mean clast widths (intermediate axis lengths) of 33 samples of 50+ stones on the Glas Mheall Liath blockslope, An Teallach.

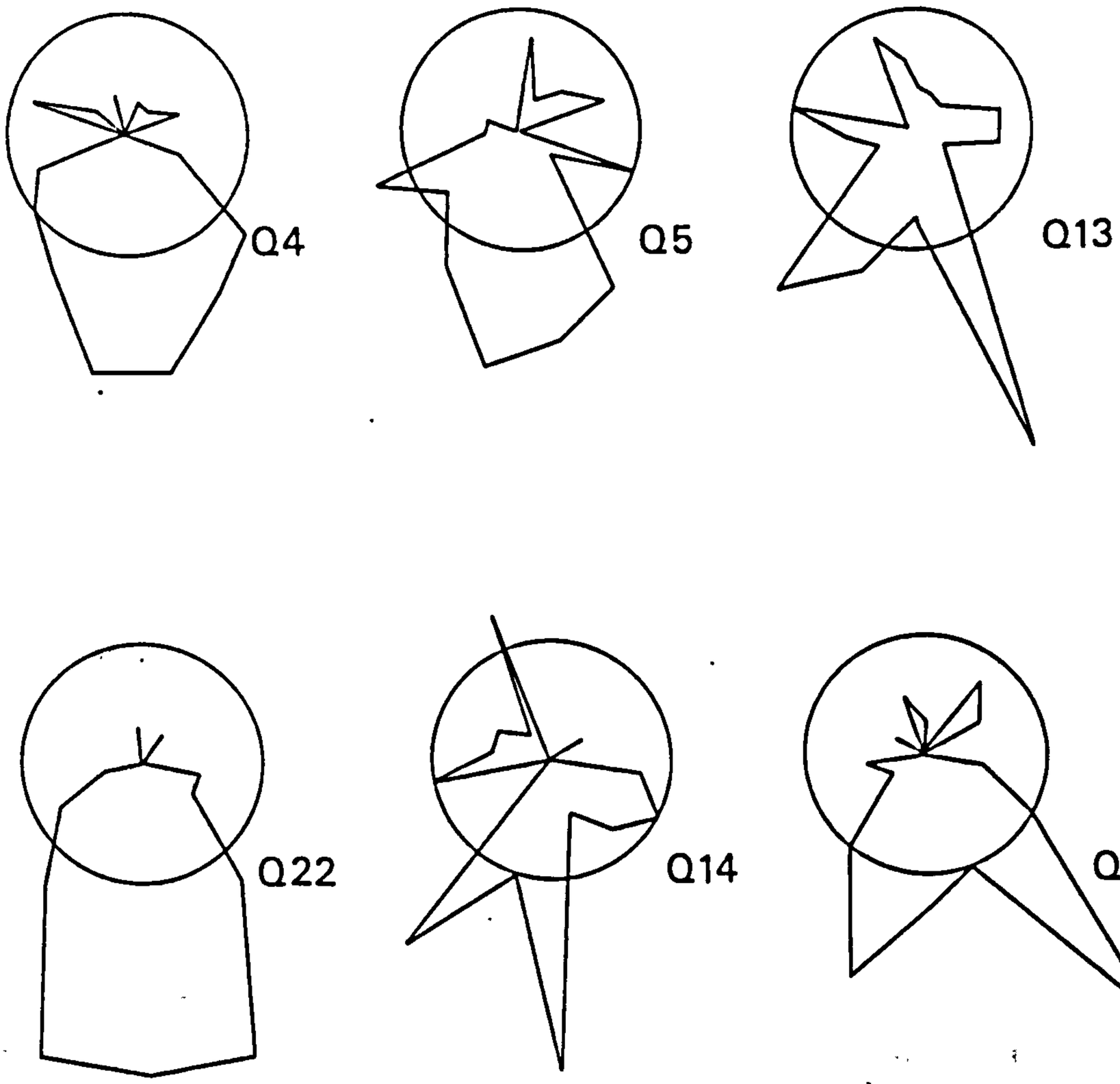


Figure 8.21: Surficial fabrics measured on the Glas Mheall Liath blockslope, An Teallach. The plotted positions of the fabrics are arbitrary; the line of maximum gradient is from the top to the foot of the page. Each circle represents four observations. Sample size = 50 clasts in each case.

described earlier (compare figures 8.9 and 8.18) indicates a similar genesis, namely macrogelivation of joint-bound blocks (these are typically oblate and flat-sided) and vertical frost sorting (giving a downward diminution in clast size for some depth below the surface). As with the blockfields, the blockslope appears to have been largely stable for a long period of time: exposed surfaces are stained, lichen-covered and rounded by microgelivation, whereas buried clasts are clean and angular. Like blockfields, blockslopes are, in the author's experience, absent from areas that lay inside the limits of the Loch Lomond Advance glaciers, which again suggests formation during the Lateglacial period.

There is, however, one vital difference in the characteristics of the two types of feature, in that blockslopes display strong evidence for former downslope movement. The fabrics measured on the Càrn Gorm blockfield are isotropic, but those on the Glas Mheall Liath blockslope display clear downslope trends. Level areas of the Sron an t-Saighdeir blockfield (map 8) and the Mheall a'Chrasgaidh blockfield (map 4) contain inactive sorted polygons, yet nearby low-angle blockslopes support fossil sorted stripes of similar width that are indicative of downslope movement during sorting. Large-scale mass-movement features are also found on blockslopes. There are poorly-developed boulder lobes on the Glas Mheall Liath blockslope, and well-developed boulder lobes and sheets on the blockslope-covered flanks of Sron an t-Saighdeir (map 8; figure 9.5). Boulder lobes and sheets are also well-developed on the quartzite blockslopes of Arkle (Sutherland), and Crampton and Carruthers (1914) reported "step-like terraces" on low-angle blockslopes in Caithness. Similar features have been interpreted as symptomatic of blockslope movement elsewhere (Rudberg, 1962, 1964; Dahl, 1966; Caine, 1968; Clark, 1972; Clapperton, 1975). Conclusive evidence of former movement of the Glas Mheall Liath blockslope is provided by the fact that the quartzite blockslope extends c. 40 m downslope over Torridon Sandstone bedrock, completely burying the latter.

Four mechanisms have been proposed to account for the downslope movement of blockslopes.

(i) Solifluction of the block deposit at a time when the blocks were encased in a matrix of fine material, the fines in the surface layers being subsequently washed out (Andersson, 1906; Lozinski, 1911; Klatka, 1961, 1962; Caine, 1968; Clapperton, 1975).

(ii) Downslope flow in the form of "rock glacier creep" as a result of plastic deformation of ice filling the interstices between blocks (Brockie, 1965; Caine, 1968).

(iii) Transport of the block deposit through solifluction of underlying fine material (Crampton and Carruthers, 1914; Benedict, 1970).

(iv) Movement through some form of creep (Richmond, 1962; Dahl, 1966), possibly as a result of frost heave of underlying fines (Rudberg, 1964).

In the context of the Glas Mheall Liath blockslope (i) seems unlikely, as it is difficult to envisage a source for the considerable quantity of fines that would be required to render the block deposit susceptible to solifluction. The lack of erratics suggests that till was not deposited on this slope and hence did not contribute to the supply of fines, and the rounding of only exposed surfaces indicates that significant microgelivation has taken place only since the blockslope stabilized. Moreover, mechanism (i) requires subsequent eluviation of large amounts of fine material that would presumably have been deposited as a colluvial sheet at the foot of the slope, and no such accumulation exists. Significantly, on level blockfields (where washing out of fines must have been minimal) considerable thicknesses of openwork blockfield are found (figure 8.18).

Explanation (ii) may be dismissed on theoretical grounds. As Whalley (1976) pointed out, glacier ice only flows under self weight when a critical thickness (about 20 m) is reached,

and the yield strength of a rock-ice mixture is much greater than that of ice alone, hence an even greater thickness would be required for flow of such a mixture to take place. The Glas Mheall Liath blockslope is however only 0.4 m thick in places, and lacks thermokarst features (kettle holes) typical of the decay of rock glaciers (Caine, 1968). Moreover, interstitial ice in such a shallow deposit must have been largely subject to annual melting, and it seems unlikely that perennial ice more than a few decimetres thick could have survived within the block deposit.

Mechanism (iii) is less easy to reject, as a matrix of fines appears to underlie all but the shallowest part of the blockslope at depths of 0.4-0.8 m. However, the fines in this layer tend to fill the voids between tightly-wedged blocks, and it is difficult to envisage movement en masse of the basal layers, even though the fairly high silt content of the matrix (table 8.4) indicates that the fines may be frost-susceptible and therefore conducive to ice segregation, which appears to be a necessary prerequisite for solifluction (see section 9.2.7). Also, as Caine (1968) pointed out, transport of the block deposit on a mobile bed of fines is unlikely to have resulted in preferred fabrics amongst surface blocks (figure 8.21).

The structure of the blockslope does, however, provide support for movement by creep (iv). In the previous section (8.3.3) it was argued that the downward diminution of clast size in level blockfields can be explained by repeated heaving and resettling of the entire deposit, during which small clasts would tend to fall down between larger ones. This process acting on a slope would give rise to large-scale frost creep, defined by Washburn (1967, p. 10) as

> " ... the rachetlike downslope movement of particles as the result of frost heaving of the ground and subsequent settling upon thawing, the heaving being predominantly normal to the slope and the settling more nearly vertical".

However, frost creep is generally associated with soils rather than block deposits. It may be, as Rudberg (1964)

Table 8.4

Grain-size analysis of fine material at the base of the Glas Mheall Liath blockslope

Size fraction	Limits (μm)	Sample 1 (% weight)	Sample 2 (% weight)
Coarse sand	600-2000	36.2	46.5
Medium sand	200-600	29.3	18.7
Fine sand	60-200	17.4	15.7
Silt	2-60	13.4	17.0
Clay	< 2	3.7	2.1

Analysis carried out by sieving to 75 μm and pippette analysis of the residue.

suggested, that blockfield creep and associated vertical sorting are generated by heave of underlying frost-susceptible fines. It is not certain, however, that this would create a preferred trend in surficial fabrics, or even if fines existed in the basal horizon of the blockslope at the time of movement. On the other hand, heaving of large blocks has been reported from arctic areas (Högbom, 1910, 1914; Yardley, 1951; Washburn, 1969a) and must have occurred during blockfield formation to have produced the block deposit in the first place. Significantly, Dahl (1966, p. 62) reported "frost in the ground at shallow depth" in blockfields in northern Scandinavia, and the "consequent high level of groundwater", which suggests that the "frost" in fact consisted of interstitial ice.

The presence of large-scale sorted polygons on the Sron an t-Saighdeir blockfield (Ballantyne and Wain-Hobson, 1980; map 8; figure 10.9) and evidence from elsewhere (chapter 3) strongly suggest that blockfields and blockslopes of Lateglacial age formed under permafrost conditions, and it is tempting to relate their thickness to the depth of the former active layer (below which macrogelivation would presumably have been ineffective) (Stromquist, 1973). If the Glas Mheall Liath blockslope were underlain by permafrost, then it is probable that meltwater and precipitation percolating into the blockslope would have frozen from the bottom up during winter freezeback, causing the entire block deposit to heave normal to the slope. Thaw from the surface down would cause the gradual resettlement of clasts slightly downslope of their original positions, and the gradual migration of smaller clasts towards the base of the deposit. A preferred orientation of surface material would be expected to form as a result of clasts adopting an alignment of least resistance.

Although most of the blockslope appears to be inactive, there is evidence for recent activity in limited areas of the slope where relatively small lichen-free clasts are abundant at the surface. It is possible that small-scale creep is taking place, but more likely that these stones

have been displaced by descending mountaineers as the areas involved are suitable for "scree running".

The descent of the block deposit to a line c. 40 m downslope of the Torridonian-Cambrian unconformity is impressive evidence for rapid rates of movement during the Lateglacial period. If it is assumed that such movement was restricted to the Loch Lomond Stadial and that the latter lasted 500-1000 years (chapter 2), then the mean rate of creep must have been between 4 cm y^{-1} and 8 cm y^{-1}. These rates are more than an order of magnitude greater than those measured by Rudberg (1964) on low-angle blockslopes in Scandinavia, but compare with frost creep rates of 0.7 cm y^{-1} to 2.2 cm y^{-1} on gentler (10-14°) slopes in Greenland (Washburn, 1967).

8.4 Debris slopes and surfaces

8.4.1 Previous research

8.4.1.1 General

Although there is an extensive literature on the features described earlier in this chapter, that on slopes and plateaux mantled by a mixture of fine and coarse detritus is practically non-existent, probably because such features have failed to excite the attention of researchers as much as those that are distinctly "periglacial", such as blockfields, blockslopes and cryoplanation terraces. Yet, as indicated in chapter 6, rocks susceptible to microgelivation are often covered by a regolith of frost-shattered clasts embedded in the products of granular disintegration. Within the study area (and indeed throughout upland Scotland) this type of periglacial deposit is much more common than, for example, blockfields and blockslopes, which are restricted in distribution to lithologies of relatively limited provenance. Moreover, the writer has observed similar deposits covering vast areas of the Barrow Surface on Cornwallis Island in the central Canadian Arctic, and on slopes on Ellesmere Island, the Jotunheimen Massif,

Norway, and the Massif Central, France. Ironically, it is on this type of deposit rather than on the more spectacular types described earlier that many of the most "typical" periglacial features are found, such as solifluction landforms and patterned ground (chapters 9 and 10).

The importance of this type of deposit was recognised by Lundqvist (1962), who coined the term "nonsorted field" to describe "unpatterned frost ground". This term, he wrote,

> " ... covers those widespread areas where frost action in the ground occurs, but no structural pattern is developed. The frost action asserts itself merely in an incomplete concentration of stones towards the surface and in a slight solifluction" (p. 64).

Where the concentration of stones at the surface is more complete, stone pavements ("dallages") may be formed (Jahn, 1961; Bout and Godard, 1973). These differ from blockfields in that stones concentrated on the surface are embedded in underlying fines.

Similar deposits have also been described on slopes (Jahn, 1975; McArthur, 1975; Sukhodrovskiy, 1975; Slaymaker and McPherson, 1977). The most complete accounts of such slopes are those given by French (1976) and Church et al. (1979). The former described "smooth debris-mantled slopes"

> " ... characterised by a relatively smooth profile with no abrupt breaks of slope, and with a continuous or near-continuous veneer of frost-shattered or solifluction debris. There is no widely developed free face or bedrock outcrop, and maximum slope angles are highly variable ..." (p. 152).

The features described here as debris surfaces and debris slopes possess characteristics similar to those described by Lundqvist and French.

8.4.1.2 Previous research on debris surfaces and debris slopes in upland Britain

Although some authors (e.g. Romans et al., 1966) have considered that the stony regolith that is widespread on plateaux and mountainsides in upland Britain is derived from

till, the majority of investigators have attributed its formation to periglacial weathering, frost sorting and mass movement (Tivy, 1962; Ragg and Bibby, 1966; Watson, 1969). On Broad Law, in the Southern Uplands, Ragg and Bibby (1966) carried out a detailed analysis of five sections in pits excavated at altitudinal intervals of 60-100 m. All profiles above 600 m revealed a rubble layer of small angular fragments of greywacke overlying a fine sand layer in which angular stones increased in size and frequency with depth. Ragg and Bibby interpreted this deposit as the product of frost weathering associated with an annual freeze-thaw cycle under permafrost conditions, and attributed formation of the coarse rubble layer to eluviation of fines, operation of the latter process being suggested by a coating of sand on the upper surfaces of stones. The lack of clay in these sections indicated mechanical breakdown of parent material (see also Tivy, 1962). Vertical and lateral sorting of such deposits is indicated by the formation of stone pavements (Kelletat, 1970a) and sorted polygons and circles (e.g. Simpson, 1932; Godard, 1965; Ryder, 1968; Ryder and McCann, 1971) on level ground and sorted stripes on slopes (e.g. Miller et al., 1954; Caine, 1962, 1963a, b; Kelletat, 1970a; Ryder and McCann, 1971). Downslope movement is evident from the preferred orientation of stones in the upper parts of such deposits (Ragg and Bibby, 1966; Ball and Goodier, 1970) and by the widespread formation of landforms associated with slow mass movement, such as lobes, terraces and sheets (Peach et al., 1913a; Hollingworth, 1934; Galloway, 1958, 1961a; Tivy, 1962; Ragg and Bibby, 1966; King, 1968, 1971b, 1972; Mottershead and White, 1969; Kelletat, 1970a, b; Sugden, 1971; Ballantyne, 1977; McMillan, 1978). The depth of these deposits is variable. Romans et al. (1966) described maximum depths of 0.5 m on the plateaux areas that they examined and depths of 0.6 m (Caine, 1963a), 0.6-1.0 m (Tivy, 1962) and 1.0 m (Galloway, 1958) have been reported on slopes up to 20° in the Lake District, Lowther Hills and Southern Uplands respectively.

The sections examined by Ragg and Bibby (1966), however, ranged from 1.5 to 3.5 m in depth, and they and others (Galloway, 1958; Tivy, 1962; Watson, 1969) have observed that the debris mantle tends to thicken downslope.

The thickness of the deposits on Broad Law and elsewhere prompted Ragg and Bibby (1966) and Romans et al. (1966) to consider that areas where these are developed must have escaped glaciation by the Late Devensian ice sheet, but this interpretation is inconsistent with evidence relating to the extent and thickness of this ice sheet (Sissons, 1974a). However, Romans et al. (1966) and Romans and Robertson (1974) have concluded from the presence of silt droplets that these deposits were formed under permafrost conditions, and hence were produced when the climate was much colder than it is at present. Since all the areas studied by these authors were ice-free during the Loch Lomond Advance, it is reasonable to infer that the frost debris deposits to which they refer were formed during the Lateglacial period.

There is, however, abundant evidence that unvegetated debris surfaces and slopes in Scotland are undergoing various forms of geomorphic activity at present. Miller et al. (1954) destroyed the sorted stripes on a debris slope at 600 m on Tinto Hill, Lanarkshire, and found that they re-formed over two winters. The present author repeated this experiment at the same site, and found perfect re-formation of stripes on ground that had been dug over to a depth of 30 cm after a period of only six months (November 1977 to May 1978), which indicated that under favourable conditions (discussed in chapter 10) frost sorting may result in patterned ground formation under present conditions over short time periods. A similar conclusion was reached for miniature patterned ground on debris surfaces in Wales by Tallis and Kershaw (1959). Hollingworth (1934) and Hay (1937) both considered that active downslope movement of detritus was taking place on striped debris slopes, and Caine (1962, 1963a, b) has demonstrated that the uppermost 10 cm of striped debris slopes on Grassmoor in the Lake

District is subject to movement under present conditions. The rates of movement he obtained for surface clasts were remarkable, ranging from 12-18 cm y^{-1} on slopes of 15^{o} and 45-50 cm y^{-1} on slopes of 25^{o}. Similar rates have been measured by the present author on striped ground on Tinto Hill (section 10.2).

Although direct measurements of movement on vegetation-covered debris slopes are lacking, there is indirect evidence that suggests that movement is taking place. Solifluction lobes developed on such slopes have been considered to be undergoing downslope movement (Galloway, 1958, 1961a; King, 1968, 1972; Kelletat, 1970a), and the movement of ploughing blocks on such slopes has been demonstrated by Tufnell (1972, 1976) in the northern Pennines and Shaw (1977) on Lochnagar. Radiocarbon dates obtained from material over-ridden by solifluction lobes by O'Brien (cited in Sugden, 1971) in the Cairngorms and by Mottershead (1978) on Ben Arkle, Sutherland, indicates that solifluction has been active on vegetated debris slopes in these areas since the mid-Flandrian.

Wind has played an important role in modifying debris surfaces and slopes by stripping the vegetation cover and removing fine material. King (1968, 1971b) described widespread wind stripes and denuded surfaces on the Cairngorm plateau, and similar features have been described elsewhere in Scotland by Kelletat (1970a), Ball and Goodier (1974) and Goodier and Ball (1975). In some areas wind erosion has resulted in the formation of relatively small-scale deflation features (those described by King are generally about 1 m wide and 2-4 m long); elsewhere deflation has produced a lag surface with only patchy vegetation cover.

8.4.2 Debris surfaces and debris slopes in the Northern Highlands

On ground over 600 m in the study areas the distribution of debris surfaces and debris slopes is largely complementary to that of blockfields, as the area occupied by other types of terrain (bare rock, talus, glacial deposits and aeolian deposits) is proportionately small. In general, debris

surfaces and slopes correspond in distribution with rocks that have proved relatively susceptible to ~~microgelivation,~~ granular disaggregation, including Torridon Sandstone, basal Cambrian Quartzite and most Moinian rocks, particularly micaceous schists (chapter 6), in areas where the slope is insufficiently steep for rockfall to take place. On such terrain there is also a correspondence between vegetation cover and rock type. The debris surfaces and slopes developed on the Moine Schists of Ben Wyvis are almost totally covered by a Rhacomitrium heath (Pearsall, 1968; figure 8.22) and on similar rocks in the Fannichs vegetation cover is extensive, though less complete. The debris slopes and surfaces on the Torridon Sandstone of An Teallach, however, are predominantly vegetation-free (figure 8.23). The difference may be partly attributable to the increased exposure of the latter area, although the poverty of clay and humus in the felspathic regolith of Torridon Sandstone areas probably contributes to the lack of vegetation (chapters 4 and 11).

8.4.3 Characteristics of debris surfaces

The structural and sedimentological characteristics of the regolith that mantles plateaux in the study areas were investigated through cutting sections and analysing the characteristics of constituent fine and coarse material. On the northern plateau of An Teallach the depth of regolith between two outcrops of Torridon Sandstone was measured by digging three pits and augering between these (figure 8.24). The maximum depth measured was c. 0.8 m; typically, depth to bedrock under the debris mantle on An Teallach and elsewhere (figure 8.18) appears to be 0.3-0.6 m.

Sections 1-5 on An Teallach (figure 8.24) exhibit similar structural characteristics. At the surface there is a concentration of slabs, the exposed surfaces of which are well-rounded by microgelivation. In places this concentration is so complete that stone pavements have formed (figure 8.25). Immediately below the surface there is a sand-rich zone with fewer and generally smaller stones. This



Figure 8.22: The vegetation-covered debris surface of Tom a' Choinnich (955 m), Ben Wyvis, with nonsorted relief stripes and solifluction features on the surrounding slopes.



Figure 8.23: The deflated debris surface of the northern plateau of An Teallach, with horizontal turf-banked terraces developed on east-facing slopes above sheets of aeolian sand deposits.

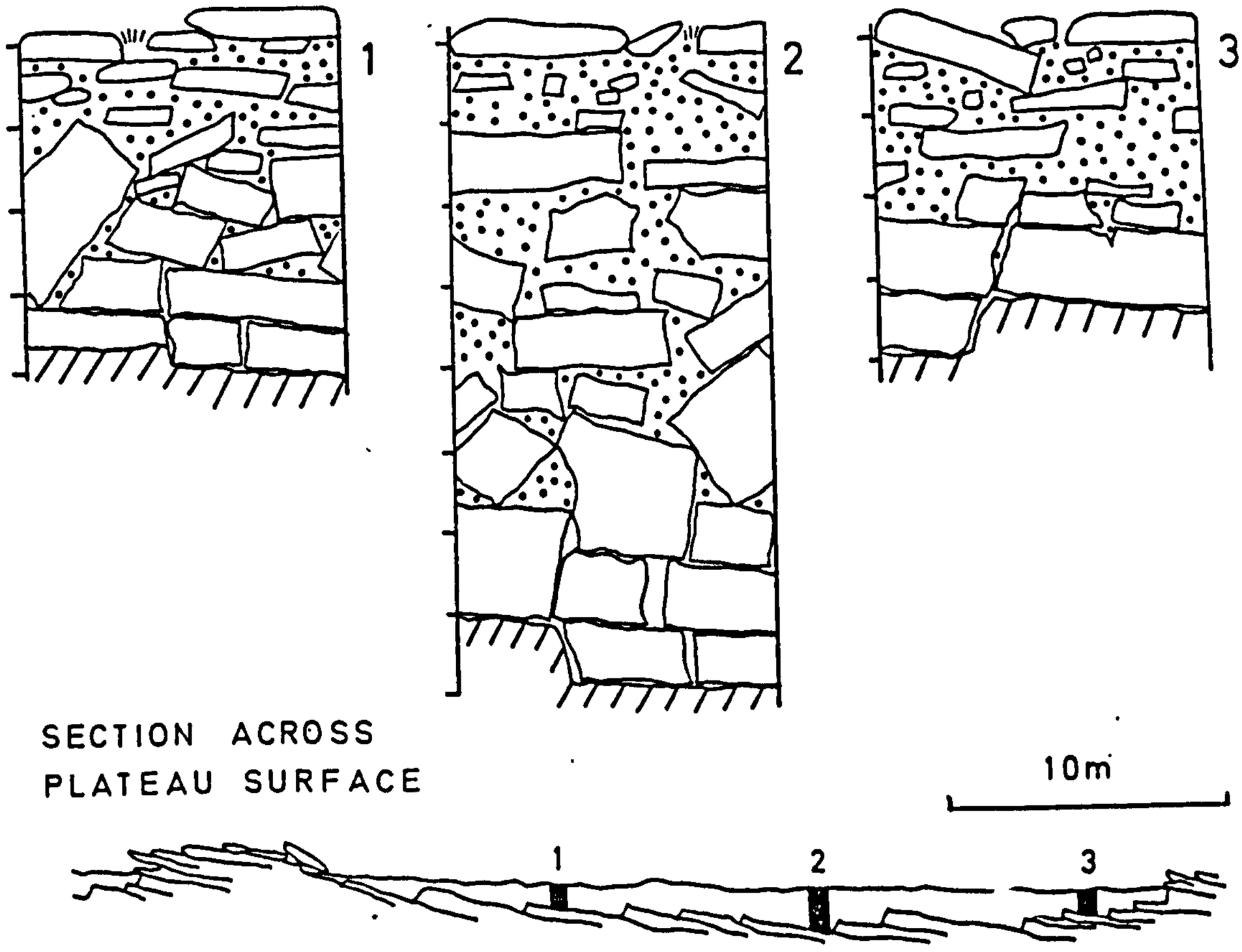


Figure 8.24: Sections trenched in the debris surface of Torridon Sandstone regolith that covers the northern plateau of An Teallach.



Figure 8.25: Stone pavement amid tor-like rock outcrops at 800 m altitude on the northern plateau of An Teallach.



Figure 8.26: Deflation surface at 780 m altitude on the northern plateau of An Teallach, showing the residual gravel lag typical of such surfaces. The belt of quartzite clasts to the left of the figure was probably deposited during the downwastage of the Late Devensian ice sheet.

grades downwards to a concentration of angular blocks that increase in size with depth. In the bottom 10-20 cm of each section these blocks are apparently *in situ* above unweathered bedrock. Section 6 on figure 8.24 differs markedly from the above pattern, in that it comprises rounded sandstone boulders and cobbles distributed apparently at random in a coarse sandy matrix. The high concentration of Cambrian Quartzite erratics at this site points to a glacial rather than periglacial origin for this deposit, which underlies the belt of quartzite boulders (figure 8.26) described in chapter 4. Section 7 also contains quartzite erratics, but is otherwise structurally similar to sections 1-5. It is suggested that at this site quartzite boulders (but not till) were left on the plateau surface by the downwasting Devensian ice sheet, and that these became incorporated in the regolith as it formed.

The characteristics of samples of stones taken from pits on the An Teallach plateau (table 8.5) bear out the above observations. In all cases there is a decline in mean diameter from the surface to the immediate subsurface, together with an increase in "blockiness" and an abrupt decrease in roundness. In chapter 6 clasts taken from the base of pits 1 and 3 were shown to be very different in shape from these at the surface (figure 6.8, samples B2b and B3b). In the same chapter it was argued that slabs of Torridon Sandstone are the products of macrogelivation of exposed rocks. If this is so, then the slabs embedded in the upper parts of these sections reflect the break-up of exposed rock outcrops, whereas the underlying blocks represent macrogelivation at depth. That surface slabs may (in some cases) have moved to their present positions from adjacent outcrops is suggested by an apparent preferred alignment of surface slabs at pits 1 and 3, where the surface slope is *c.* 3^{o}.

Particle size analysis of fines taken from depths of 10 cm and 60 cm in pit 2 (figure 8.27) shows that over 90% of the matrix consists of angular medium and coarse sand and granules (i.e. particles > 0.2 mm in diameter) and that clay

Table 8.5: Characteristics of clasts in plateau surface pits.

Site:		An Teallach Pit 1		An Teallach Pit 3		An Teallach Pit 5		Ben Wyvis Pit		
Depth (cm):		0	20	0	10	0	5	5	20	40
Sample size:		61	50	71	50	25	32	63	44	15
$\frac{(a+b+c)}{3}$ (cm)	Mean	12.2	8.6	10.5	6.3	13.1	11.3	5.9	8.4	8.4
	S.D.	12.3	5.6	6.2	2.6	6.1	3.9	3.0	3.7	3.1
(a + b) / 2c	Mean	3.8	2.5	3.8	2.3	5.4	3.8	5.2	4.8	5.3
(b / a)	Mean	0.66	0.73	0.66	0.71	0.78	0.76	0.65	0.66	0.64
(c / a)	Mean	0.27	0.38	0.28	0.41	0.19	0.24	0.22	0.30	0.32
1000 (2r / a)	Mean	169.9	47.7	145.1	56.1	N.M.	N.M.	50.1	46.5	31.7

N.M. - Not measured.

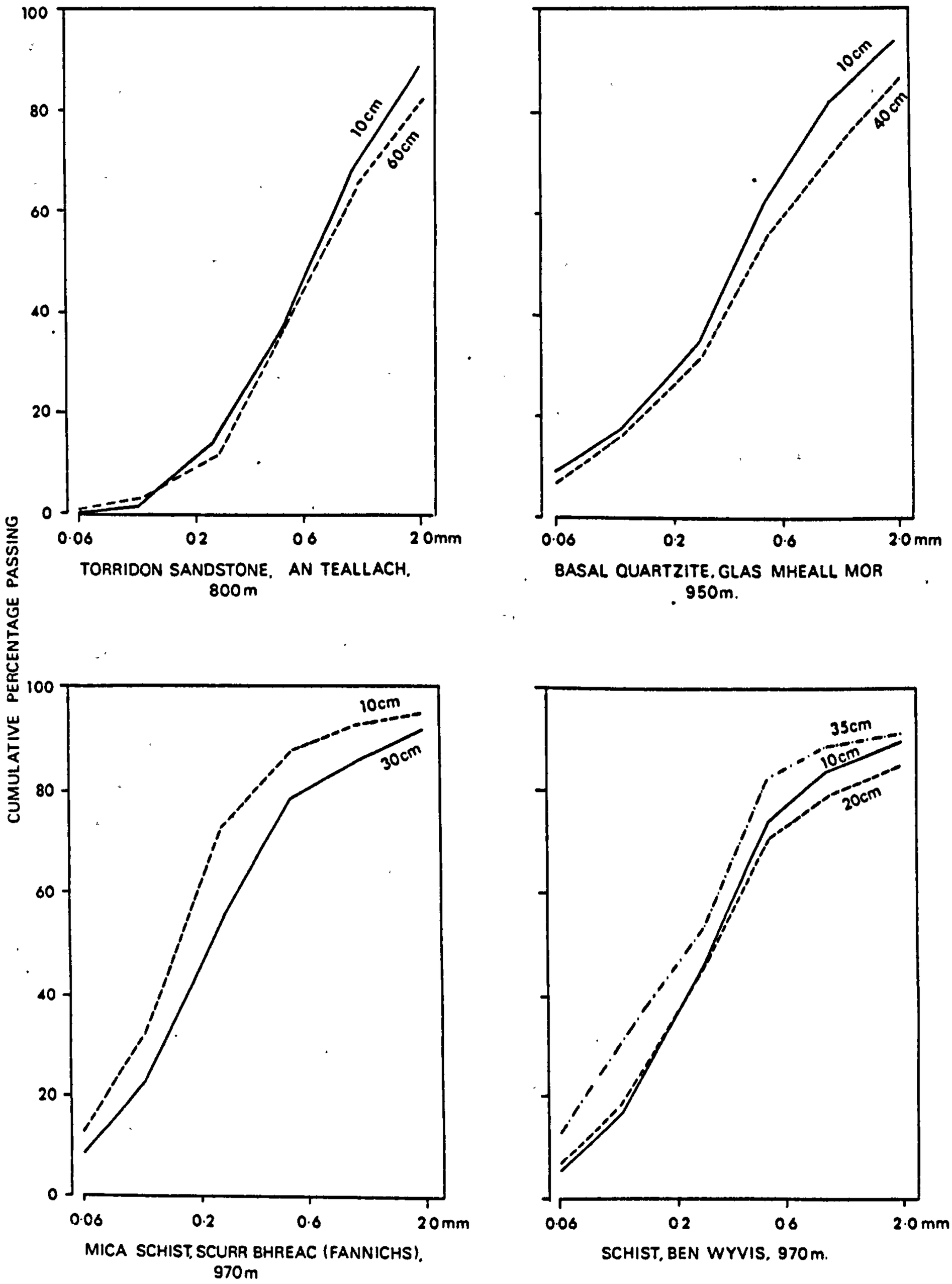


Figure 8.27: Grain-size distributions of matrix material collected from the depths shown in pits dug in debris surfaces developed on Torridon Sandstone, basal Cambrian Quartzite and Moine schist.

and silt (< 0.06 mm) are almost absent. The similarity of particle size distribution of the two samples suggests that the matrix is homogeneous. At the surface, however, sand is often absent and the spaces between slabs in many areas of the An Teallach plateau are covered by moderately-sorted fine gravel made up of both small sandstone clasts and pebbles weathered out of conglomeratic bands in fairly equal proportions (figure 8.28). This is interpreted as a lag deposit formed through the removal of finer particles by wind (chapter 11). The areas of plateau where this occurs have been classified and mapped as deflation surfaces; an example is illustrated on figure 8.26.

The above observations allow several inferences to be made concerning the origin of debris surfaces on An Teallach.

(i) The absence of erratics and glacially-rounded clasts in five of the seven sections indicated that much of the regolith is autochthonous. Limited areas, however, apparently consist of sandy till (pit 6), and in others erratics are embedded in an autochthonous deposit (pit 7).

(ii) The blocks at the base of autochthonous deposits are apparently the product of the macrogelivation of mechanically sound bedrock. The angularity of the blocks suggests that chemical weathering has not contributed to block formation. The regularity of many subsurface blocks and the fact that blocks lie <u>in situ</u> at the base of the regolith are symptomatic of the operation of macrogelivation along joint planes. The coarseness and angularity of the matrix material suggest that it is the product of granular disintegration of Torridon Sandstone, as does the abundance of pebbles weathered out of conglomeratic strata. The slabs on the surface of the deposit are apparently the product of the macrogelivation of initially exposed bedrock (chapter 6).

(iii) The debris mantle has undergone limited vertical sorting and localised surficial mass-movement. The former is indicated by concentration of slabs at the surface above a sand-rich zone, and by the presence of poorly-developed inactive sorted

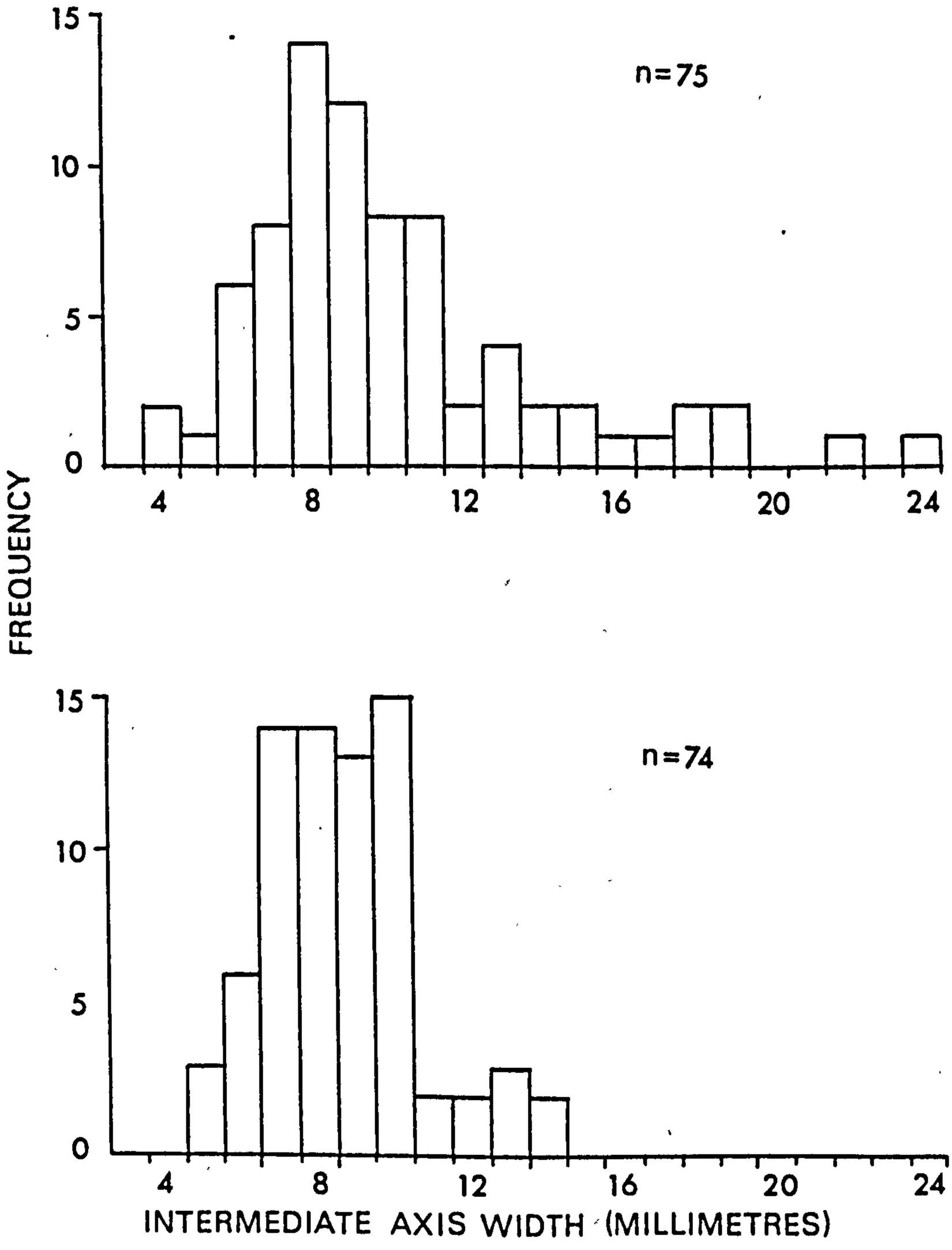


Figure 8.28: Particle-size distributions for a sample of fine gravel collected from a deflation surface on the northern plateau of An Teallach. The upper histogram shows the widths (intermediate axis lengths) of sandstone clasts; the lower shows the widths of pebbles weathered from Torridonian conglomerate. The proportions (75:74) are those of the original sample. Widths were measured using a micrometer.

polygons on parts of the plateau (chapter 10; figure 10.5). The latter is indicated by the preferred downslope orientation of surface slabs on gentle slopes.

(iv) Apart from erosion by wind (chapter 11) and wash (chapter 12) the debris surfaces are apparently inactive at present, and have been so for a long time. This is indicated by the rounding of exposed surfaces by ~~microgelivation~~ granular disintegration and the preservation of a gravel lag on deflation surfaces.

The inferences made above concerning the formation of autochthonous debris surfaces on Torridon Sandstone suggest a genesis similar to that envisaged for blockfields (section 8.3.3), except that (i) macrogelivation has affected a much shallower layer of bedrock, probably reflecting the shallow depth of open joints, (ii) ~~microgelivation~~ granular disintegration has been much more effective, producing a fine matrix at all depths, and (iii) surface clasts have apparently moved downslope on gentle gradients (although this may be true of blockfields also). As with the blockfield areas, the debris surfaces on the northern plateau of An Teallach were covered by the Late Devensian ice sheet, but escaped glaciation during the Loch Lomond Advance. It would therefore seem likely that these deposits are essentially of Lateglacial age, although subsequently modified by microgelivation, wash, deflation and near-surface frost creep (see below).

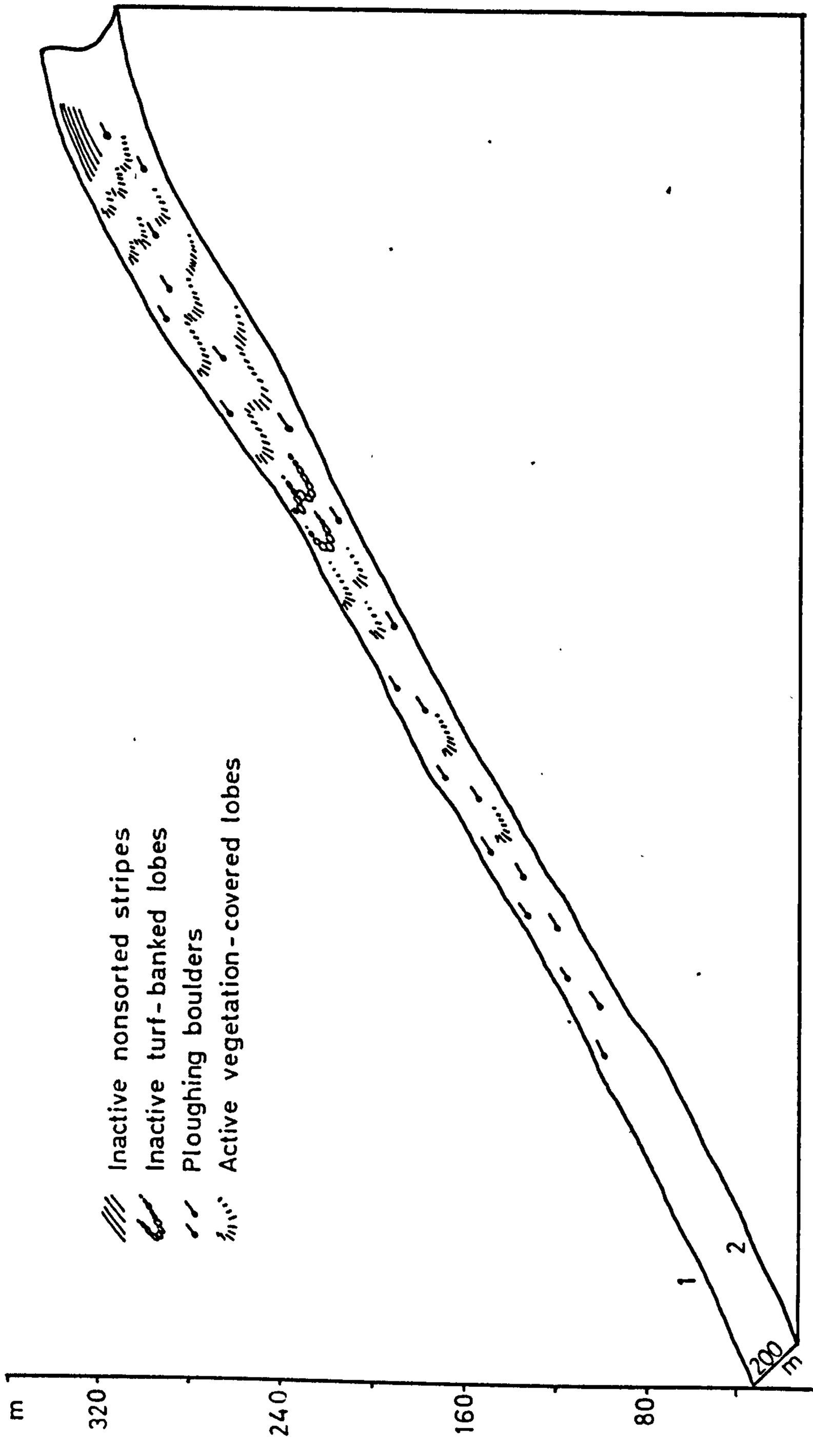
The debris mantle structures on other rock types (figure 8.18, sections 1, 2 and 6) indicate an evolution similar to that proposed for that on Torridon Sandstone, although evidence for vertical sorting is absent (table 8.5) and the shape and size characteristics of the detritus reflect different responses to periglacial weathering. In all cases buried clasts are predominantly angular and erratics are absent, suggesting that the deposits are authochthonous. At the base of each pit there is a transition from clasts in situ to unweathered bedrock. Near and at the surface clasts show evidence of modification by ~~microgelivation;~~ granular disintegration; on schistose rocks (pits 1 and 2, figure 8.18) these take the form of slightly rounded (table 8.5) platy slabs from which

smaller fragments have flaked along lines of schistosity, and on basal Cambrian Quartzite (pit 6, figure 8.18) surface debris is slabby and rounded. In all cases the deposit is apparently inactive at present, and soil horizons have developed on the vegetation-covered deposits, rankers on Scùrr Breac and a well-developed podzol on Ben Wyvis. The formation of the latter implies that the deposit has been stable for several hundreds (possibly thousands) of years, and points again to the formation of the regolith in the Lateglacial period.

The particle size distribution of the matrix material in these sections (figure 8.27) reveals little change with depth but considerable variation between sites. The clay-silt fraction is much greater at all three sites than on An Teallach, and increases with depth on schistose rocks, reflecting eluviation from the uppermost horizons. On schistose rocks between 70% and 90% of the matrix is finer than 500μm (mainly fine and medium sand), compared with 50% to 65% on basal quartzite and less than 40% on Torridon Sandstone. These differences in matrix constitution provide an explanation for the difference in vegetation cover on different regoliths, in that a more abundant fine fraction aids the retention of nutrients and moisture in the soil and increases cohesion, thereby minimising wind erosion. As will be seen in the next four chapters, these differences in the particle size of the matrix have also been of crucial importance in determining the nature of periglacial activity on different terrains, and hence the type and distribution of periglacial landforms.

8.4.4 Characteristics of some debris slopes

Profiles were surveyed on two of the steepest debris slopes in the study area, namely a vegetated debris slope on the north-west flank of Ben Wyvis (downslope of NH 453674; figure 8.29) and an unvegetated debris slope at the head of Coire a'Mhuilinn, An Teallach (figure 8.30). The former supports both fossil and active mass movement features (Galloway, 1958, 1961a; McMillan, 1978). As the structural characteristics of these landforms are crucial to the



Surveyed profiles on the Ben Wyvis debris slope showing the distribution of landforms developed on this slope. Although plotted as a block diagram, the profiles are undistorted.

Figure 8.29:

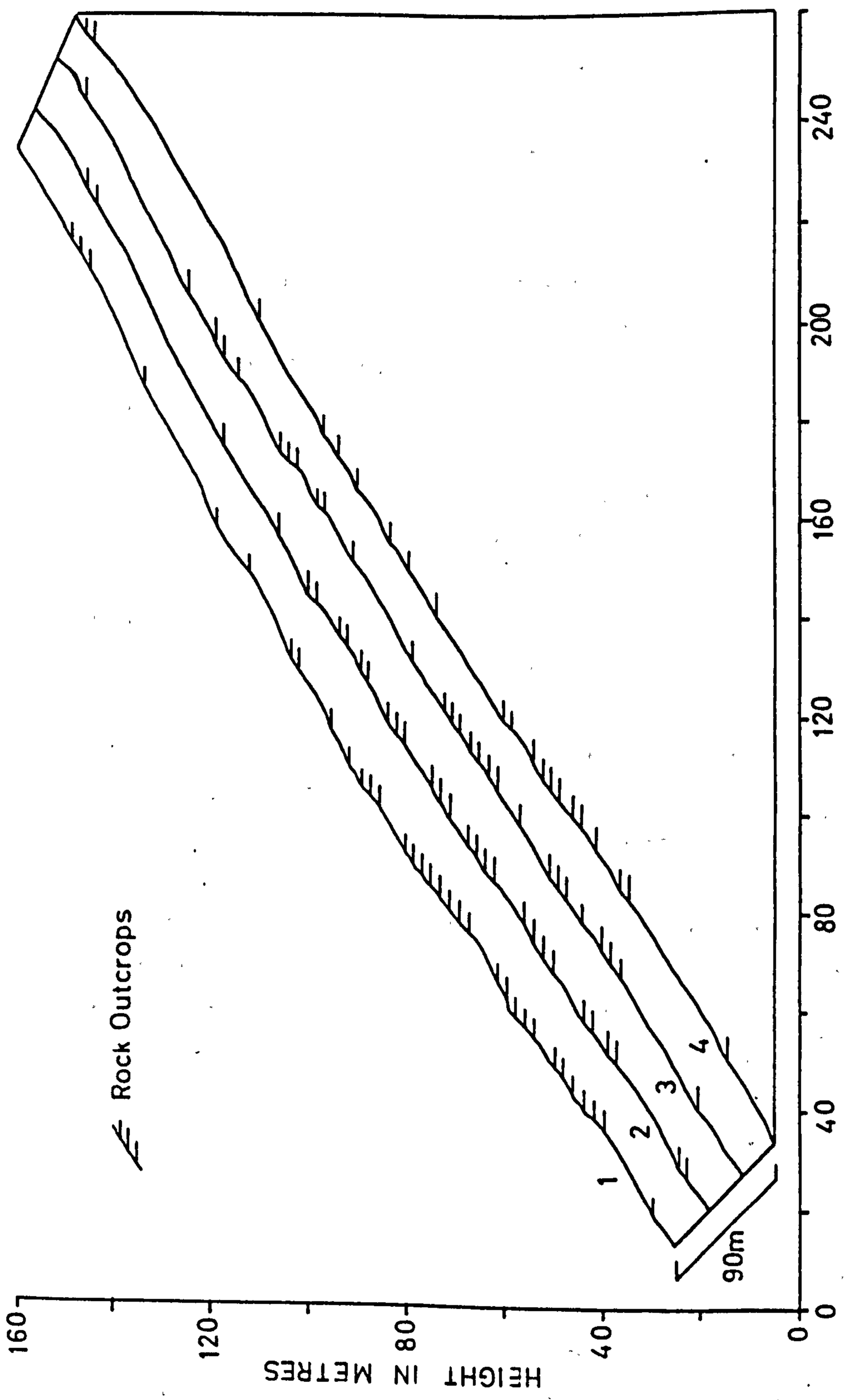


Figure 8.30: Surveyed profiles on the Coire a'Mhuilinn debris slope, showing outcrops of Torridon Sandstone bedrock. Although plotted as a block diagram, individual profiles are undistorted.

interpretation of the evolution of the deposits on this slope, this topic is dealt with in chapter 9. For the moment it is sufficient to note (i) that bedrock outcrops are rare and of limited extent, and that the slope ranges in angle from 22° to 36°, with a pronounced modal range of 28°-30°, (ii) that slow mass movement (solifluction) is currently operating over at least the upper half of this slope, except in areas occupied by fossil forms (chapter 9) and (iii) that the upper part of the slope deposit consists of frost-shattered clasts embedded in the products of microgelivation (figure 8.9), being similar in composition to the section on the Ben Wyvis plateau discussed in the previous section, and similar in structure to the deposits investigated by Ragg and Bibby (1966) on Broad Law. The main formative processes are considered to be frost weathering and vertical sorting during the Lateglacial period and solifluction of the weathered mass, the last mentioned being active at present.

The Coire a'Mhuilinn debris slope on An Teallach faces north and is about 600 m long, descending from the summit of Glas Mheall Mór (981 m) to the head of Coire a'Mhuilinn (c. 650 m). What little vegetation is found on this slope (mainly Calluna Vulgaris and Festuca spp.) generally grows on the risers of poorly-developed turf-banked terraces (chapter 9) or on windblown sand that clings to the lower parts of the slope. The contrast in vegetation cover between this debris slope and that on Ben Wyvis can be explained in terms of the removal by wind of the products of ~~microgelivation,~~ granular disintegration, leaving only a stony regolith.

The surveyed profiles on the Coire a'Mhuilinn debris slope are much less regular than those on other detritus slopes (compare figure 8.30 with figures 8.4, 8.10, 8.19 and 8.29). This results from the protrusion of rock outcrops through the shallow debris mantle. Yet the modal angle of slope is 31°, less than that of the Glas Mheall Liath blockslope where bedrock outcrops are absent. The difference highlights the relative resistance of Torridon Sandstone to macrogelivation, which has destroyed all outcrops on the

quartzite blockslope. The lower parts of the debris slope were also protected from the operation of macrogelivation during the Lateglacial Stadial by the Loch Lomond Advance glacier that occupied Coire a'Mhuilinn (Sissons, 1977a).

Mean clast size varies little over the surface of the debris slope (figure 8.31) and shows no downslope trend. The largest clasts are often found perched on rock outcrops with smaller material piled up behind them, suggesting that the stones reached their present position through downslope movement. This is supported by the strong preferred downslope orientation displayed by twelve fabrics that were measured on surficial clasts on this slope (figure 8.32). All fabrics except D1 and D2 show preferred downslope orientations significant at the 0.025 level (A_n180 statistic) and reversed imbrication is rare: all twelve samples have a preferred downslope dip direction significant at the 0.001 level (A_n360 statistic). Those fabrics with weak directional trends (D1 and D2) were measured at sites where debris was piled up behind an obstruction, as described above. Those with modal orientations differing from that of maximum slope (D4 and D14) were measured at sites where the debris was "streaming" obliquely between rock outcrops.

Debris flows have stripped the regolith completely from parts of the Coire a'Mhuilinn debris slope, revealing in long section a deposit of variable depth (nearly always less than 1 m) overlying a crudely "stepped" bedrock profile. The lowermost clasts are angular, blocky and sometimes in situ (figure 8.9), but those in the uppermost 25 cm tend to be slabby, oriented downslope and rounded by microgelivation. The bottom part of the deposit appears to have been produced through macrogelivation of bedrock and infilling of interstices by the products of ~~microgelivation~~ granular disintegration. The top 25 cm, however, has apparently moved downslope over the in situ deposit. Where the debris mantle is shallow only the mobile layer is found.

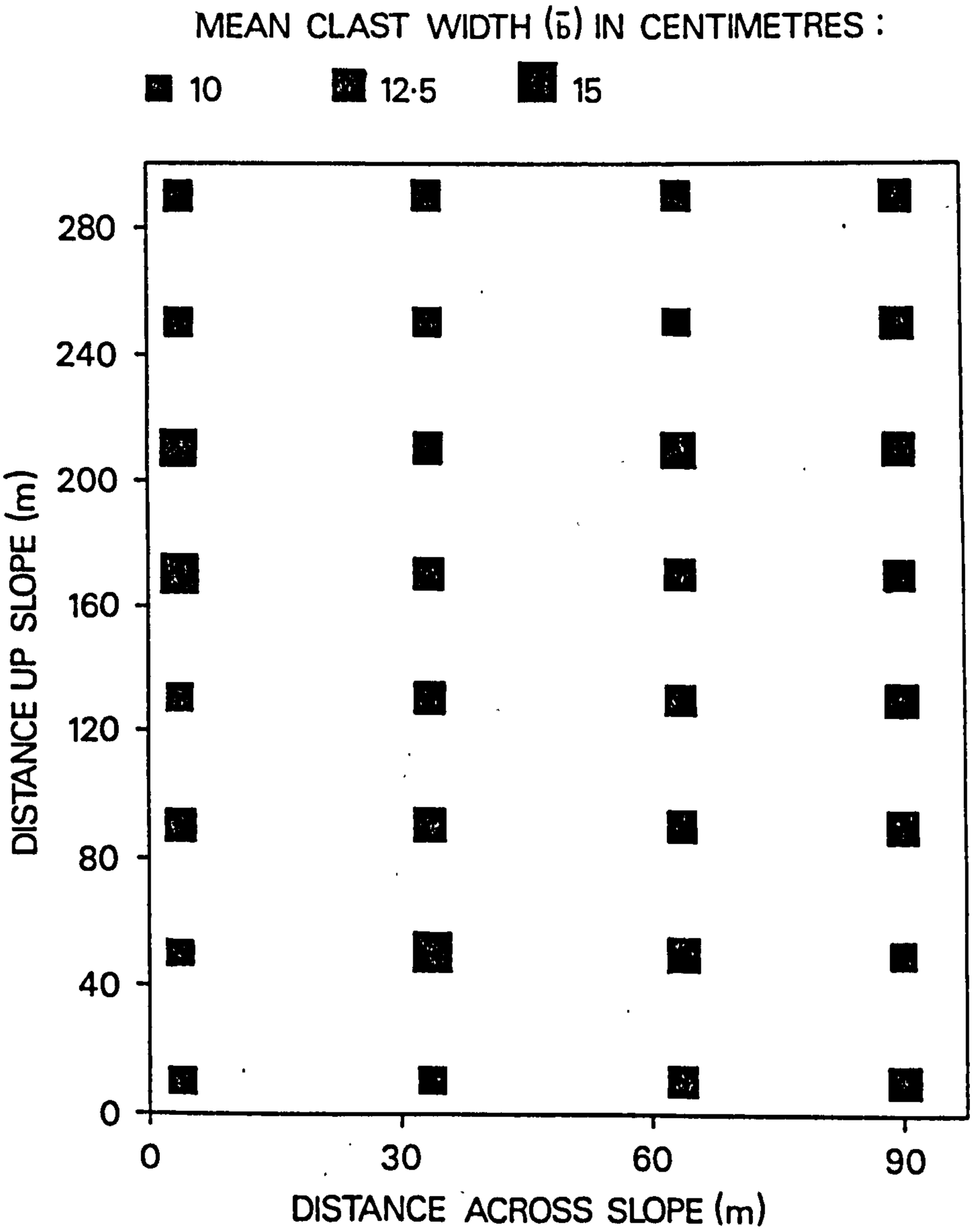


Figure 8.31: Mean clast widths (intermediate axis lengths) of 32 samples of 50 + stones on the Coire a'Mhuilinn debris slope, An Teallach.

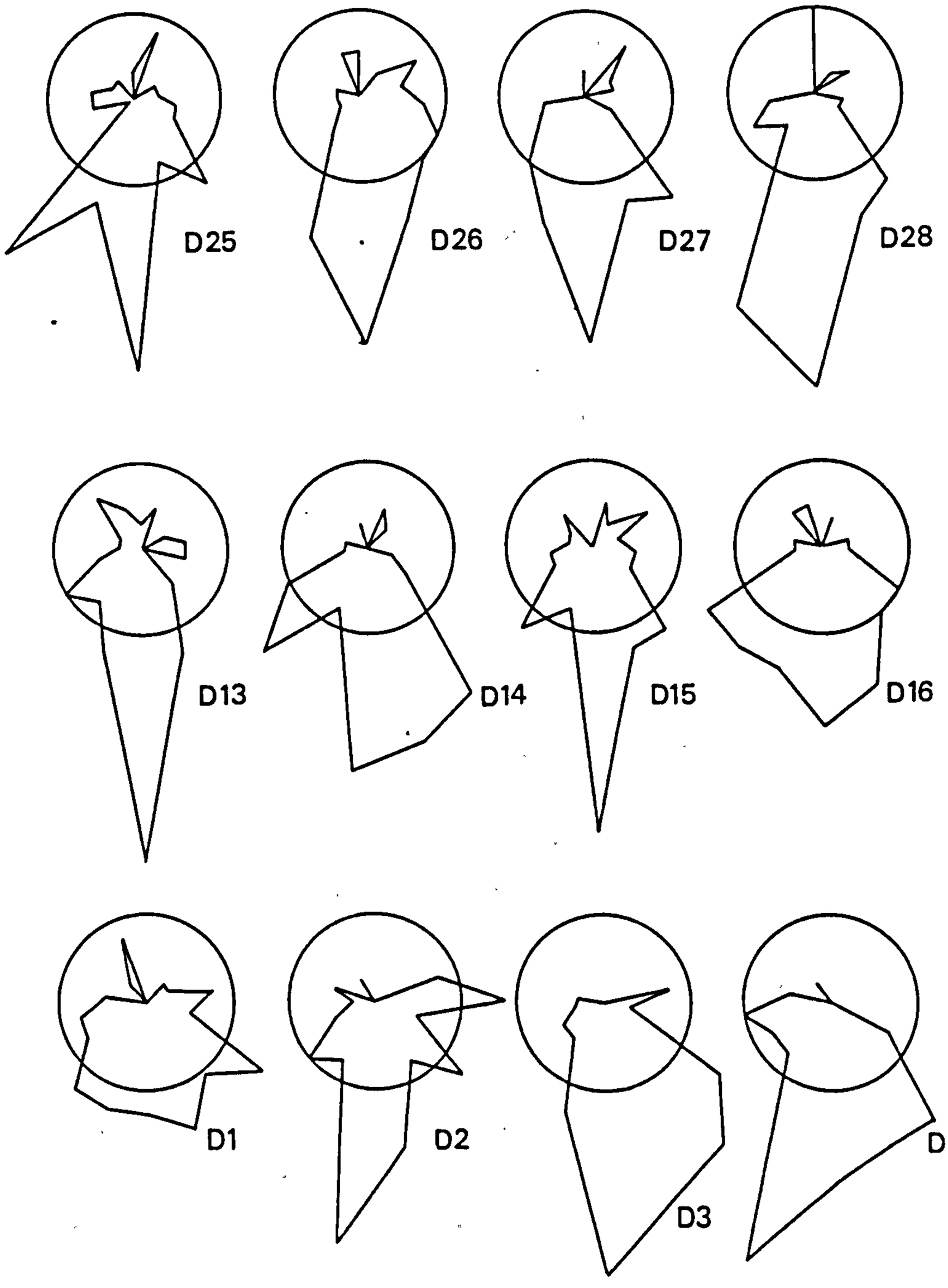


Figure 8.32: Surficial fabrics measured on the Coire a'Mhuilinn debris slope. The fabrics are plotted in the correct position relative to each other with the line of maximum slope from the top to the foot of the page. Each circle represents four observations. Sample size = 50 in each case.

8.4.5 Rates of movement on unvegetated debris slopes

There are several indications that the unvegetated debris slopes on An Teallach are undergoing movement at present. In spring, surface clasts were observed to have moved downslope out of the "sockets" of fine material that they previously occupied. In addition, needle ice was observed to heave individual clasts away from the surface, and repeated heaving of the ground was measured on a terraced debris slope (chapter 9). In order to assess current rates of movement, mass-movement sites were established at twelve points on unterraced debris slopes at different aspects, as described in the appendix The results from five of these sites were rejected as there was evidence that the marker poles had been displaced by animals or frost heave. Rates of surficial movement for the remaining seven sites over two years (1976-77 and 1977-78) are plotted on figure 8.33. These show great variation both within and between sites, with markers at every site remaining stationary or near-stationary, a maximum downslope displacement of 186.5 mm (93.2 mm y^{-1}) and four markers showing net upslope movement. Much of this variation can be explained in terms of four factors: differences in slope, a decline in displacement with depth, the rolling and sliding of clasts following slight displacement, and the nature of downslope movement.

In general, displacement increases with slope. The markers at site 2, on a slope of 16°, exhibit least displacement, and those on slopes of over 30° (5 and 6) have a relatively high average rate of movement, although the pattern is confused by the high values at site 1, on a slope of 20°. A broad correspondence can also be identified between increasing displacement and increasing altitude, so it is not certain whether increase in gradient or increase in exposure is responsible for greater movement. Slope is almost certainly a cause of intra-site variation, as the local gradient varies considerably along the line of measurement.

Decline in displacement with depth was measured (at site 4) by inserting two lengths of rubber tubing vertically into the ground and studying their displacement after two years (section

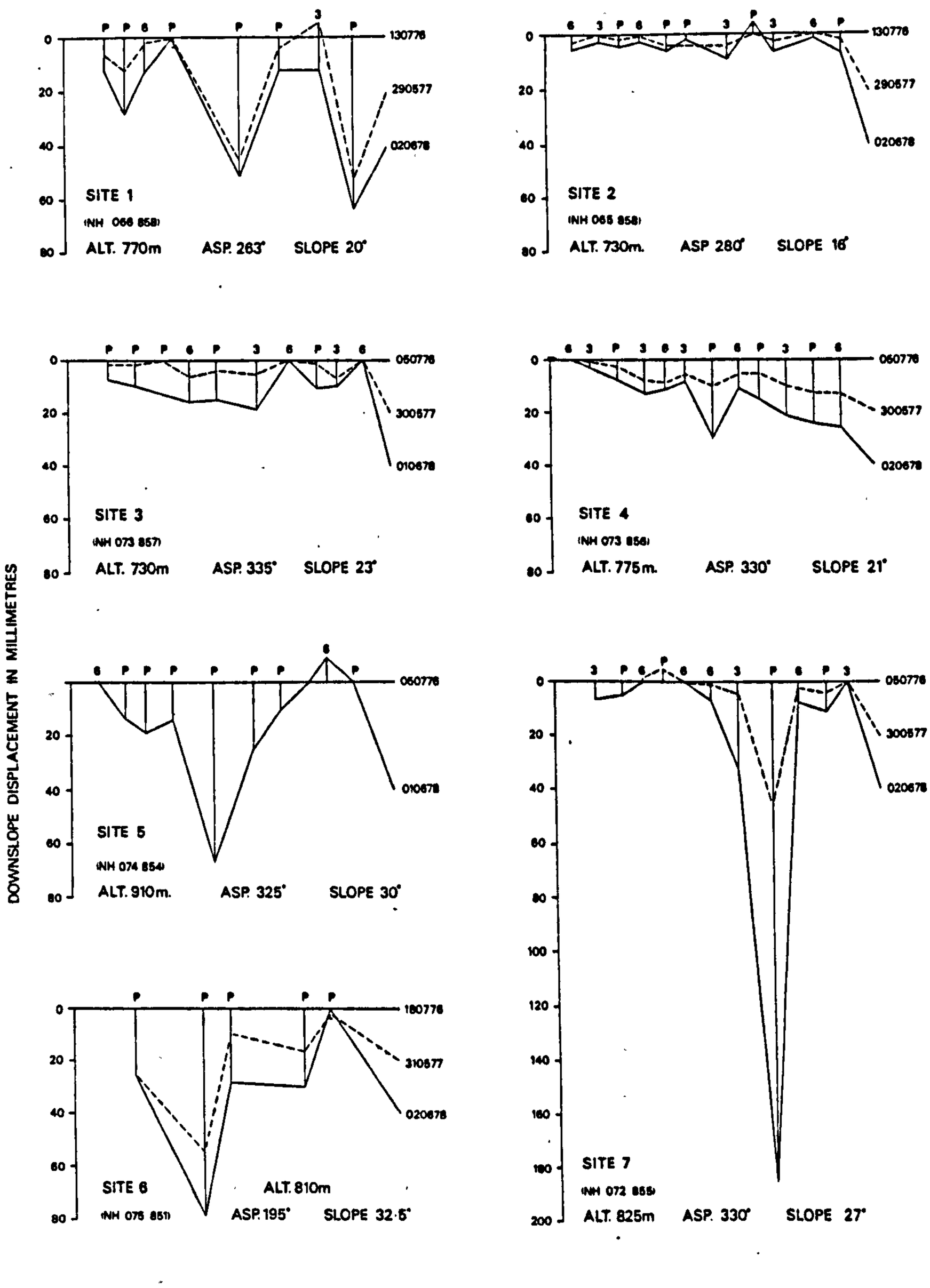


Figure 8.33: Downslope displacement of clasts (P), 3" nails (3) and 6" nails (6) at seven mass-movement sites (of the type described in appendix 1) installed across unvegetated debris slopes on An Teallach. The dashed lines mark the displacement of individual markers after one year (1976-7), the continuous lines mark displacement after two years (1976-8).

9.3.4). Unfortunately these tubes were partly displaced by frost heave, but their configuration on excavation indicated that measurable displacement (> 1 mm) could be detected only in the top 3 cm of regolith. When the displacement of different types of marker (pins adhered to surface clasts, 3" nails and 6" nails) are plotted on dispersion diagrams (figure 8.34) the median values for pins (indicating surface clast movement) are highest (5.5 and 8.9 mm y^{-1}), those for 3" nails are intermediate (3.9 and 4.9 mm y^{-1}) and those for 6" nails are lowest (2.0 and 3.6 mm y^{-1}), which also suggests that displacement decreases rapidly with depth and is confined to the top few centimetres of the regolith. The results for all three types of marker indicate greater displacement for 1977-78 than for the previous year. On figure 8.34 it is also apparent that few markers moved more than 20 mm downslope in either year, and those that did often moved a great deal farther (the majority over 40 mm y^{-1}). With one exception, all markers travelling in excess of 20 mm y^{-1} were pins mounted on surface clasts. Such clasts have apparently slid over steeper reaches following slight displacement; the exception, a 3" nail, had apparently tilted over the edge of a similar steep reach. Rolling of stones over steep banks also occurs, though less often, resulting in displacement of several cm y^{-1}, but this process tends to dislodge marker pins and such movements are not recorded on figures 8.33 and 8.34; if they were, median and mean rates of clast displacement would be slightly increased.

A picture emerges of an annual surface displacement of 0 - 2 cm y^{-1} on such slopes, with individual clasts occasionally sliding or rolling farther. Such movement apparently affects only the top few centimetres of the regolith, is highly variable over short distances, tends to increase with increasing slope and/or altitude and may involve occasional slight displacement upslope. Movement is largely restricted to the "winter" months from September to May: measurements made at sites 1 and 2 for the period 290577 - 170977 indicated slight displacement of only four of the markers. The development of needle ice at shallow depth, the movement of stones from their "sockets" described

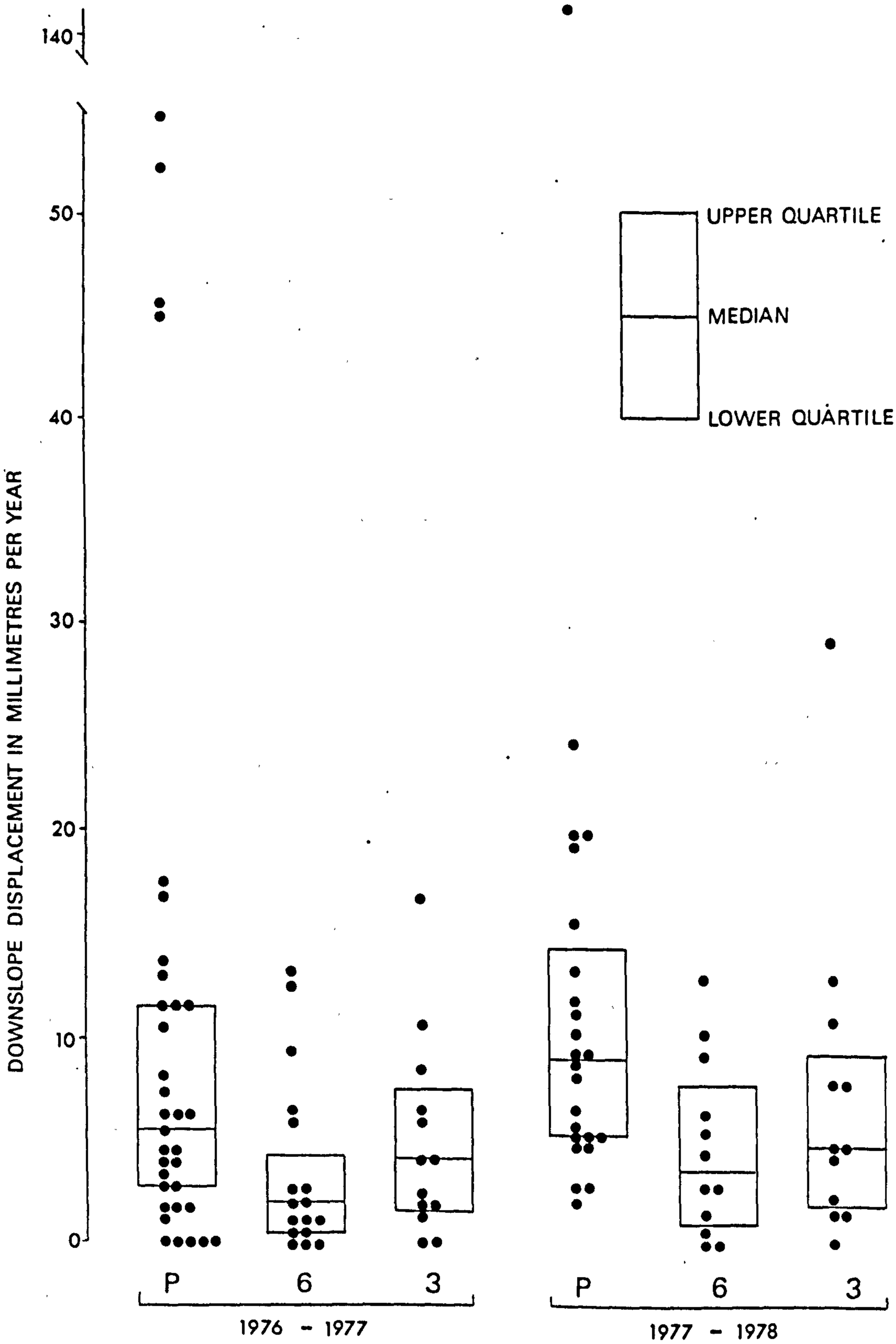


Figure 8.34: Dispersion diagram showing the range of downslope displacement of clasts (P), 6" nails (6) and 3" nails at seven sites on unvegetated debris slopes on An Teallach over the period 1976-8.

earlier and measurements reported in chapter 9 demonstrate the operation of near-surface frost heave. Indeed the characteristics of movement described above all point to frost creep as the main agency of present-day activity on unvegetated debris slopes (Caine, 1962, 1963a).

The mean rate of movement of surface clasts measured at sites 6 and 7 (on slopes over 30^{o}) is 11.2 mm y^{-1}. If this figure is considered representative and extrapolated for the period since the downwastage of the Late Devensian ice sheet (say 14,000 years), then the total downslope movement of clasts on steep unvegetated debris slopes would be around 150 m. However, Cambrian Quartzite boulders apparently weathered from the quartzite cap on Glas Mheall Mór (NH 078855) can be traced in large numbers down the steep (c. 30^{o}) unvegetated debris slopes on the underlying Torridon Sandstone for much greater distances (figure 8.35), being common up to 300 m downslope of the Torridonian-Cambrian unconformity on the north side of the hill and nearly 600 m downslope on the south side. Unless rates of recent displacement have been grossly underestimated, this discrepancy indicates that at some time in the past movement of quartzite clasts has been far more rapid than at present. In view of the apparent rapidity of blockfield movement during the Lateglacial period (section 8.3.4) it is reasonable to infer that debris slopes also underwent accelerated activity at this time, particularly on south-facing slopes. In this context it may be significant that the debris mantle on the south flank of Glas Mheall Mór is thicker than that on similar slopes north of the summit ridge, bedrock outcrops being rare on the former. This suggests that both periglacial weathering and mass-movement during the Lateglacial period were more effective on south-facing than north-facing slopes, possibly because of an increase in the number of effective freeze-thaw cycles. The validity of this hypothesis, however, requires testing over a wider area.

8.5 Synthesis and conclusions

In this section the characteristics of different types of

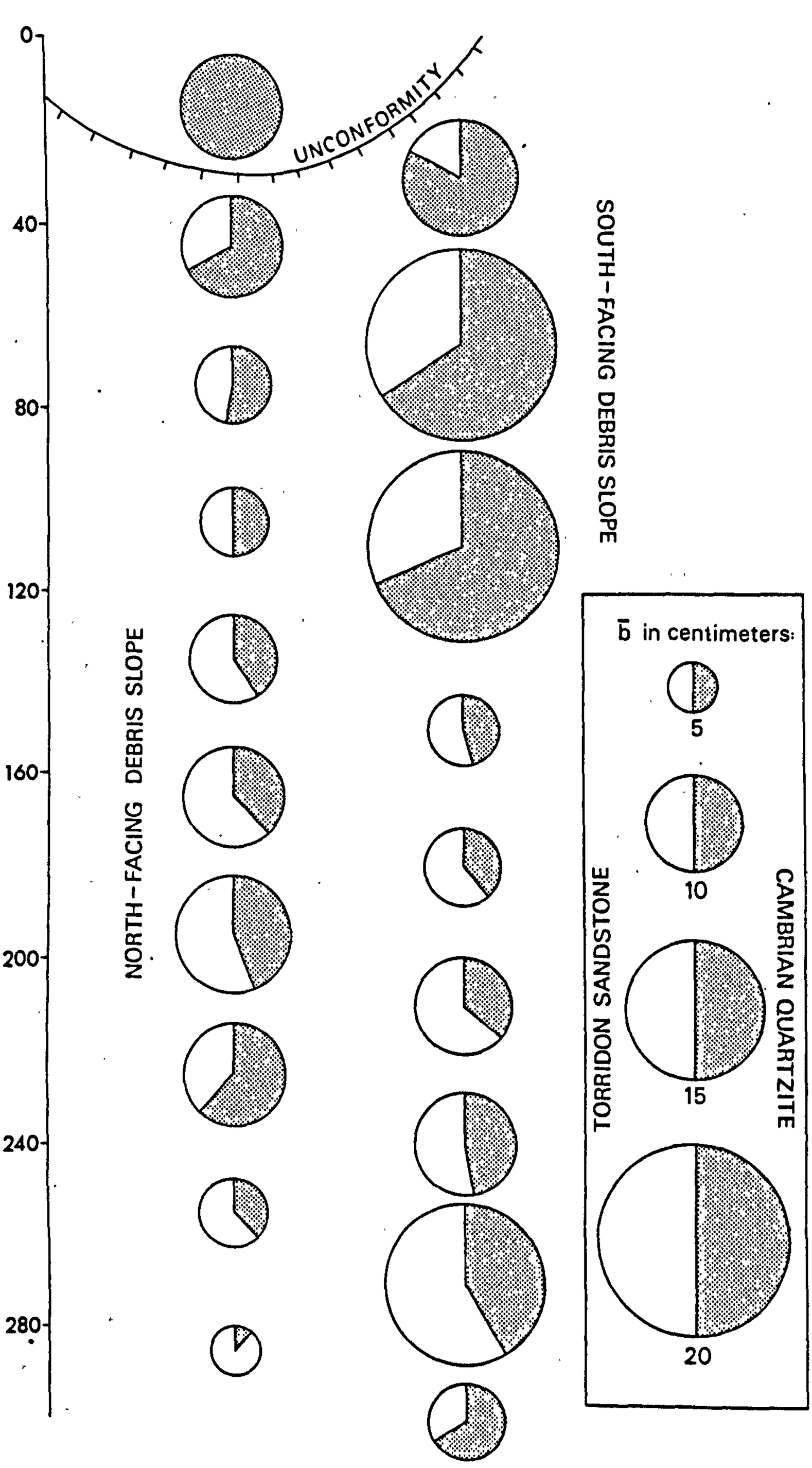


Figure 8.35: Concentrations of Cambrian Quartzite clasts downslope of the Torridonian-Cambrian unconformity on Glas Mheall Mór, An Teallach. The size the circles is proportional to the mean intermediate axis length for each sample of 50 clasts.

detritus slopes are compared in order to identify the main factors that have determined slope type, and the nature of past and present activity is summarized.

On all types of detritus, slope facets with gradients over 40° are rare (figure 8.36), and although the debris slopes and blockslopes surveyed are amongst the steepest in the area, they are less so than the upper parts of talus slopes. The Glas Mheall Liath blockslope and Ben Wyvis debris slope do not exceed 38°, and angles greater than this on the Coire a'Mhuilinn debris slope are associated with rock outcrops. From this it can be inferred that rock slopes with angles greater than around 40° do not support a detritus mantle, but form free faces. Significantly, an angle of 39 - 40° is generally considered to be the angle of residual shear of coarse aggregates (Van Burkalow, 1945; Scheidegger, 1970; Chandler, 1973); on steeper slopes detritus is liable to slide or roll and accumulate downslope as talus. Slopes with an initial gradient greater than about 40° therefore give rise to a talus/free face system, but those at lesser angles are capable of supporting an autochthonous debris mantle in the form of blockslopes or debris slopes; these are subject only to slow mass movement or infrequent localised failure in the form of shallow slab sliding or debris flow (chapter 9).

A second major determinant of slope type is bedrock response to periglacial weathering. Figure 8.37 summarizes the shape characteristics of surface clasts on the sample slopes studied. Although there is no apparent difference in elongation ($\overline{b/a}$) between types, all three measures of "blockiness" or sphericity demonstrate that only the clasts on the Coire a'Mhuilinn debris slope display the characteristic "slabbiness" of in situ weathered Torridonian rocks. The talus material is very "blocky", even more so than the frost-riven quartzite of the Glas Mheall Liath blockslope; the thickness and "blockiness" of the clasts on talus slopes suggests that macrogelivation plays little part in their genesis. The roundness of surface clasts (table 8.6) gives a clue to the importance of microgelivation on each slope type. Fresh talus is highly angular, superficial clasts on vegetated talus slightly

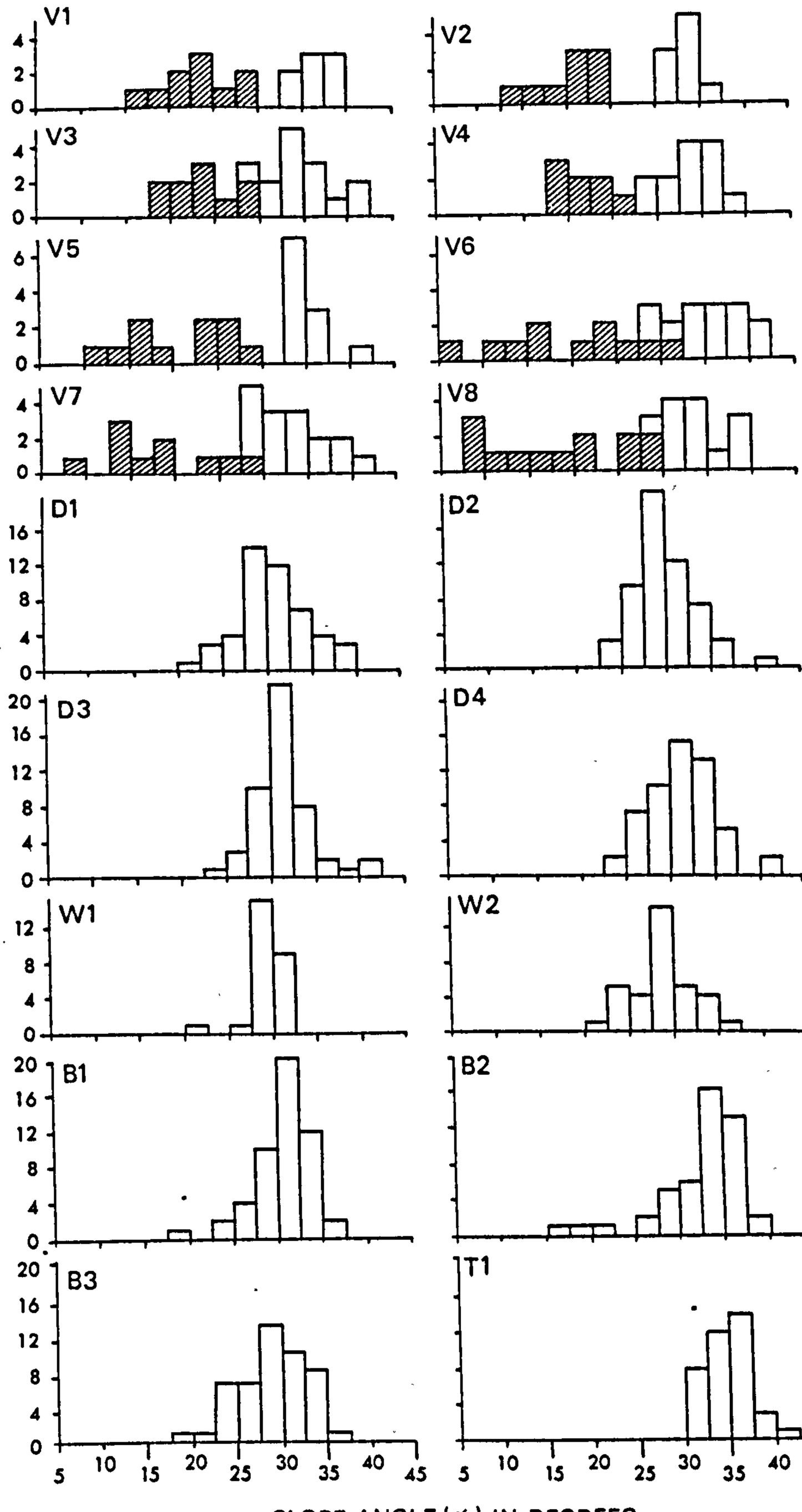


Figure 8.36: Frequency distribution of gradients of surveyed slope facets on detritus slopes in the study areas.

V 1-8: Vegetation-covered talus, Coire a'Mhuilinn (shaded bars refer to the lower zone depicted in figure 8.10).
D 1-4: Unvegetated debris slope, Coire a'Mhuilinn (figure 8.30).
W 1-2: Vegetation-covered debris slope, Ben Wyvis (figure 8.29).
B 1-3: Quartzite blockslope, Glas Mheall Liath (figure 8.19).
T1 : Glas Tholl talus cone (figure 8.4).

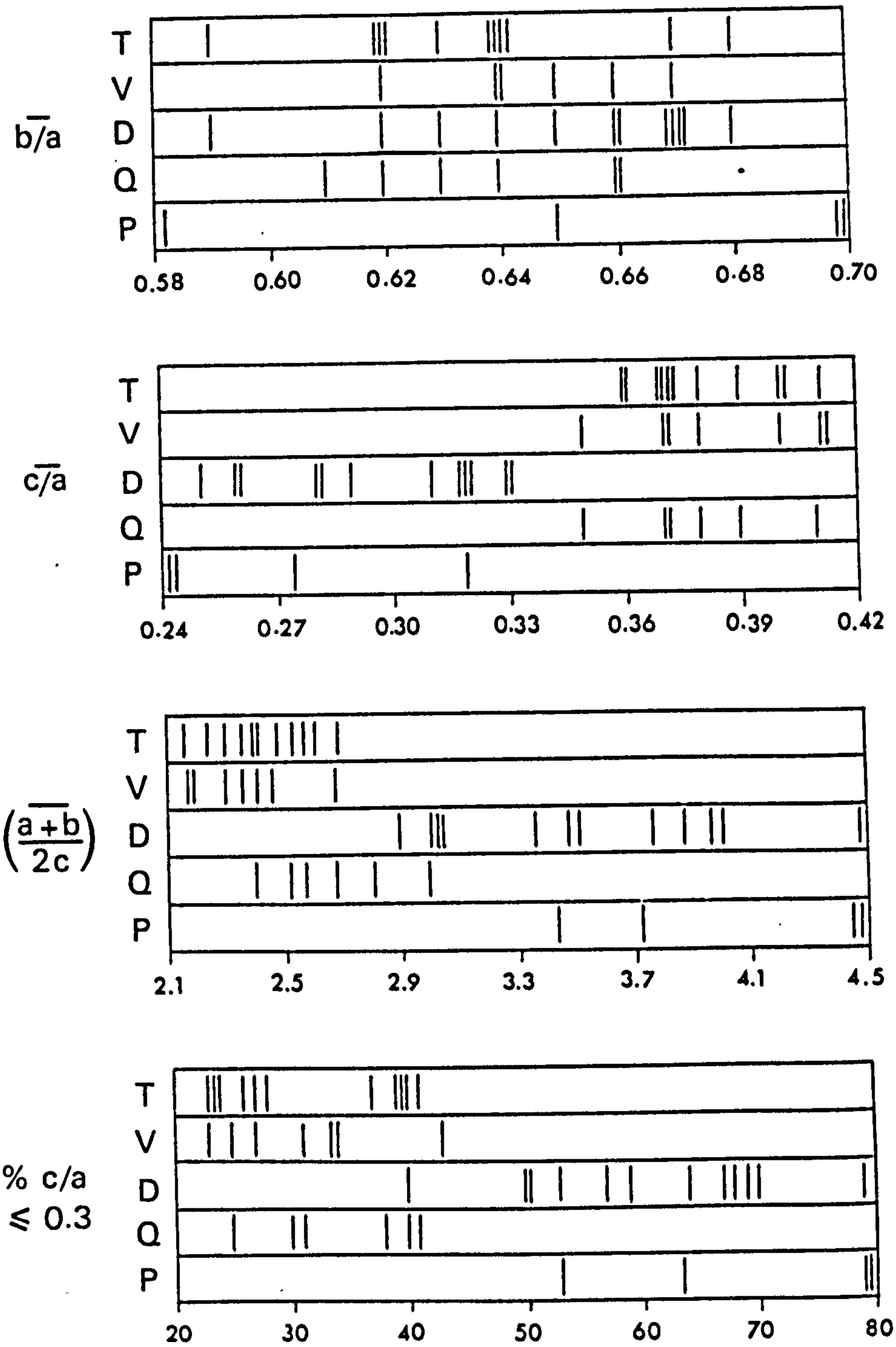


Figure 8.37: Summary of the shape characteristics of samples of 50+ surface clasts measured on detritus slopes in the study areas. T - Glas Tholl talus cone. V - Coire a'Mhuilinn vegetated talus. D - Coire a'Mhuilinn debris slope. Q - Glas Mheall Liath blockslope. P - In situ weathered slabs of Torridon Sandstone, An Teallach.

Table 8.6

Roundness of surface clasts on detritus slopes

Sample	Slope Type	Lithology	$\bar{r}$	$\bar{R}$
T 1	Talus Cone	Torridon Sandstone	0.58	74
T 2	Talus Cone	Torridon Sandstone	0.37	43
V 13	Vegetated Talus	Torridon Sandstone	0.50	56
V 19	Vegetated Talus	Torridon Sandstone	0.68	93
V 20	Vegetated Talus	Torridon Sandstone	0.68	110
V 31	Vegetated Talus	Torridon Sandstone	0.68	74
D 4	Debris Slope	Torridon Sandstone	1.11	142
D 16	Debris Slope	Torridon Sandstone	1.33	151
D 28	Debris Slope	Torridon Sandstone	1.92	218
Q 5	Blockslope	Cambrian Quartzite	2.44	132
Q 23	Blockslope	Cambrian Quartzite	0.90	106

$\bar{r}$ = mean minimum radius of curvature (cm) on the principal plane of measured clasts.

$$\bar{R} = 1000 \left[\frac{1}{n} \sum_{i=1}^{n} \frac{2r_i}{a_i} \right]$$

where the a_i are the lengths of the principal axes of measured clasts, the r_i are the minimum radii of curvature of measured clasts and n is sample size (50).

less so (as it incorporates debris slope material from the slope above the free face); quartzite blocks, despite long exposure, are only moderately rounded, but Torridon Sandstone slabs on the surfaces of debris slopes are well-rounded. These differences are reflected in the depth and development of a fine matrix in different slope deposits (figures 8.9, 8.18 and 8.24). The formation of a block deposit rather than a debris mantle would appear to depend primarily on resistance to ~~microgelivation~~ granular disintegration, and secondarily on depth of open jointing, such that in deep deposits the products of ~~microgelivation~~ granular disintegration can be accommodated in the lowest layers.

The strength of preferred downslope alignments of clastic detritus also differs amongst slope types (figure 8.38), reflecting the operation of different forms of movement. Near-surface frost creep has produced generally strong trends on debris slopes and those produced by large-scale frost creep on blockslopes are only slightly weaker. However, only about half of the talus cone fabrics display statistically significant preferred trends (reflecting the alignment of falling particles on final impact), and on vegetated talus the majority of fabrics are isotropic with respect to orientation. The majority of fabrics on all slope types show a statistically significant downslope dip direction ($A_n 360$), but again the tendency is much stronger for creep-dominated slopes than for rockfall slopes.

The only essential difference in the characteristics of detritus-mantled surfaces and detritus-mantled slopes is that the latter display signs of past and/or present movement. Thus, although a gradient of 5° has been employed (arbitrarily) to separate the two groups, a sounder procedure is to consider blockfields and debris surfaces as static deposits, and blockslopes and debris slopes as those that have undergone downslope movement, no matter how gentle the slope. If this is done, then it becomes possible to generalise as to the nature of slope types that develop under periglacial conditions on slopes inherited from glaciation in terms of only three factors, namely initial gradient, lithological properties of

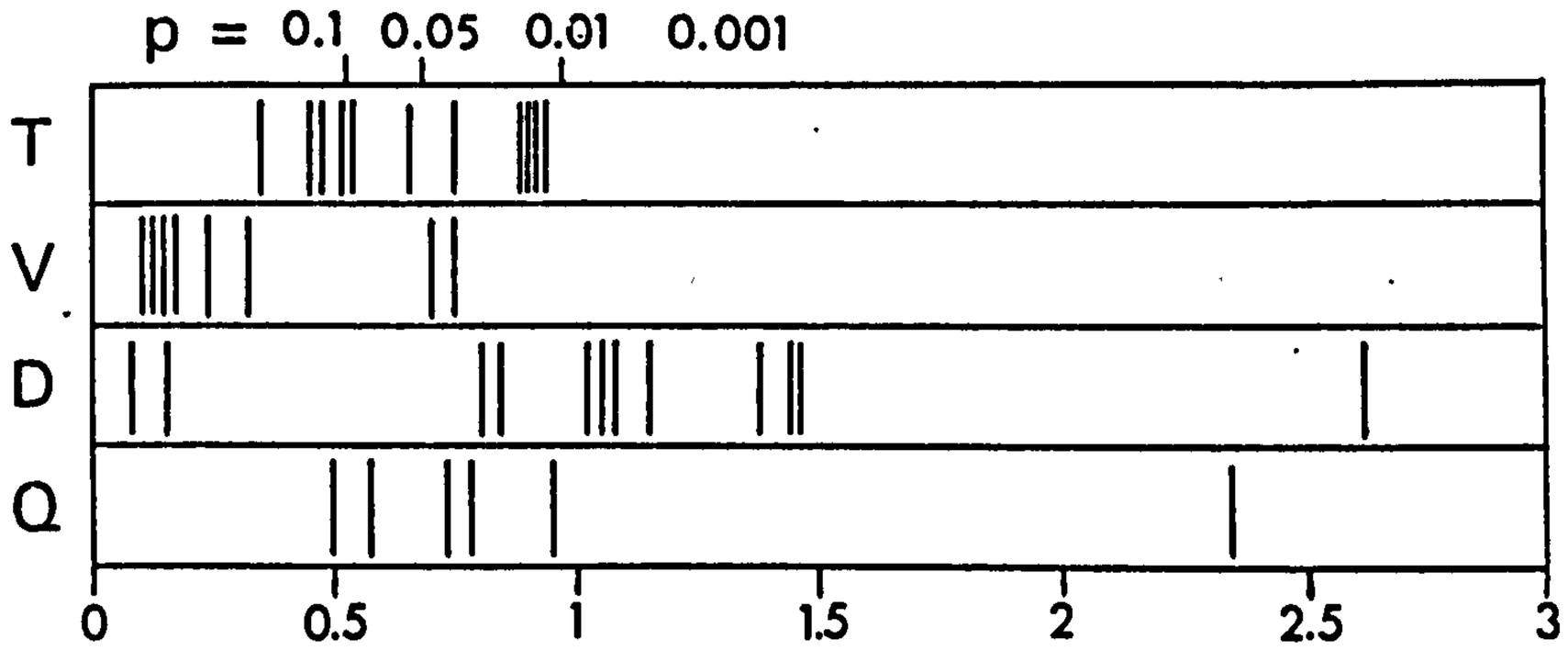
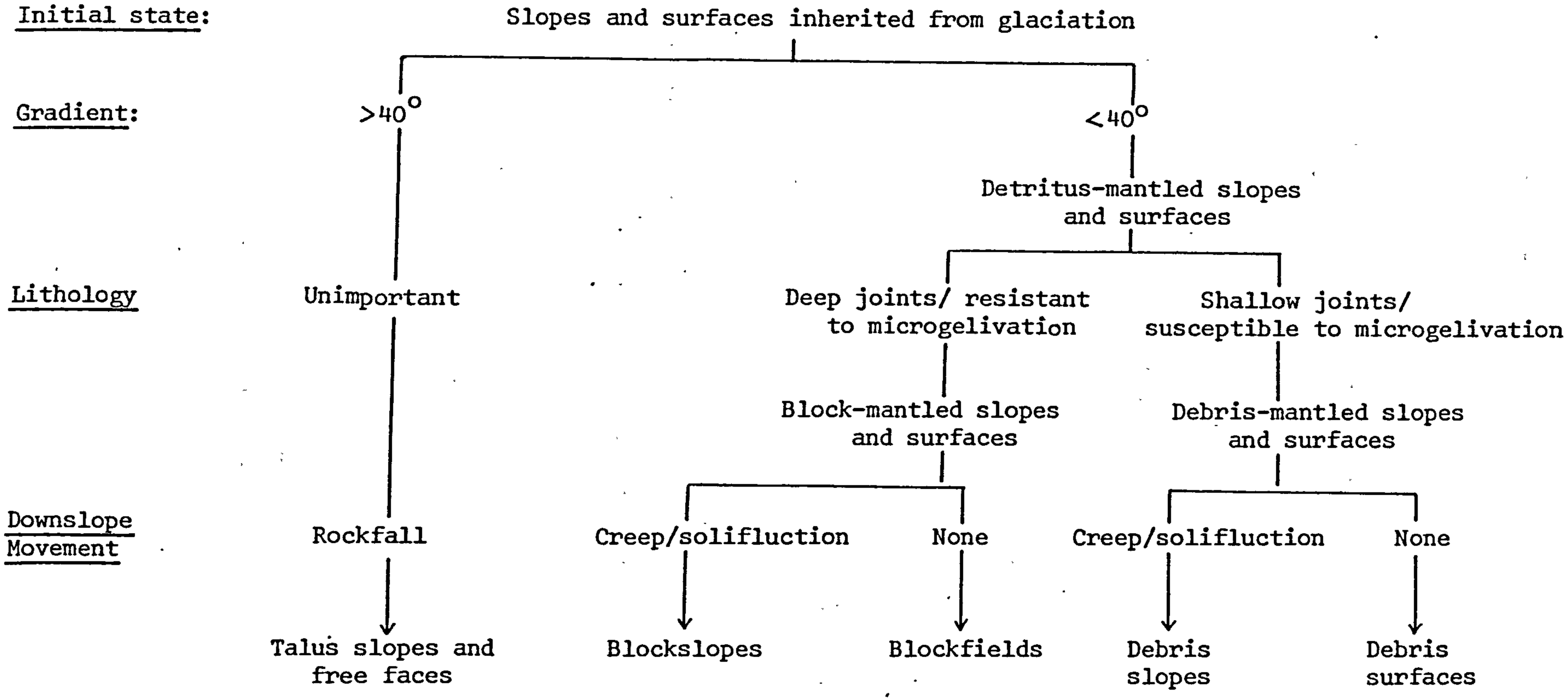


Figure 8.38: Strength of preferred orientation (top) and downslope dip for fabrics measured from samples of 50 surface clasts on detritus slopes on An Teallach. Each vertical line represents a single sample; the upper figures (p) refer to significance level, the lower figures to the strength of trend assessed using the A_n180 and A_n360 orientation statistics (Dale and Ballantyne, 1980). T - Glas Tholl talus cone. V - Coire a'Mhuilinn vegetated talus. D - Coire a'Mhuilinn debris slope. Q - Glas Mheall Liath blockslope.

bedrock and the occurrence of downslope movement (table 8.7). A fourth factor, deflation, can be used to differentiate debris surfaces and slopes into vegetation-covered and unvegetated types.

In the course of this chapter, various inferences have been made about the nature of the processes that operated under Lateglacial conditions. Rockfall was apparently very frequent, producing mature talus in unglacierized areas over periods of several hundreds of years; snow avalanches contributed to the modification of such talus; large scale frost-wedging operated at depths of over 1.6 m; and the annual freezing and thawing of the upper layers of the ground above a probable permafrost table resulted in vertical size sorting, the production of large-scale sorted patterned ground and stone pavements, and downslope creep of the detritus mantle at rates of several centimetres per year. At present many of these processes still operate, but only at a much reduced rate and scale. Rockfall, including occasional large-scale falls, is frequent from high cliffs above "immature" talus; the surface of unvegetated regolith is creeping downslope at average rates of a few millimetres per year; and many areas are subject to solifluction, frost sorting, deflation, debris flow and wash. The nature of this activity and the distinctive landforms it has produced are treated in greater depth in the next four chapters.

Table 8.7: Principal influences on slope type



CHAPTER 9 MASS-MOVEMENT FEATURES

9.1 Introduction

Landforms produced by the downslope movement of detritus constitute the most widespread manifestation of periglacial activity in upland Britain. Features formed by the slow mass-movement of regolith are particularly common, occupying virtually all slopes above 800 m and many at lower altitudes, except where these exceed approximately 35°, lack sufficient detritus or are covered by peat. It is not surprising, therefore, that mass-movement phenomena were amongst the first upland periglacial features in Great Britain to be described (e.g. Crampton, 1911; Peach et al., 1912, 1913a, b; Crampton and Carruthers, 1914) and that they have subsequently attracted considerable interest (section 3.3.3).

In general, previous studies have attempted to identify particular types of landform (e.g. lobes, terraces, ploughing boulders) and to explain the characteristics of each type in isolation. Although this procedure is necessary in view of the range of different forms that have resulted from periglacial mass-movement, it has tended to obscure the relationship between types and to leave a fundamental question unasked: why have similar slopes responded in different ways to periglacial conditions? Comparison of maps 2 (An Teallach) and 5 (Ben Wyvis) shows that whereas the flanks of the former are occupied by featureless debris slopes and various types of turf-banked terrace, the latter is dominated by sheets and lobes. What factors underlie such differences? In the present study an attempt is made not only to explain the characteristics of individual types of landform, but also to demonstrate the relationships between these types and the principal factors responsible for their distribution.

9.1.1 Problems of classification

Although many of the features in the mass-movement "family" described in chapter 7 can be clearly and unambiguously defined, landforms produced by the slow mass-movement of regolith form a

complex morphological continuum that defies simple or rigid classification. The failure of previous workers (e.g. Galloway, 1958, 1961a; King, 1968, 1972; Kelletat, 1970a, b; Ball and Goodier, 1970; Shaw, 1977) to recognise the existence of such a continuum led to their treating related features in isolation. Study of maps 2-8, however, reveals many instances of morphological transition: that from straight-fronted detritus sheets through lobate sheets to individual lobes, for example, is illustrated on the N.W. flank of Ben Wyvis; the slope S.W. of Scùrr Mór on map 4 shows a transition from horizontal terraces via oblique terraces to lobate sheets; and on several areas of the An Teallach plateau (e.g. 065853) there is a transition from horizontal terraces through interconnecting terraces to oblique terraces. The complete range of transitions known to the author is illustrated on figure 9.1 which, though schematic, is based closely on field sketches made in the study areas.

Two elements are common to all of the features illustrated in figure 9.1 and the accompanying photographs (figures 9.2-9.18), namely a step-like riser with an angle exceeding that of the general slope and a tread with a gradient less than or equal to that of the general slope. Forms possessing these elements were distinguished according to three criteria:

(i) Relationship between tread angle (α_t) and general slope angle (α). For features on the left hand side of figure 9.1 (sheets and lobes) α_t is only slightly less than α ; for non-oblique terraces, however, α_t rarely exceeds 12^{o}, irrespective of the value of α . For example, regression of α_t against α for a sample of 118 lobes (see below) yielded the relationship

$$\alpha_t = 1.063\,\alpha - 2.161 \qquad (r^2 = 0.897,\ p<.00001) \qquad (9.1)$$

implying that on average α_t is about 2^{o} less than α . For a sample of 333 terraces, the relationship proved nonlinear and less regular:

$$\alpha_t = 0.886\,\alpha^{0.779} \qquad (r^2 = 0.336,\ p<.00001) \qquad (9.2)$$

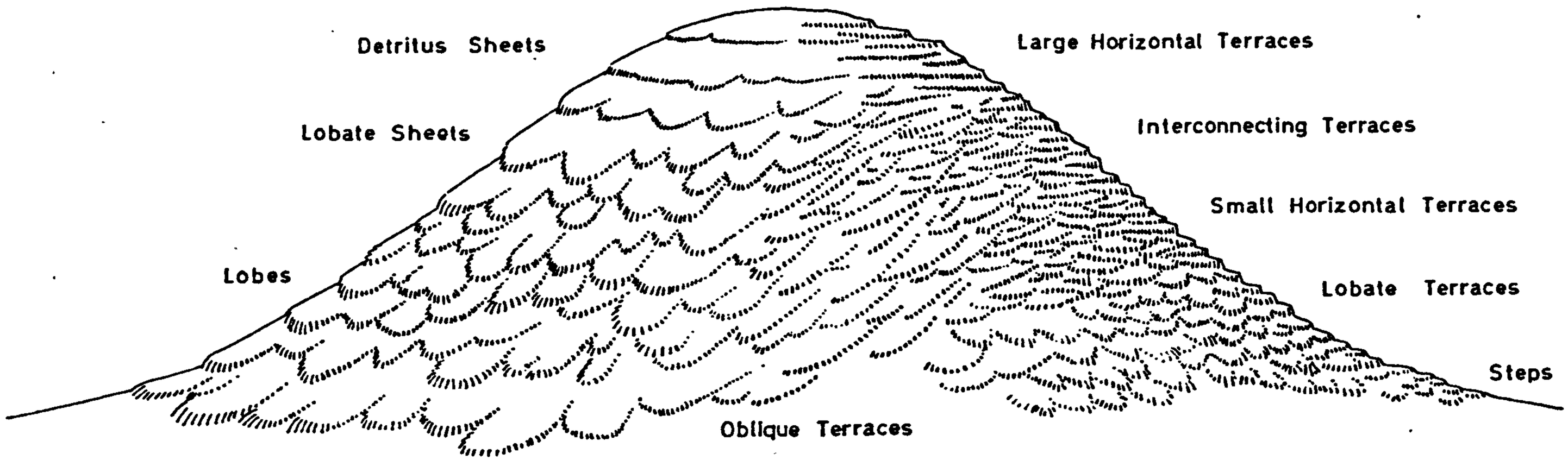


Figure 9.1: Diagram illustrating the range of forms produced by slow downslope movement of regolith on the mountains of northern Scotland. Schematic; based on field sketches.

In this case the exponent indicates that as α increases, the difference $(\alpha - \alpha_t)$ also increases: for $\alpha = 10^{\circ}$, $\alpha_t = 5.3^{\circ}$; for $\alpha = 20^{\circ}$, $\alpha_t = 9.1^{\circ}$; for $\alpha = 30^{\circ}$, $\alpha_t = 12.5^{\circ}$. Many previous authors (e.g. Galloway, 1958, 1961a; King, 1968, 1972; Shaw, 1977) have referred to sheets as "terraces", but this term is obviously inappropriate for features that slope downhill at an angle only slightly less than the general slope angle, and is here reserved for features with α_t significantly lower than α.

(ii) Plan form of the riser. A crenulate plan distinguishes lobate sheets and lobate terraces from other types. Individual lobes are often formed through the over-riding of one lobate sheet by another, and the features widely referred to in the literature as lobes generally represent the U-shaped frontal parts of detritus sheets or terraces. It follows, therefore, that many lobes may properly be regarded as secondary features that reflect differing rates of mass-movement at the fronts of sheets or terraces.

(iii) Obliquity relative to contours. This criterion was used to distinguish terraces that are aligned approximately parallel to the contours from those that dip (often steeply) across-slope. The interaction of the two types on the same slope produces a network of horizontal and oblique treads and risers, here referred to as interconnecting terraces.

Previous studies of periglacial mass-movement features in upland Britain have usually distinguished lobes and terraces (the latter term often used to refer to non-lobate sheets) as fundamentally different types. A more meaningful distinction exists, however, between the features described here as sheets and terraces; lobes, although often the most eye-catching features on a hillslope, are best regarded as adjuncts to these primary types, particularly the former (figures 9.5-9.8).

9.1.2 Structure of the chapter

Given the range of types of mass-movement feature in the study areas (of those identified in the classification scheme



Figure 9.2: Turf-banked debris sheet at 760 m altitude, northern Plateau of An Teallach.



Figure 9.3: Vegetation-covered boulder sheet, probably of Lateglacial age, on low-angle (7^{o}-11^{o}) slopes at 1000 m altitude on Ben Wyvis.



Figure 9.4: Lobate solifluction sheet at 940 m altitude on A'Chailleach (western Fannichs).



Figure 9.5: Lateglacial boulder lobes developed on a microgranite blockslope at 500 m altitude on Sron an t-Saighdeir, Rhum.



Figure 9.6: Active solifluction lobe on a 29.5° slope at c. 850 m altitude on the N.W. flank of Ben Wyvis.



Figure 9.7: Active solifluction lobes on a 24° slope at c. 800 m altitude on the N.W. flank of Ben Wyvis.



Figure 9.8: Vegetation-covered lobe at the basal break of slope. Photographed at c. 750 m altitude at the head of Coire Mór, An Teallach.



Figure 9.9: Small active solifluction lobe at 850 m altitude on Meall a'Chrasgaidh, eastern Fannichs.



Figure 9.10: Lobate boulder sheet at 920 m altitude on Scùrr Breac, western Fannichs.



Figure 9.11: Vegetation-covered solifluction sheets at 850 m altitude above Coire Granda, Ben Wyvis. These features exhibit a transition between solifluction sheets and vegetation-covered terraces.



Figure 9.12: Horizontal turf-banked terraces (deflation terraces) at 600 m altitude on the Ruinsival-Scùrr nan Gillean ridge in the Rhum Cuillin. These features grade into adjacent wind stripes.



Figure 9.13: Interconnecting turf-banked terraces on a north-facing slope at 720 m, An Teallach.



Figure 9.14: Large intersecting turf-banked terraces on a south-east facing slope at 730 m altitude, An Teallach.



Figure 9.15: Small horizontal turf-banked terraces or steps on an east-facing slope at 780 m, An Teallach.



Figure 9.16: Turf-banked steps at 710 m altitude in Coire Mór, An Teallach.



Figure 9.17: Oblique turf-banked terraces on the south flank of Glas Mheall Mór, An Teallach.



Figure 9.18: A small oblique terrace on a north-facing slope at 750 m altitude on An Teallach.

only rock glaciers are absent) it is necessary, at least initially, to consider different types separately. Following the distinction made above, detritus sheets and associated lobes are considered in section 9.2 and terraces (including oblique terraces) in section 9.3. Ploughing boulders form a third distinct category (section 9.4), and features produced by rapid mass-movement are considered in section 9.5. The significance of such features in terms of the response of hillslopes to periglaciation is explored in the concluding section (9.6).

9.2 Lobes and sheets

9.2.1 Previous research in upland Britain

For convenience, the terminology employed in this review follows the scheme outlined in chapter 7 rather than that used in the original sources as the latter is non-systematic and sometimes misleading.

Although early descriptions of detritus sheets and lobes in the Cairngorms appeared in papers written by ecologists (Watt and Jones, 1948; Metcalfe, 1950), the first geomorphological study of such forms in Scotland was that carried out by Galloway (1958, 1961a) in several mountain areas. Galloway believed that what he termed "solifluction" features are widespread throughout Scotland at altitudes above 600 m, a conclusion subsequently attested by the survey carried out by Kelletat (1970a, b) and by reports of lobes and sheets on high ground in such widely-separated areas as the Hebrides (Ryder, 1968; Ryder and McCann, 1971; Birks, 1973; Ballantyne and Wain-Hobson, 1980), the Northern Highlands (Godard, 1965; Mottershead and White, 1969; Robinson, 1977; Sissons, 1977a), northern England (Tufnell, 1969; Sissons, 1980) and Wales (Goodier and Ball, 1969; Ball and Goodier, 1970).

Largely on the basis of observations made on Ben Wyvis and Lochnagar, Galloway differentiated relatively small "turf-banked" (actually vegetation-covered) lobes from "stone-banked" (stone- and turf-banked) types. The former he believed to be active, the latter of Lateglacial or early Postglacial age. All he

attributed to "solifluction", but he did not elaborate on the mechanisms of this process. In contrast, King (1968, 1972) considered the vegetation-covered lobes of the Cairngorms to be older (early Postglacial) than the stone-banked variety (Little Ice Age) and attributed the formation of all types to "saturated flowage". His conclusions regarding age, however, were largely dictated by adherence to Sugden's (1970a) interpretation of deglaciation in the Cairngorms, subsequently demonstrated to be incorrect (Sissons, 1979b). Sugden (1970b, 1971) himself considered small lobes active and large lobes relict, but later (1973) criticized the use of this form of differentiation. Mottershead and White (1969) differentiated stone-banked lobes, turf-banked terraces and vegetation-covered lobes on Arkle (Northern Highlands), but attributed the formation of different types primarily to slope angle rather than changing environmental conditions. Kelletat (1970a, 1971) distinguished active "rasenloben" (vegetation-covered lobes), active "solifluktionsloben" (debris lobes) and inactive "blockloben" (boulder lobes) and equivalent sheet forms, but Birks (1973) identified only two primary types (vegetation-covered lobes and sheets, stone-banked lobes and sheets) on Skye, and Whyte (1970) found only two types (turf-banked and stone-banked) in the Mamores; the former he considered active, the latter "partly active".

The above summary indicates the confusion that exists in the study of lobes and sheets in upland Britain. Part of this is terminological, in that various authors have used the same terms to refer to different features or vice versa. Partly it results from research in areas in which only a limited range of features occurs. It is also possible that the use of vegetation cover or size to delimit types is inappropriate. In the present study an attempt is made to resolve the confusion by determining (i) meaningful criteria for distinguishing fundamentally different types, (ii) the age of different types, and (iii) their mode of formation.

The literature provides some indications as to how these aims may be met. Sissons and Grant (1972) found that large stone-banked or boulder lobes and sheets in the Lochnagar area are absent

from ground that was occupied by glaciers during the Loch Lomond Stadial, yet are abundant immediately outside the former glacial limit. This relationship, subsequently found to hold good for other parts of upland Britain (e.g. Sissons, 1977a, 1979b), implies that these features have not been active since final deglaciation, c. 10,500-10,000 B.P. In contrast, samples of organic material taken from under a "turf-banked" (vegetation-covered?) lobe in the Cairngorms and a vegetation-covered sheet in the N.W. Highlands yielded radiocarbon dates in the range 5441±55 to 2,680±120 B.P. (O'Brien, cited in Sugden, 1971; Mottershead, 1978). These dates indicate mass-movement of certain types of lobes and sheets in the second half of the Flandrian period.

Descriptions of lobes and sheets of presumed Lateglacial age and those active in the late Flandrian indicate that systematic differences may exist between the two. One fundamental difference appears to be the nature of their internal constitution. Shaw (1977) trenched granite lobes of presumed Lateglacial age and found " ... a mass of boulders without visible interstitial fines" (p. 111), and King (1968, 1972) described a mixture of boulders and soil in a stone-banked lobe located outside the Loch Lomond limit in the Cairngorms. In contrast, the vegetation-covered sheet overlying mid- to late-Flandrian peat described by Mottershead (1978) consisted mainly of sand with some angular stones.

Another distinction of possible significance is that of size. Several authors (e.g. Ball and Goodier, 1970; Sugden 1971; Sissons, 1976a) have considered Lateglacial features to be generally larger than more recent forms. In particular, lobe thickness (expressed as riser height) may be expected to be generally greater for features formed under the severe conditions of the Loch Lomond Stadial when frost penetration must have been much deeper than now or indeed any time during the Flandrian (Williams, 1961). Several observations support a distinction based on riser height: Galloway, for example, described boulder sheets 1-5 m high on White Mounth (Eastern Grampians) and vegetation-covered lobes 0.3-1.3 m high in the same area, and McMillan (1978) found a significant difference in the heights of turf-banked and vegetation-covered lobes on a single slope on Ben Wyvis. Shaw (1977), however,

established that variations in riser height reflect size of constituent detritus as well as differences in the types or ages of features.

The validity of using vegetation cover to distinguish fundamentally different types of sheets and lobes is less certain. King (1968, 1972) found that stone-banked and vegetation-covered lobes in the Cairngorms are of similar size and constitution, and Shaw (1977) concluded that vegetation cover on boulder lobes on Lochnagar merely reflectspatterns of moisture concentration or exposure.

In summary, present information suggests that there are two fundamental types of lobes and sheets in upland Britain: relatively large relict forms dating from the Lateglacial cold periods and containing predominantly coarse detritus, and smaller (or at least less thick) lobes and sheets that were active during at least the latter part of the Flandrian period. In the pages that follow, this distinction is treated as hypothesis. Following Shaw (1977), features or coarse detritus are hereafter referred to as boulder sheets and lobes. Those composed of predominantly fine material are referred to as solifluction sheets and lobes.

9.2.2 Methodology

To test this hypothesis, four approaches were employed:

(i) Mapping and analysis of distribution. The procedure employed in mapping was described in chapter 7. Before the maps could be analysed, however, the five categories of lobes and sheets recognised in the a priori classification had to be reduced to the two types identified by the hypothesis (section 9.2.3). The distribution of these two types was then analysed in terms of altitude, aspect, slope, lithology and the limits of former glaciers.

(ii) Morphometric analysis. Although lobes were described above as adjuncts of detritus sheets rather than primary features, analysis of lobe morphology provides a useful tool for investigating the differences between types and the relationships between lobe form and environment. This type of analysis was first applied in upland Britain by King (1968, 1972) and Whyte (1970), refined by Shaw (1977)

and used to test the differences between types by McMillan (1978). In the present study distances and heights were measured using a 30 m tape, orientations using a Suunto compass and slope angles with an Abney level. Figure 9.19 illustrates the measurements made on each lobe, and table 9.1 defines measured and derived variables.

(iii) Sedimentological analysis. The structure and internal constitution of different types of lobe were investigated not only to establish the characteristics of different types, but to allow inferences to be made concerning the nature of mass-movement and lobe formation. Trenches were dug upslope from the snouts of representative examples of different types of lobe until a downslope section at least 3 m in length was exposed. These sections were mapped and samples of fine material removed for grain-size analysis. Fabric and clast-size measurements were made at the front, both sides and (in some cases) centre of lobes of each type.

(iv) Measurement of current rates of movement. Mass-movement sites of the type described in the appendix were set up across various types of lobe (including those thought to be relict) in an attempt to detect present activity.

9.2.3 Characteristics of sheets and lobes in the Northern Highlands

9.2.3.1 Vegetation cover

Sheets and lobes in upland Britain have often been classified on the basis of the presence or absence of vegetation on risers or treads (e.g. Galloway, 1958, 1961a; Ryder, 1968; King, 1968, 1972; Goodier and Ball, 1969; Mottershead and White, 1969; Whyte, 1970; Ryder and McCann, 1971; Birks, 1973; McMillan, 1978). The use of terms such as "stone-banked", "turf-banked" and "vegetation-covered" to describe sheets and lobes in upland Britain follows a long-established international tradition (e.g. Dutkiewicz, 1962; Benedict, 1970, 1976; Harris, 1972, 1973, 1977; Embleton and King, 1975). This traditional scheme was employed in the present study (chapter 7) as the distribution of vegetation cover on a sheet or lobe is readily assessed and thus facilitates subdivision of types for mapping purposes. The categories of "debris" and "boulder" sheets and lobes were added

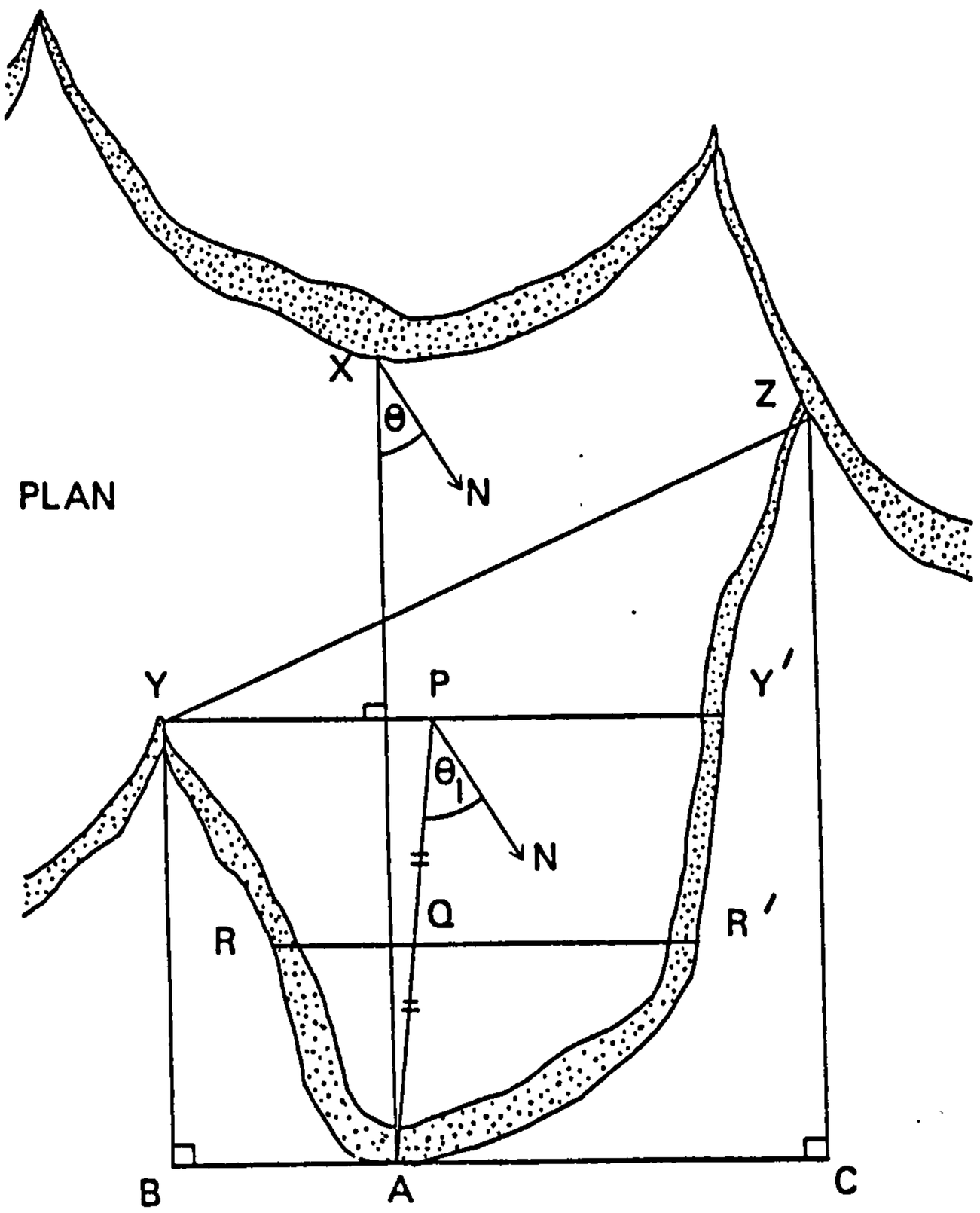


Figure 9.19: Definition of morphological variables measured on lobes (see table 9.1). Stippled area represents the lobe riser.

Table 9.1

Definitions of measured and derived variables used in morphometric analysis of lobes

Variable name	Symbol	Definition (see figure 9.19)
Width	w_1	The line YY'
Midpoint width	w_2	The line RR' (where PQ=QA)
Mean riser length	$\bar{\ell}$	(YB + ZC)/2
Lobe length	ℓ	The line PA (where YP=PY')
Spacing	L	The line XA
Riser width	w_r	The line A'D
Riser height	h_r	The line AD
Shape	Sh	w_1 / w_2
Ratio	Rt	w_1 / ℓ
Aspect	θ	Bearing of XA relative to magnetic north (N).
Orientation	θ_1	Bearing of PA relative to magnetic north (N).
Skew	θ'	$\lvert \theta - \theta_1 \rvert$
Slope angle	α	General slope angle
Tread angle	α_t	See figure 9.19
Riser angle	α_r	Tan^{-1} (AD / A'D)
Area	A	$(w_1 \lvert ZC - YB \rvert)/2 + (YB.w_2) \times 3/4 + ((w_1 - w_2) YB)/4$ (where YB is the length of the shorter riser).

to describe features for which vegetation cover is incomplete but not concentrated on riser or tread. Shaw (1977), however, questioned the validity of a classification based on vegetation cover, contending that the present distribution of vegetation on a sheet or lobe may reflect local variations in exposure or moisture, and have little relevance in terms of the age, formation or activity of such features.

If it is assumed that differences in the constitution of boulder sheets and lobes and solifluction sheets and lobes represent differences in age, formation or both, the weakness of a classification based on vegetation cover becomes apparent. The turf-banked sheets and lobes of Ben Wyvis appear to be composed largely of coarse detritus, but turf-banked solifluction sheets and lobes are found in the Cairngorms (Kelletat, 1970a, plate 5). Similarly, although most vegetation-covered sheets and lobes are of the solifluction type (figures 9.5-9.9), some massive vegetation-covered features are composed mainly of boulders (figure 9.3; King, 1972), as are all of the stone-banked lobes investigated. The present distribution of vegetation cover on sheets and lobes apparently has little significance in terms of their age and formation, although it may influence current activity (Antevs, 1932; Sharp, 1942a; Embleton and King, 1975; Washburn, 1979).

9.2.3.2 Size and morphology

Differentiation of boulder features from solifluction features is facilitated by corresponding differences in morphology and size. Table 9.2 summarizes the means of seven variables for three groups of solifluction lobes and five groups of boulder lobes in Scotland and for 20 active solifluction lobes in the Jotunheimen Massif, Norway. Several points emerge:

(i) Although sampled from different areas, all three groups of Scottish solifluction features have very similar mean dimensions. This suggests that all features considered (on the basis of constitution) to be solifluction lobes are indeed of the same type. In every size dimension the Scottish lobes are smaller than the Norwegian lobes. This suggests that size of feature is related to climate, as inferred by Williams (1961).

Table 9.2: Solifluction lobes and boulder lobes: mean values of certain dimensions and vegetation cover

	n	Length ($\bar{\ell}$) (metres)	Width (w_l) (metres)	Riser Height (h_r) (metres)	Riser slope (α_r) (degrees)	Riser Width (w_r) (metres)	Riser Vegn. (% cover)	Tread Vegn. (% cover)
Ben Wyvis: Solifluction lobes	88	4.6	5.7	0.52	46	0.54	98	100
East Fannichs: Solifluction lobes	30	5.2	6.1	0.52	36	0.71	93	62
An Teallach: Solifluction lobes	4	5.5	6.4	0.62	46	0.67	100	100
Storbreen (Norway): Solifluction lobes	20	7.3	8.8	0.97	42	1.38	100	100
Ben Wyvis: "shallow" T.B. boulder lobes	65	4.6	7.3	0.75	51	0.65	84	43
Ben Wyvis: "massive" S.B. boulder lobes	5	9.1	7.1	1.00	20	2.92	60	100
Ben Wyvis: "massive" V.C. boulder lobes	20	7.6	7.6	1.20	19	3.94	97	100
An Teallach: "massive" S.B. boulder lobes	8	9.7	9.6	2.40	29	5.76	34	49
An Teallach: "massive" V.C. boulder lobes	9	7.7	9.4	2.00	22	4.89	97	96

T.B. - Turf-banked S.B. - Stone-banked V.C. - Vegetation-covered

(ii) In general, features identified as boulder lobes are larger in all respects than solifluction lobes (apart from the Norwegian examples) and are characterized by a gentler riser slope. The exceptions to this are the shallow turf-banked boulder lobes on Ben Wyvis.

(iii) As might be expected, vegetation cover tends to be more complete on solifluction lobes than on boulder lobes, though this is not always so.

Differences in the dimensions of solifluction lobes and boulder lobes are illustrated in figure 9.20. For each of the variables illustrated, differences between the two types were tested at the 0.05 significance level using the Mann-Whitney U statistic. This indicated no difference in length between the two types, but that boulder lobes are significantly wider, with significantly higher and wider risers. Interestingly, the test revealed no differences between the slope angles on which the two types are found, and indeed figure 9.20 reveals almost identical distributions of slope angle for the solifluction and boulder lobe samples. King (1968, 1972) and Mottershead and White (1969) believed that gradient determined the distribution of lobes of different types, but the data summarized in figure 9.20 indicate that boulder lobes and solifluction lobes occupy an almost identical range of slopes (10^{o} - 36^{o}).

9.2.3.3 Configuration

Figure 9.1 depicts a transition from relatively straight-fronted detritus sheets via lobate sheets to individual lobes. This transition is superbly illustrated by vegetation-covered solifluction features on the N.W. slope of Ben Wyvis. Sheet risers are absent on the low-angle slopes of the Ben Wyvis plateau, but appear at angles of 6^{o}-10^{o} on the convex shoulder of the hill. Here they are low (< 40 cm) and straight-fronted, with no frontal lobes. The front of the riser is, however, sometimes aligned obliquely across slope, though at a low angle to the contours. Risers remain low as the slope increases to

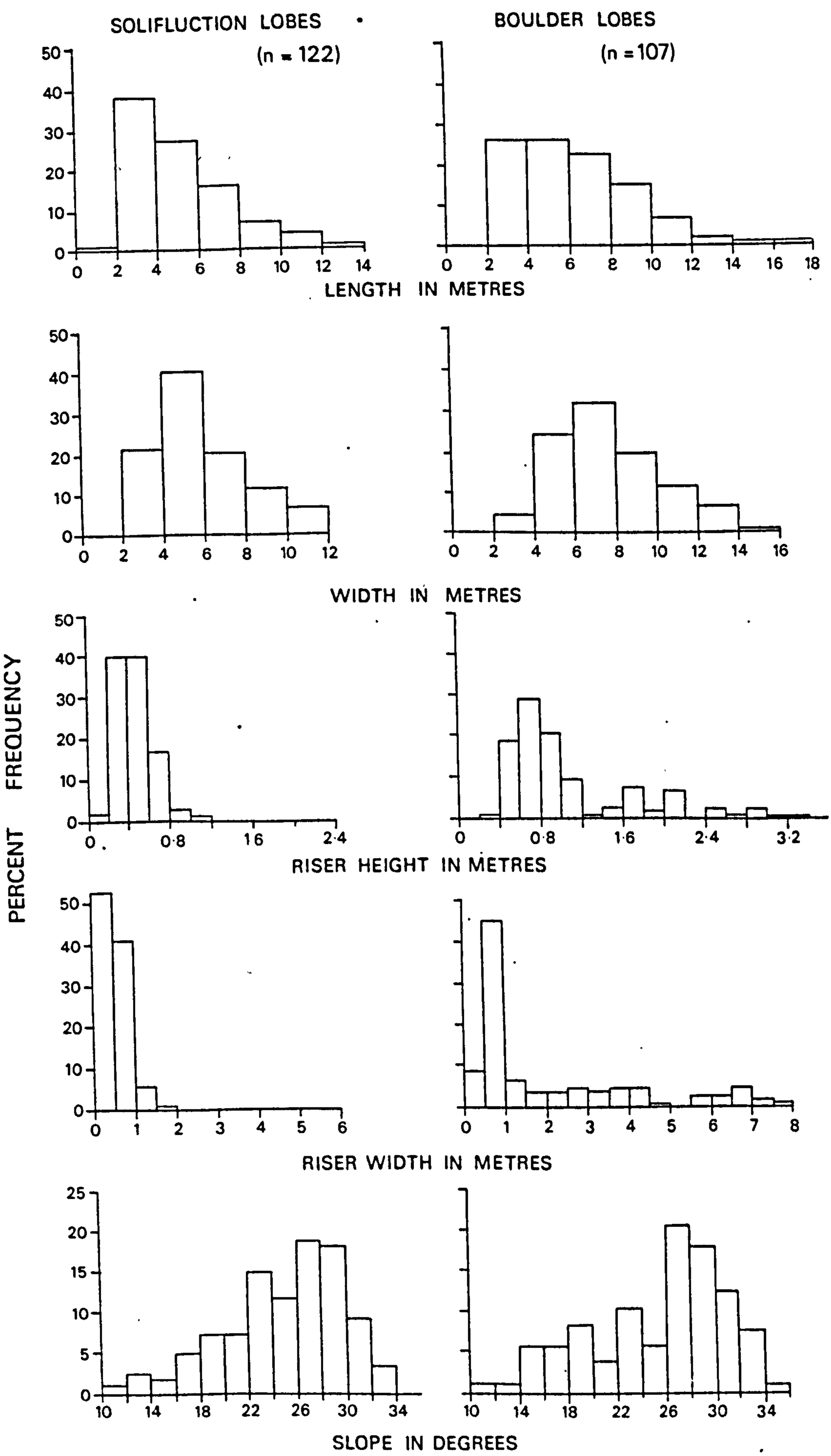


Figure 9.20: Percentage frequency distributions summarizing some morphological characteristics of the solifluction and boulder lobes surveyed in the field. The boulder lobe sample is dominated by the 65 "shallow" turf-banked boulder lobes surveyed on Ben Wyvis.

10°-15°, but small lobes appear at the riser front. As the slope increases beyond 15° well-developed lobate sheets with risers up to one metre in height appear. The obliquity of solifluction sheets appears to increase with slope. At gradients in excess of 25° over-riding of sheets is common, and individual sheets can be traced for only relatively short distances. On some steep parts of the slope the sheet pattern is completely obscured and the hillside gives the appearance of being covered by individual lobes.

Although there is no equivalent transect of boulder features in the three study areas, boulder sheets and lobes in the Cairngorms (e.g. in Coire an Lochain) appear to follow a similar if less regular pattern. In the Northern Highlands, straight-fronted boulder sheets are found on low gradients ($< 10^{\circ}$), the outstanding examples being that identified by Galloway (1961a) as a "continuous turf-banked terrace" on Ben Wyvis (figure 9.3; NH 473682) and a massive boulder sheet in the valley west of An Teallach at NH 040867. On steep slopes the fronts of boulder sheets tend to be less regular than those of solifluction sheets (McMillan, 1978), and lobes are interspersed with boulder streams up to 30 m long. In terms of surface area, the largest solifluction and boulder lobes are often found where there is a relatively sharp reduction in gradient, as at the foot of slopes (figure 9.8).

Lobes have sometimes been represented as possessing a regular U-shaped plan (e.g. King, 1972, figure 2), but even approximately symmetrical features are rare. Figure 9.21 depicts the surveyed plan form of four solifluction lobes and six boulder lobes. Although all the ten examples shown approximate a "U" shape, six have one riser much longer than the other (cf. Shaw, 1977) and in some cases the "U" is flattened or elongated downslope.

Despite such deviations from the "ideal" lobe shape, analysis of the morphometric data reveals relationships between certain dimensions of both solifluction lobes and boulder lobes (table 9.3). For a sample of 118 solifluction lobes on Ben Wyvis (88) and the Fannichs (30), midpoint width (w_2) and mean

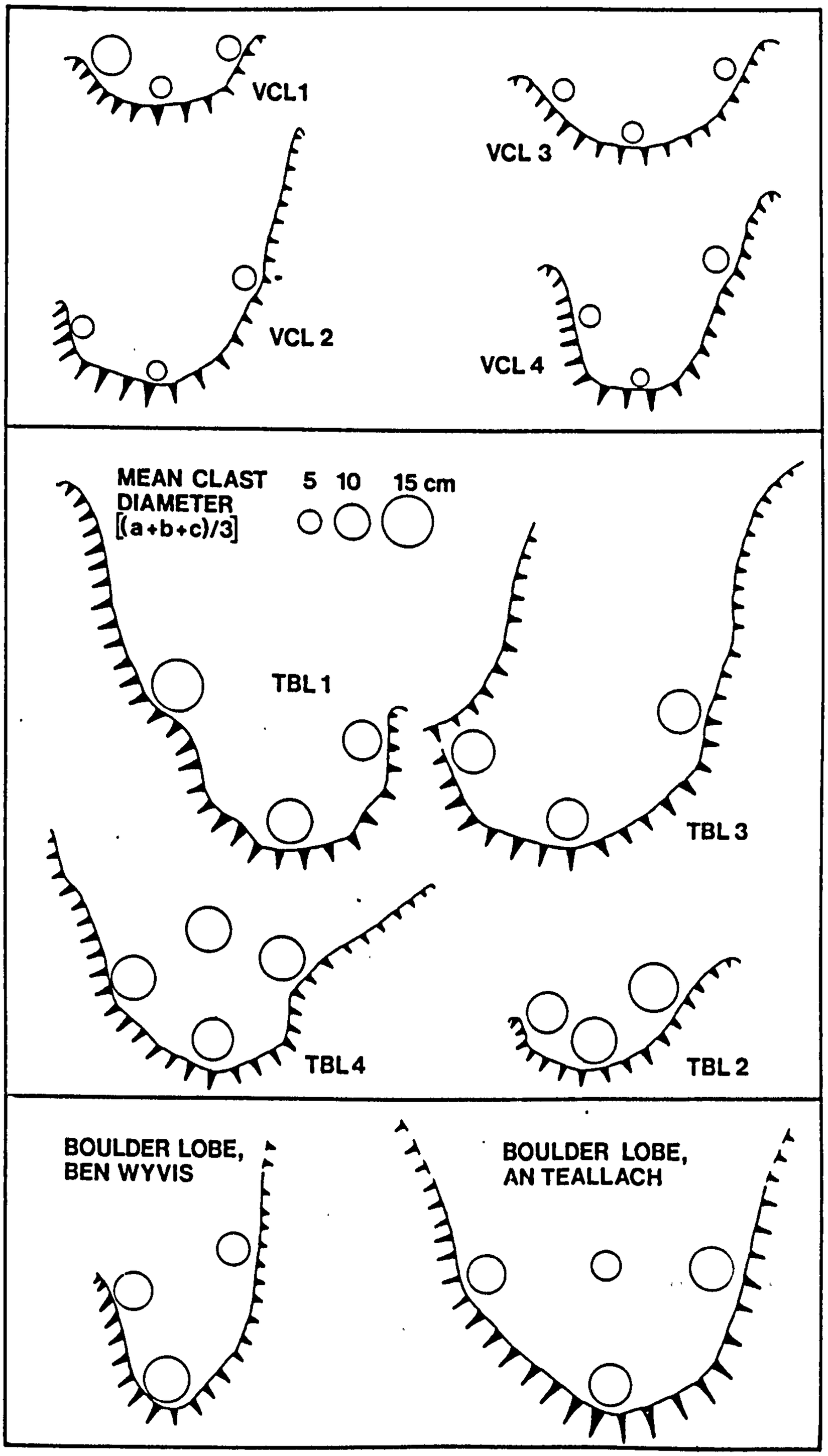


Figure 9.21: Plan form of four solifluction lobes (VCL 1-4) and four "shallow" turf-banked boulder lobes (TBL 1-4) on Ben Wyvis and two massive boulder lobes. Based on tape survey. The circles indicate the mean diameter of 50+ clasts sampled at the locations shown.

Table 9.3

Correlation between selected lobe dimensions and slope, lobe width and riser height

Variable	Slope (α)		Width (w_1)		Riser Height (h_r)	
	S.L.	B.L.	S.L.	B.L.	S.L.	B.L.
Width (w_1)	0	0	-	-	3	0
Midpoint width (w_2)	0	0	5	5	3	0
Mean riser length ($\bar{\ell}$)	1	0	5	5	3	2
Lobe length (ℓ)	1	0	5	5	3	2
Area (A)	1	0	5	5	3	0
Tread angle (α_t)	5	5	0	0	5	0
Riser angle (α_r)	2	0	0	0	0	0
Riser width (w_r)	0	0	2	0	5	5
Riser height (h_r)	5	0	3	0	-	-

S.L. - solifluction lobes on the Fannichs and Ben Wyvis (n = 118)

B.L. - shallow turf-banked boulder lobes on Ben Wyvis (n = 65)

Numbers refer to significance level of correlation:

0 - no significant relationship; 1 - significant at the .05 level;

2 - significant at the .01 level; 3 - significant at the .001 level;

4 - significant at the .00001 level. All correlations are positive.

riser length ($\bar{\ell}$) are linearly related to width (w_1) as

$$w_2 = 0.598\ w_1 + 0.036 \quad (r^2 = 0.783;\ \ p < .00001) \qquad (9.3)$$

and $$\bar{\ell} = 0.717\ w_1 + 0.580 \quad (r^2 = 0.378;\ \ p < .00001) \qquad (9.4)$$

the equivalent relationships for a sample of 65 turf-banked boulder lobes on Ben Wyvis are similar:

$$w_2 = 0.640\ w_1 + 0.123 \quad (r^2 = 0.701;\ \ p < .00001) \qquad (9.5)$$

and $$\bar{\ell} = 0.443\ w_1 + 1.345 \quad (r^2 = 0.265;\ \ p < .00001) \qquad (9.6)$$

although the smaller coefficient of w_1 in equation (9.6) compared with that in equation (9.4) indicates that solifluction lobes are generally more elongate than boulder lobes; particularly when w_1 is large.

For the solifluction lobe sample, width, midpoint width and mean riser length all display a positive linear relationship with riser height, significant at the .001 level. This indicates that, in general, the areal dimensions of these lobes are related to lobe thickness. Between riser width (w_r) and riser height (h_r) on solifluction lobes there is an approximately 1:1 relationship:

$$w_r = 1.004\ h_r + 0.076 \quad (r^2 = 0.311;\ \ p < .00001) \qquad (9.7)$$

implying that these lobes tend to have riser angles around 45°.

For the boulder lobe sample, $\bar{\ell}$ is weakly correlated with h_r ($r^2 = 0.114$), but there is no significant relationship between the width variables and riser height. The relationship between w_r and h_r, given by the regression

$$w_r = 0.840\ h_r + 0.018 \quad (r^2 = 0.245;\ \ p < .00001) \qquad (9.8)$$

implies generally steep riser angles (around 50°), but this is not typical of other samples of boulder lobes (table 9.2), which usually have riser angles between 20° and 30°. This may be related to the fact that the shallow turf-banked boulder lobes of Ben Wyvis contain a much higher proportion of fines than other more massive boulder lobes on this mountain or An Teallach (see below).

9.2.3.4 Discussion

From the above it is apparent that boulder lobes and solifluction lobes, differentiated in terms of internal constitution, are similar in some other characteristics but different in size. Although vegetation cover is not an infallible distinguishing factor, most vegetation-covered features in the study area are of the solifluction type and most turf-banked (and all stone-banked) features are of the boulder type. Both types occupy a similar range of slopes and show a downslope transition, as slope angle increases, from sheets via lobate sheets to lobes. The plan dimensions ($\bar{l}$, w_1, w_2) of both types are related. Solifluction lobes, however, are generally shorter, narrower and less high, with steeper risers. Descriptions by other workers of the dimensions of the two types suggest that this distinction is general. The heights of boulder lobes elsewhere in Great Britain have been given as 1-4 m (Metcalfe, 1950), 1-5 m (Galloway, 1958), "up to 5 m" (King, 1968), 1-3 m (Ball and Goodier, 1970) and 0.2-5.9 m (Shaw, 1977). In contrast, reports of solifluction lobes in Britain describe features generally less than 1 m in height (Tivy, 1962; Ball and Goodier, 1970; Kelletat, 1970a). Indeed, a survey of the literature on solifluction lobes elsewhere indicates that although these may reach 2 m in height (Williams, 1957; Price, 1970; Harris, 1972, 1973) and occasionally more (Benedict, 1970; Price, 1972; Washburn, 1973, figure 5.14), they are often less than 1 m high (Sharp, 1942a; Dutkiewicz, 1962; Kallander, 1967; Washburn, 1967). The lobes measured in Norway by the author ranged from 0.4 to 1.8 m in height. Comparison of other dimensions is less easy as lobe width and length have been measured in different ways by different workers, but it is notable that mean figures given for boulder lobe width by King (1968: 13.6 m) and Shaw (1977: 14.8 m on granite, 6.3 m on metamorphics) suggest that, whilst the solifluction lobes measured in the present survey are generally narrower (mean = 5.85 m), there is wide variation in this dimension of boulder lobes, depending on size of constituent material.

In summary, morphometric data from the study area support the hypothesis stated in section 9.2.1, namely that there are two distinct populations of lobes and sheets. The following sections explore the distributional and sedimentological characteristics of each population in an attempt to establish the causes underlying the differences.

9.2.4 Distribution and environmental relationships

The distribution of both types of sheet and lobe may be considered to reflect six possible influences: nature of regolith, slope, the limits of former glaciers, the extent of peat cover, aspect and altitude. The final two factors are effectively measures of climatic control, the former broadly reflecting the influence of insolation, snow-lie and dominant winds, the latter of temperature and precipitation. The relationships between these factors are complex, however: the altitude to which sheets and lobes descend, for example, may reflect climatic controls but is more often determined by a change in lithology or slope, or the upper limits of peat or former glaciers (Shaw, 1977). For this reason it is possible only to hazard generalizations concerning the relative importance of each of the six factors, and to suggest approximate environmental thresholds.

9.2.4.1 Regolith

In chapter 6 it was demonstrated that the nature of upland regolith reflects the relative effectiveness of macrogelivation and ~~microgelivation~~ granular disintegration on underlying rock type. Not surprisingly, the distribution of boulder sheets and lobes is related to that of lithologies on which macrogelivation has proved dominant, producing blockslopes on which fine material is absent at the surface. This is beautifully illustrated in the Fannichs, where boulder lobes are largely restricted to areas underlain by Lewisian Gneiss (e.g. west of Sgùrr nan Clach Geala) or Moinian quartzose schist (e.g. around Beinn Liath Mhór). In contrast, the micaceous rocks of the western Fannichs and the Sgùrr nan Clach Geala and Sgùrr Mór areas of the eastern Fannichs (figure 4.3)

have yielded a regolith rich in fines and dominated by vegetation-covered solifluction lobes and sheets. This dichotomy extends to Ben Wyvis, where turf-banked boulder sheets and lobes tend to occupy areas of siliceous schist but vegetation-covered solifluction lobes dominate the micaceous pelites (Galloway, 1958, 1961a). On An Teallach, rather poorly-developed boulder lobes are found on the quartzite blockslopes (e.g. NH 081840) but are rare on Torridon Sandstone, being restricted to a steep debris slope around NH 059841. Solifluction features are also rare on An Teallach, being best developed immediately downslope of the aforementioned boulder lobes.

The correspondence between type and lithology is therefore not perfect, as features composed of predominantly coarse detritus are found on rocks that normally yield an abundance of fines on weathering. Solifluction features nonetheless appear to be limited in distribution to areas where the regolith contains a high proportion of fine (< 2 mm diameter) detritus. Not all such areas support solifluction features, however, as testified by their rarity on Torridon Sandstone terrain.

9.2.4.2 Slope

The morphometric data presented in section 9.2.3.2 indicate that solifluction and boulder lobes occupy a similar range of slopes (10^{o}-36^{o}). This range is virtually identical to that found by King (1968, 1972) and Shaw (1977) for boulder lobes in the Cairngorms (10^{o}-35^{o}) and Grampians (10^{o}-34^{o}) respectively. Galloway (1958, 1961a) stated that solifluction and boulder lobes on Ben Wyvis occupy slopes of 5^{o}-20^{o}, but as the modal range for both types on this hill exceeds 25^{o} it is clear that his figures reflect undermeasurement

Sheets without crenulate risers are found on lower gradients than lobes. Solifluction sheets on Ben Wyvis occur on gradients of 6^{o}, and an extensive boulder sheet at NH 473682 on the same hill (figure 9.3) is developed on slopes of 7^{o}-11^{o}. At slopes over 36^{o} rock frequently crops out through the regolith (chapter 8) and no lobes or sheets are found. The limiting angles for both

solifluction and boulder features would therefore appear to be 6^{o} and 36^{o}.

9.2.4.3 Limits of former glaciers

The absence of boulder or "large-scale solifluction" features from ground occupied by glaciers during the Loch Lomond Stadial has been established by Sissons (1972, 1974a, 1976a, 1977a, 1979b; Sissons and Grant, 1972; Sissons et al., 1973). The relationship between the limits of former glaciers and boulder sheets and lobes in the study areas is not clearly defined, however, as the upper margins of former glaciers cannot be determined by independent evidence (drift limits or lateral moraines). Nonetheless, there are no boulder sheets or lobes in areas that were apparently ice-covered during the Loch Lomond Stadial, and coarse detritus within the limits of the Loch Lomond Advance glaciers shows no sign of mass-movement. Sissons' inference that boulder lobes have been inactive throughout the Flandrian appears justified.

At several localities, such as the Gaick Plateau (Kelletat, 1970a) and above Strath Dionard (Sissons, 1977a) solifluction lobes are known to occur within the limits of Loch Lomond Advance glaciers, indicating formation since the disappearance of glacier ice. The only instances of sheets or lobes within the Loch Lomond Advance limits in the study area are the debris lobes at the base of a talus sheet at NH 217707 (map 4) and vegetation-covered solifluction lobes at NH 061845 on An Teallach. In both cases the lobes are poorly-formed, and indeed in the latter case the glacial limit is contentious (chapter 4). It seems reasonable to conclude that although solifluction sheets and lobes occasionally occur on ground previously covered by Loch Lomond Advance glaciers, they tend to be more common on slopes that escaped glaciation during the Loch Lomond Stadial. In the Fannichs, the paucity of solifluction features inside the limits of former local glaciers probably reflects the coarseness (or lack) of detritus within the glacial limit compared with neighbouring ridges on which frost-weathered fines are abundant.

9.2.4.4 Peat

Shaw (1977) suggested that the true downslope limit of boulder lobes in the Lochnagar area has locally been obscured by peat growth, but there are no obvious examples of this in the study areas. It is also possible that peat has inhibited recent (i.e. late Flandrian) mass-movement by absorbing water, insulating the underlying regolith against freezing and providing a stable, cohesive surface layer of relatively low bulk density. At several points in the Fannichs (e.g. NH 160708, NH 184723, NH 198700) the lower limits of solifluction features and the upper limits of peat coincide, suggesting that peat has prevented the downhill extension of such features. On the N.W. slope of Ben Wyvis the soils below 700 m (the approximate lower limit of solifluction features) support a thin peat cover that increases in thickness downslope. This may account for the uniform downslope termination of such features.

9.2.4.5 Aspect

Various researchers have proposed that boulder lobes are most common on slopes with a westerly aspect. Those in the Cairngorms have been considered to be preferentially developed on north-facing or west-facing slopes (Watt and Jones, 1948; Galloway, 1958; King, 1968) and Shaw (1977) detected a preference for west-facing slopes amongst the granite lobes of Lochnagar, whilst Sissons (1980) considered those in the Lake District to be most extensively developed on south-facing or west-facing slopes. A general preference for westerly aspects is not proven, however: Shaw also found that metamorphic lobes tend to face between east and south, although he considered the sample analysed (50) too small to be representative.

Examination of maps 4-6, on which turf-banked boulder sheets and lobes are well represented, shows that although these occur at all aspects the great majority face between 180° and 360°. This bias, however, reflects a similar bias in the aspects of suitable slopes (i.e. slopes of 10°-36° above 750 m, underlain

by siliceous rocks and outside the limits of Loch Lomond Advance glaciers), so that no inferences can be made with regard to the aspects of boulder sheets and lobes in these areas.

Solifluction sheets and lobes within the study areas display no directional bias. Those surrounding the summits of A'Chailleach and Sgùrr Breac (map 3), for example, face all directions, as do those around the summit of Ben Wyvis. Aspect does not appear to affect the distribution of solifluction features.

9.2.4.6 Altitude

As Shaw (1977) pointed out, the altitude to which lobes and sheets descend often reflects a change in terrain (slope, lithology, peat cover or glacial limits) rather than climatic control, so that such data must be interpreted with caution. The altitudes of the lowest boulder features in the study areas rise eastwards from 420 m (a boulder sheet west of An Teallach at NH 040867) to 590 m on Càrn Gorm, Ben Wyvis. The limited data available from elsewhere suggest that this eastward rise in the lower limit of boulder features is typical. Peacock (personal communication) has reported boulder lobes at 300 m in the Outer Hebrides, and those on Rhum descend to 350 m (map 8). In contrast, the lowest boulder lobes in the Cairngorms occur at 550 m (King, 1968) and those on Lochnagar are not found below 580 m (Shaw, 1977). Farther south the lower limit rises to around 700 m in the Southern Uplands (unpublished data) and 800 m in Snowdonia (Ball and Goodier, 1970). Overall these figures indicate a general northward and westward decline in the lower limit of boulder sheets and lobes.

It is tempting to relate this decline to a parallel decline in the altitude of blockslopes (section 8.3.2) on which boulder features are commonly developed. This in turn may reflect increased exposure, in that removal of snow by high winds permits deeper frost penetration and hence more effective macrogelivation and cryoturbation. Such an interpretation must remain speculative, however, until more data are available on the lower limits of

blockslopes and boulder features, and the relationship between the two.

The lower limit of solifluction features in the Scottish Highlands has been studied by Kelletat (1970a). This also appears to decline northwards and westwards. It is highest in the Cairngorms (760 m) and lies around 730 m on other mountains in the central and eastern Highlands (e.g. Glas Maol, the Drumochter Hills and Ben Lawers) but declines westwards to 580 m on Ben Nevis and Skye and 550 m on Mull, and northwards to 580 m on Ben Klibreck (Sutherland). The figures obtained from maps 2-6 agree well with this pattern (An Teallach, 670 m; Fannichs c. 650 m; Ben Wyvis c. 650 m) although the An Teallach figure probably reflects topography; the potential lower limit on this hill may well be lower, possibly around the 610 m figure that Kelletat obtained for Applecross. This pattern suggests an inverse relationship between the lower limit of solifluction features and present precipitation (chapter 2), but again exposure may represent a major control.

9.2.4.7 Environmental relationships

Some of the six factors discussed above influence not only the distribution of sheets and lobes, but also their size and shape. Regression analysis of the morphometric data for 118 solifluction lobes on Ben Wyvis and the Fannichs shows that these tend to increase in thickness (h_r) with increasing slope (α):

$$h_r = 0.227\ e^{0.29\alpha} \quad (r^2 = 0.203;\ p < .0001) \qquad (9.9)$$

Other relationships with slope, although statistically significant, are weak (table 9.3): lobe length tends to increase with slope (to be expected in view of the slope-determined transition from sheets to lobes) and riser angle to steepen. None of these relationships holds for the sample of 65 boulder lobes, however. For a large sample of boulder lobes on Lochnagar Shaw (1977) also did not find any relationships between morphometric variables and slope, except that riser angle was positively correlated with slope angle.

The overall size of boulder lobes appears, however, to reflect size of constituent material (Shaw, 1977) and hence lithology. Table 9.2 shows that the mean thickness (h_r) of boulder lobes on An Teallach is more than twice that for similar features on Ben Wyvis. This appears to reflect the generally coarser material inside the An Teallach lobes (see below). Clastsize does not, however, influence the average dimensions of solifluction lobes; these are similar in each area.

No meaningful relationships were found between lobe morphology and aspect or altitude.

9.2.4.8 Summary

The distributional characteristics of solifluction and boulder features in the study areas may be summarised as follows.

(i) Lithology or, more specifically, the response of different rocks to periglacial weathering provides a fundamental control on the distribution of both types of feature. Solifluction features only occur where the regolith is rich in fines, but boulder features are most common on blockslopes.

(ii) Both types of feature are limited to slopes between 6^{o} and 36^{o}.

(iii) Boulder features are not found within the limits of the Loch Lomond Advance glaciers. Solifluction features are found within these limits, but only rarely.

(iv) Peat appears to inhibit the formation of solifluction sheets and lobes.

(v) Solifluction features display no directional preferences, and there is no evidence for such a preference in the distribution of boulder features.

(vi) The lowest altitudes to which solifluction and boulder features descend in the study areas form part of a general

westward and northward decline in such limits. In general, the lowest altitude reached by boulder features in any area is several tens of metres below that of solifluction features. Locally, however, solifluction features are encountered downslope of boulder features (e.g. on An Teallach and the N.W. slope of Ben Wyvis).

(vii) The thickness of solifluction lobes is related positively to slope, whereas that of boulder lobes reflects size of constituent detritus.

9.2.4.9 Discussion

It was concluded in section 9.2.3.4 that sheets and lobes distinguished on the basis of internal constitution are similar in configuration but different in size. The above summary highlights other similarities and differences. Perhaps the most significant difference lies in the relationship of each type to the limits of the Loch Lomond Advance glaciers, which implies that boulder features have not been active in Flandrian times but that some solifluction features at least have formed and moved downslope since final deglaciation. Support is given to this interpretation by the mid and late Flandrian radiocarbon dates obtained by O'Brien (Sugden, 1971) and Mottershead (1978) on organic material under solifluction features on the Cairngorms and Ben Arkle, and by the measurements of mass-movement reported below. These show that solifluction lobes on Ben Wyvis are moving at the present day but boulder lobes on the same slope and on An Teallach are apparently immobile. It is therefore possible to interpret boulder features as the Lateglacial ancestors of Flandrian solifluction features, produced by the operation of similar processes but under a much more severe periglacial climate, and subsequently immobilised through the eluviation of fines (Galloway, 1958; Kelletat, 1970b). Others, particularly Shaw (1977) have concluded that although the boulder lobes of Lochnagar are of Lateglacial age, they moved downslope under a completely different mechanism from that responsible for the

movement of solifluction lobes. The results of sedimentological analyses designed to resolve this question and to throw light on the processes involved in movement are reported in the following section.

9.2.5 Sedimentology

9.2.5.1 Structure

The internal structures of solifluction and boulder lobes are illustrated in figure 9.22. The sections through the solifluction lobes on Ben Wyvis and the Fannichs (1-3) reveal almost identical structures. In each case the body of the lobe consists of a homogeneous structureless brown soil containing plant roots and many angular stones (figures 9.23 and 9.24). The podzolic soil horizons downslope of the risers of these lobes may be traced upslope under the brown soil layer. In the case of the Ben Wyvis examples (1 and 2), however, these buried soil horizons are much fragmented and overlie a concentration of boulders (figure 9.24), the significance of which is discussed below. The profiles of these three lobes closely resemble that of a lobe excavated in Norway (5), although the latter is composed of soliflucted till and contains layers of organic material above the main buried soil horizon. The An Teallach solifluction lobe (4) is remarkable in that the body of the lobe is composed entirely of structureless sand containing no stones whatsoever. Again, however, a poorly-preserved illuviated humic horizon can be traced under the lobe from the soil downslope of the riser.

The sections excavated through the boulder lobes (6-9) revealed an entirely different structure. The uppermost 50-100 cm of all three excavated boulder lobes consisted essentially of openwork boulders, cobbles and pebbles, although the interstices between clasts often contained loose fines. A striking feature of this zone is the downward decrease in clast size (figures 9.25 and 9.26). Immediately below this zone of vertical sorting clasts of all sizes are embedded in a predominantly sandy matrix that grades into a stony C horizon. In one of the Ben Wyvis examples (6 and 9), however, the boundary between the body of the lobe and

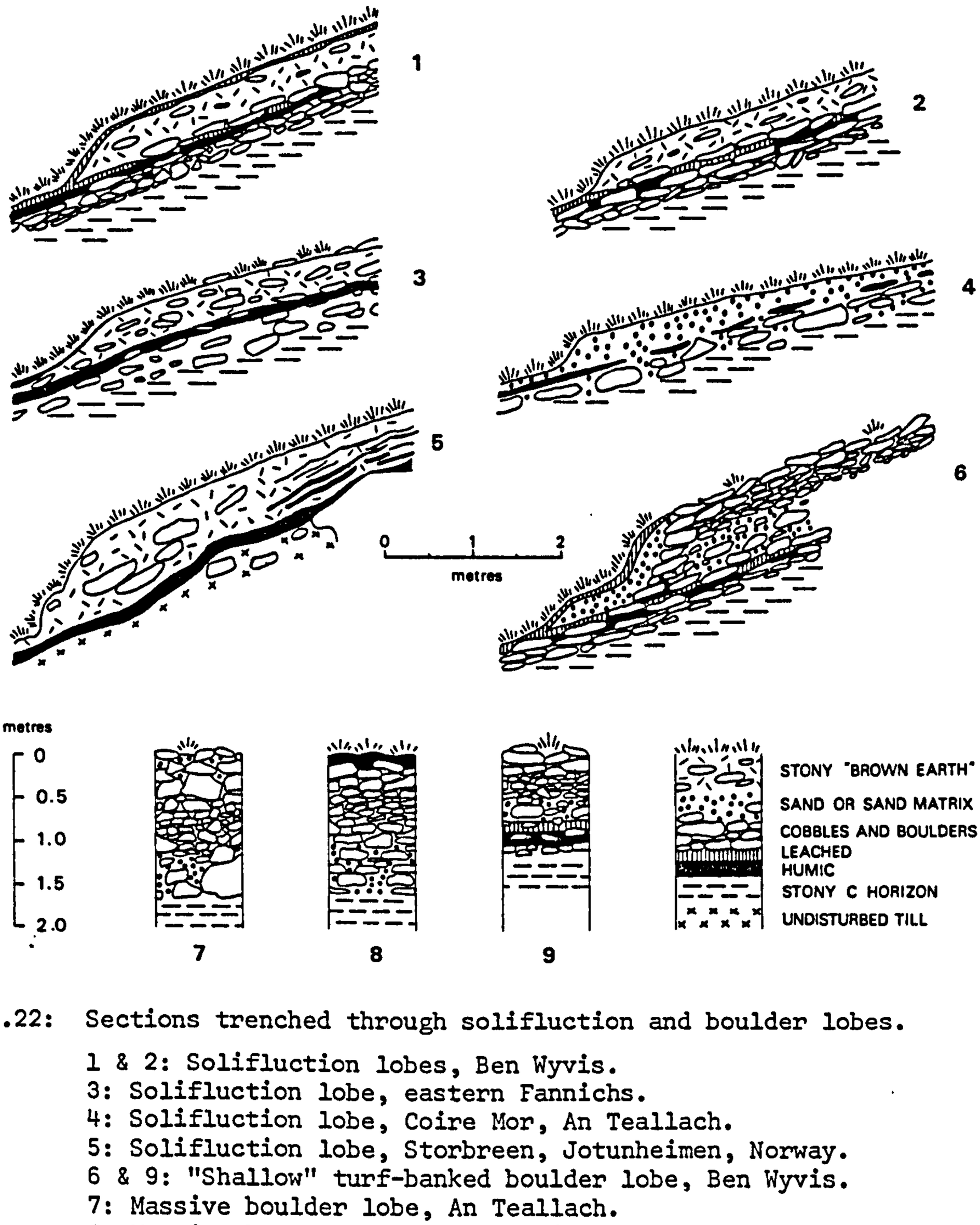


Figure 9.22: Sections trenched through solifluction and boulder lobes.

1 & 2: Solifluction lobes, Ben Wyvis.
3: Solifluction lobe, eastern Fannichs.
4: Solifluction lobe, Coire Mor, An Teallach.
5: Solifluction lobe, Storbreen, Jotunheimen, Norway.
6 & 9: "Shallow" turf-banked boulder lobe, Ben Wyvis.
7: Massive boulder lobe, An Teallach.
8: Massive boulder lobe, Ben Wyvis.



Figure 9.23: Rear of the trench cut into a solifluction lobe at c. 800 m altitude south of Meall a'Chrasgaidh in the eastern Fannichs. The body of the lobe comprises a brown, structure-less soil that overlies a buried organic soil horizon.



Figure 9.24: Rear of the trench cut into a solifluction lobe (2, figure 9.22) at c. 800 m altitude on the N.W. flank of Ben Wyvis. The body of the lobe consists of a structureless brown soil that overlies a boulder concentration typical of the upper layers of "shallow" turf-banked boulder lobes on the same slope, suggesting that the solifluction lobe has over-ridden a vertically-sorted boulder sheet (compare with figure 9.25).



Figure 9.25: Rear of the trench cut into a "shallow" turf-banked boulder lobe at c. 800 m altitude on the N.W. slope of Ben Wyvis. (6 and 9, figure 9.22). The uppermost openwork zone is not shown, but a downward diminution in clast size is evident, indicative of vertical sorting.

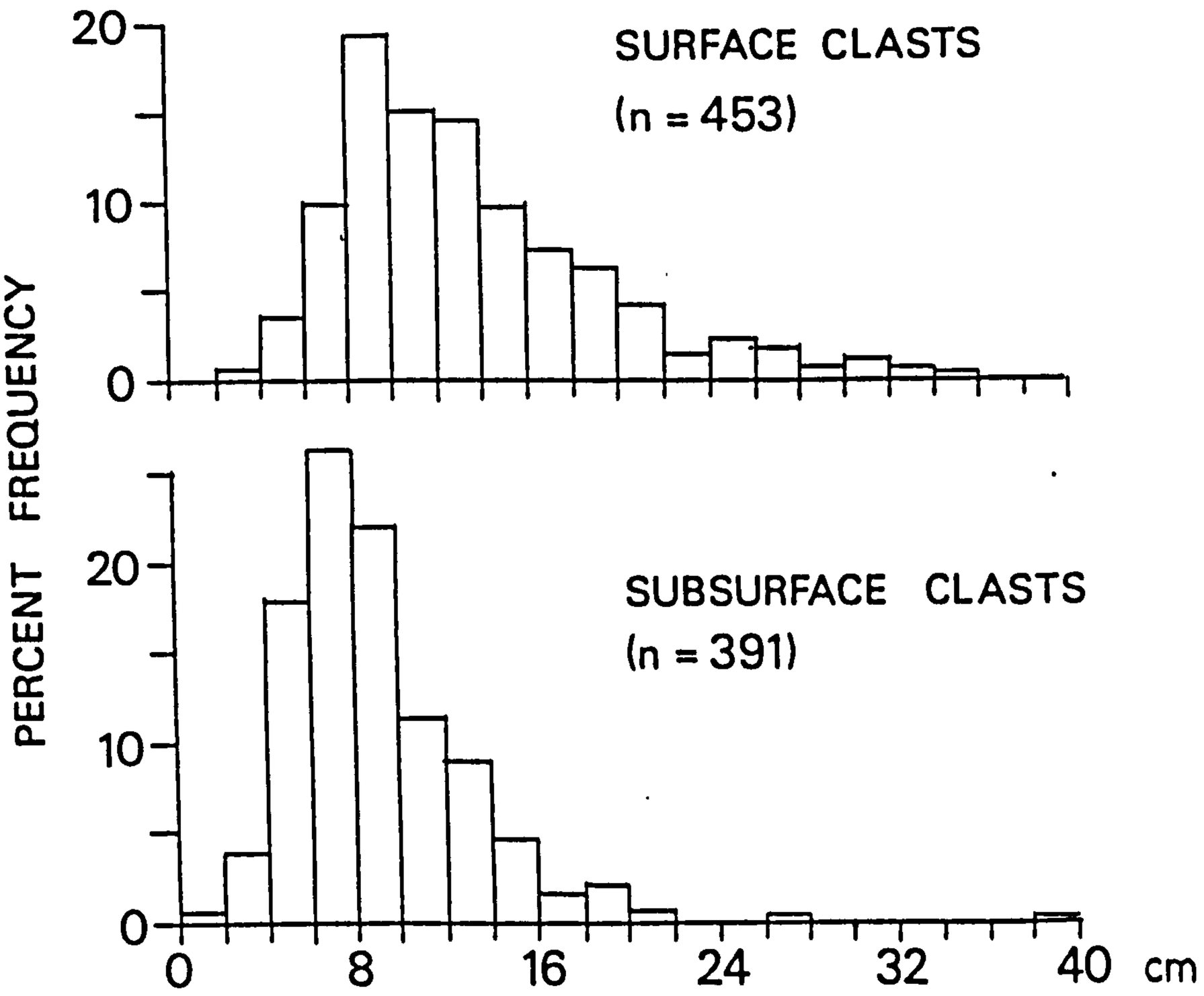


Figure 9.26: Frequency distributions of intermediate clast length for samples of clasts measured on the surface and immediate subsurface of four "shallow" turf-banked boulder lobes at c. 800 m altitude on the N.W. flank of Ben Wyvis.

the underlying C horizon is marked by what appears to be a buried podzolic soil, although this horizon is largely fragmented.

Two further points are noteworthy. First, the sections through boulder lobes are clearly similar in structure to those excavated in blockslopes (figure 8.9, sections 3 and 6). Secondly, examination of the stone layer under the solifluction lobe shown in figure 9.24 showed that it is very similar in structure to that of the boulder lobe section in figure 9.25. Both occur on the N.W. slope of Ben Wyvis, the former some distance upslope of the boulder lobe zone. This indicates that solifluction sheets and lobes have over-ridden boulder features in this area, again suggesting that the former have been active in more recent times than the latter.

9.2.5.2 Fabric

Fabrics measured at the front and both sides of four solifluction lobes on the N.W. slope of Ben Wyvis generally show a strong preferred downslope orientation and dip (figure 9.27). The former, assessed using the A_n 180 statistic, is significant at the .05 level in all cases but one. Dip direction shows an even stronger downslope bias, giving A_n 360 scores that are in all cases significant at the .005 level. The orientation diagrams indicate that reversed (upslope) imbrication and cross-slope alignment of clasts are rare. Comparison of the three orientation diagrams for each lobe reveals evidence of divergence at the lobe margins, with clasts on the left-hand (northern) riser trending slightly to the north of slope aspect and those on the right-hand (southern) riser trending slightly to the south of slope aspect.

The fabrics of four shallow boulder lobes on the same hillslope (figure 9.28) also display preferred downslope orientation and dip, but the dispersion of the distributions about the modal orientation range is greater. Several diagrams also display prominent secondary modes at 90° or 180° to the principal mode, indicating cross-slope orientation and reverse imbrication respectively. These characteristics are also

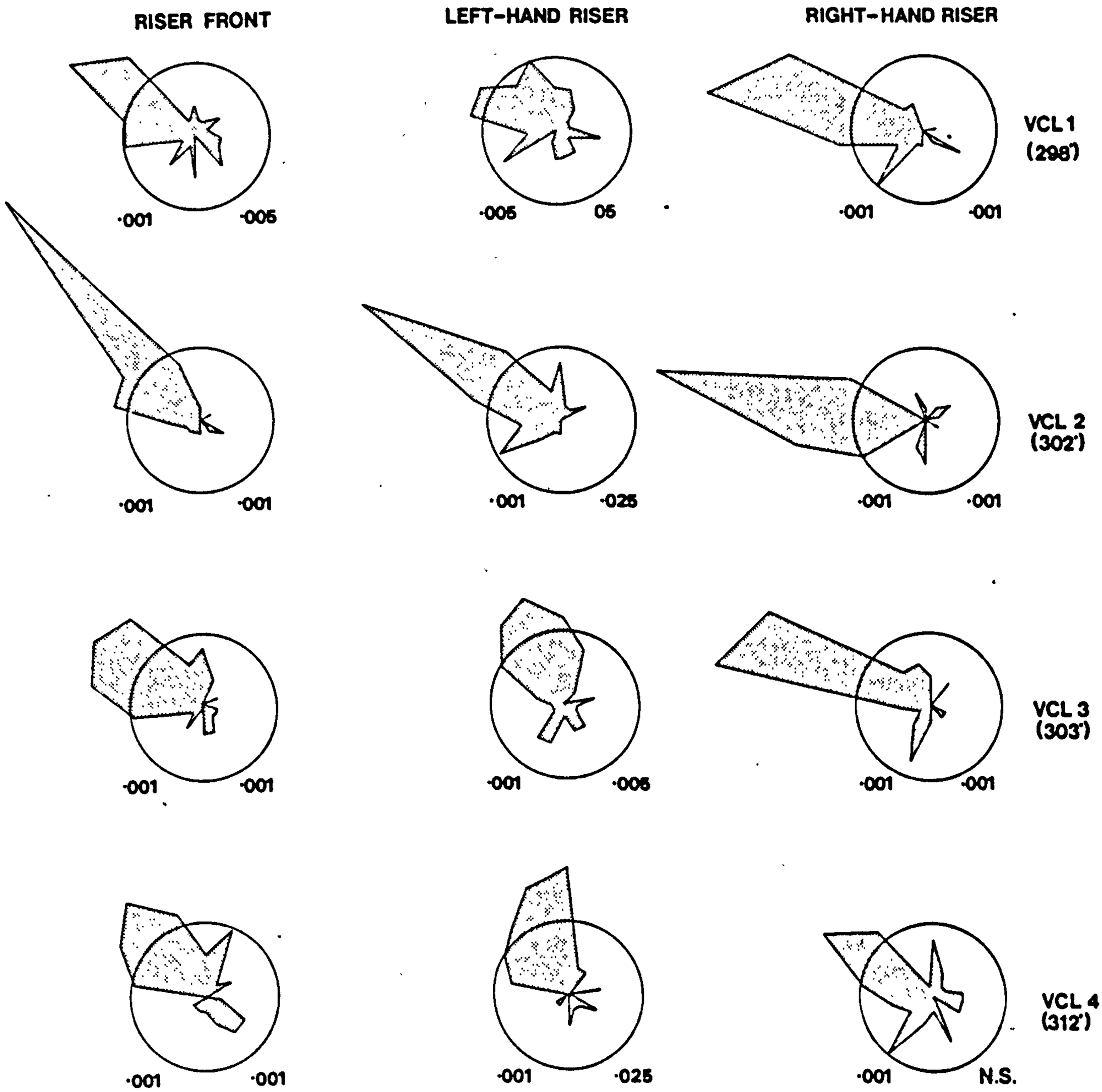


Figure 9.27: Fabrics measured on four solifluction lobes on the N.W. flank of Ben Wyvis. N.S. = not significant.

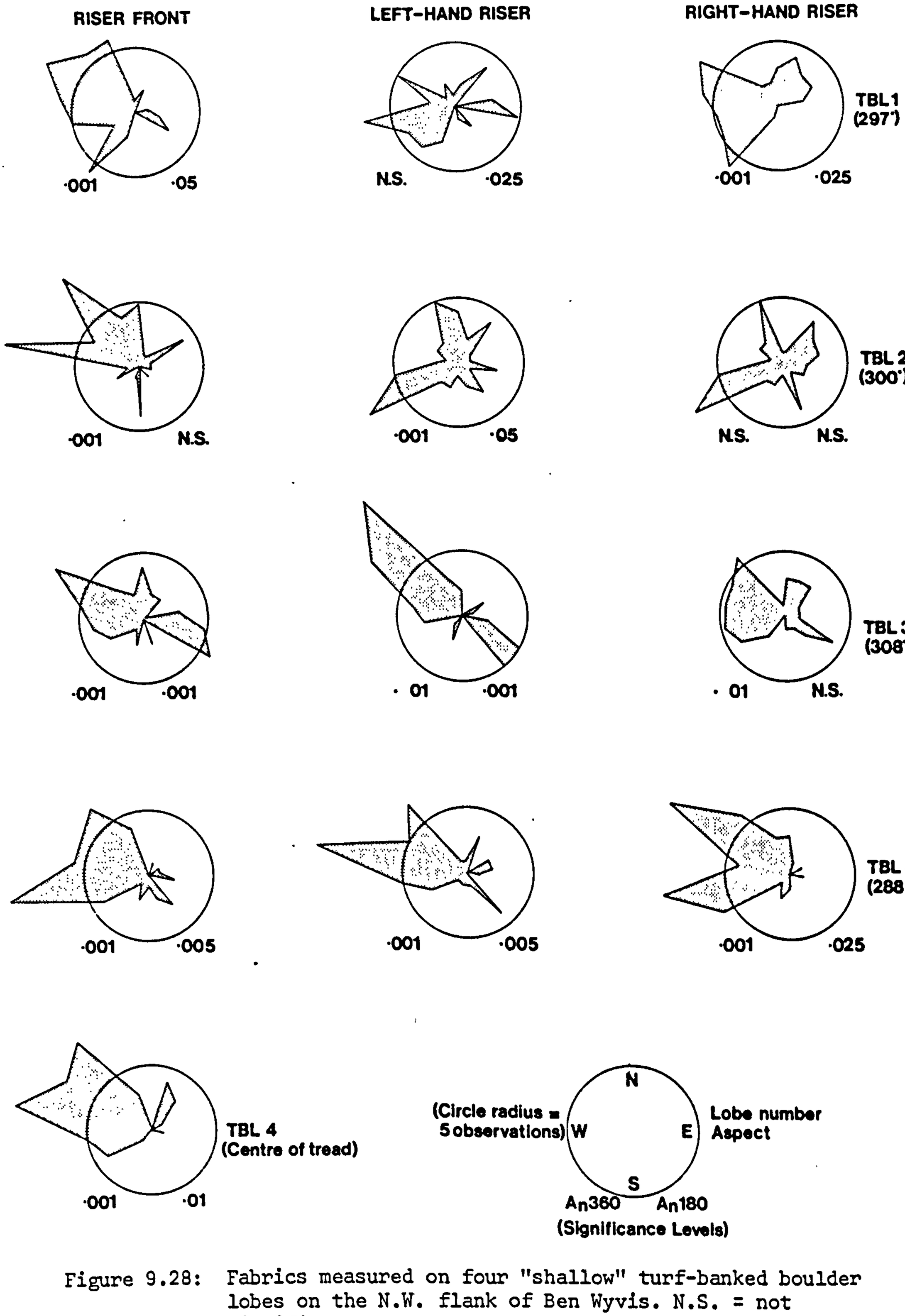


Figure 9.28: Fabrics measured on four "shallow" turf-banked boulder lobes on the N.W. flank of Ben Wyvis. N.S. = not significant.

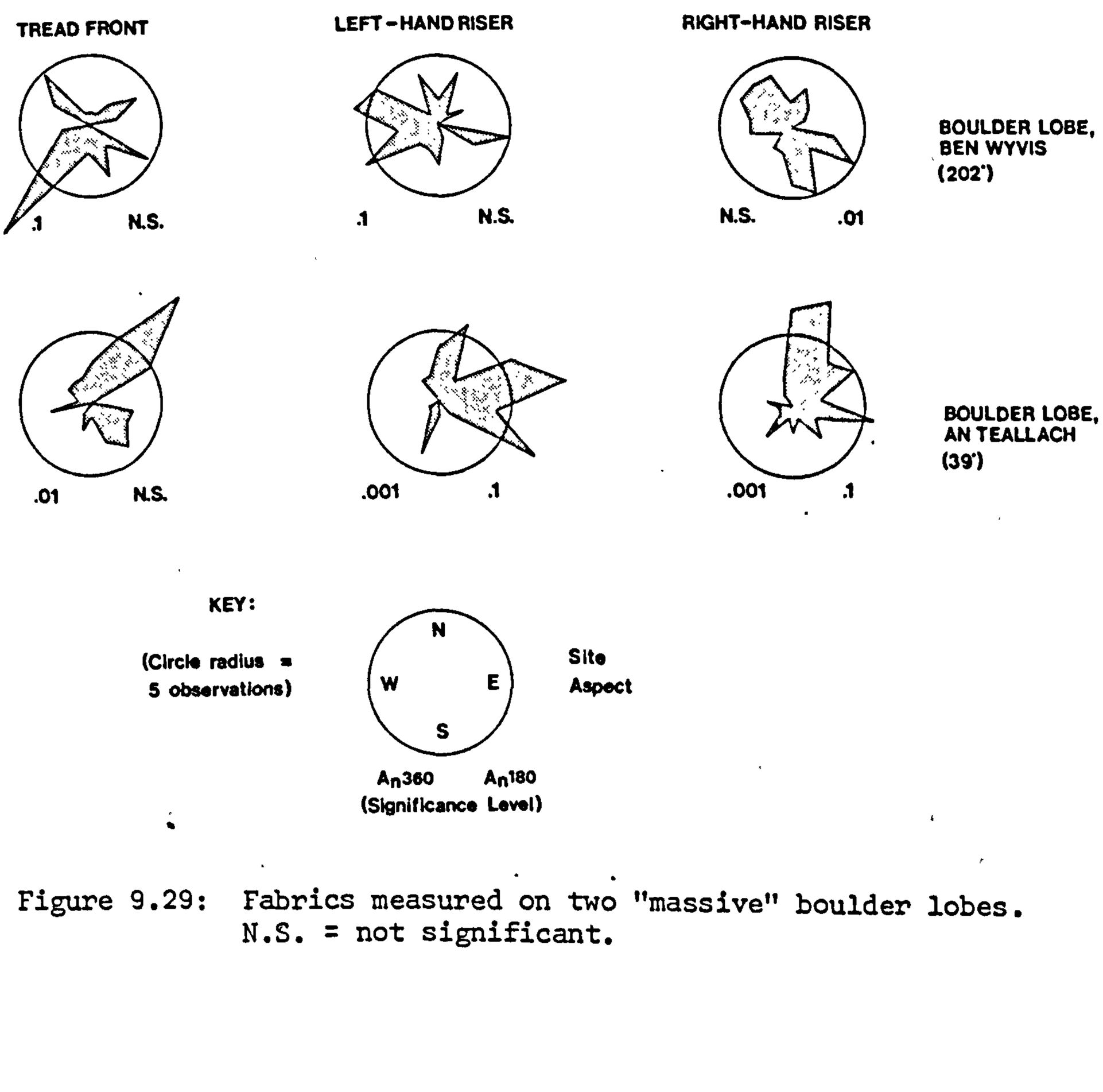


Figure 9.29: Fabrics measured on two "massive" boulder lobes. N.S. = not significant.

typical of fabrics measured at the margins of massive boulder lobes on another slope on Ben Wyvis (NH 482665) and on An Teallach (NH 060841) (figure 9.29). In consequence, the A_n statistic scores for boulder lobe fabrics tend to fall below those for solifluction fabrics (figure 9.30), although there is overlap between the two. Of the 19 boulder lobe fabrics, only 11 display a downslope orientation significant at the .05 level (A_n 180) and only 10 have a preferred downslope dip significant at the .005 level (A_n 360).

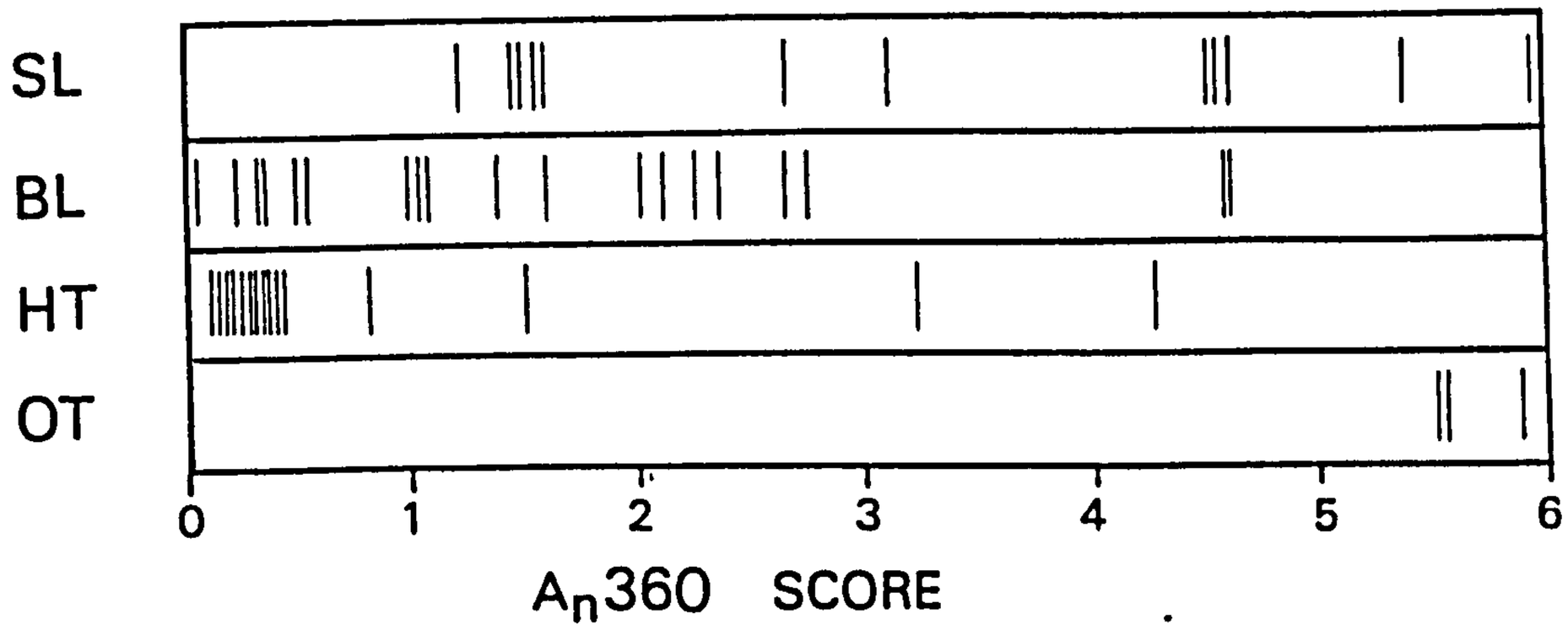
9.2.5.3 Clast size

Figure 9.21 illustrated considerable differences in mean clast size between solifluction lobes and boulder lobes. As the boulder lobe samples were taken from the surface or immediate subsurface, however, the difference between the two types appears greater than it would have been had the boulder lobe samples been taken across a range of depths. Even so, the diagrams highlight a difference in size that is conspicuous in the field, even though the excavated solifluction lobes contained occasional large (up to 70 cm in length) boulders.

Another feature revealed by figure 9.21 is the homogeneity of mean clast size over the surface of individual lobes, particularly boulder lobes. There are some exceptions, however, the margins of some boulder lobes being composed of much larger clasts than the treads. Fifty clasts sampled from the margins of the boulder lobe shown in figure 9.11, for example, had a mean intermediate axis length of 25.1 cm, but the mean for 50 stones from the surface of the tread was only 12.1 cm. The equivalent figures for a small quartzite lobe on Glas Mheall Liath, An Teallach, were 21.8 cm and 7.8 cm. Such figures suggest that lateral as well as vertical sorting has operated on some boulder lobes.

9.2.5.4 Grain size

The grain size characteristics of samples of fine material taken from solifluction lobes and the lower layers of boulder



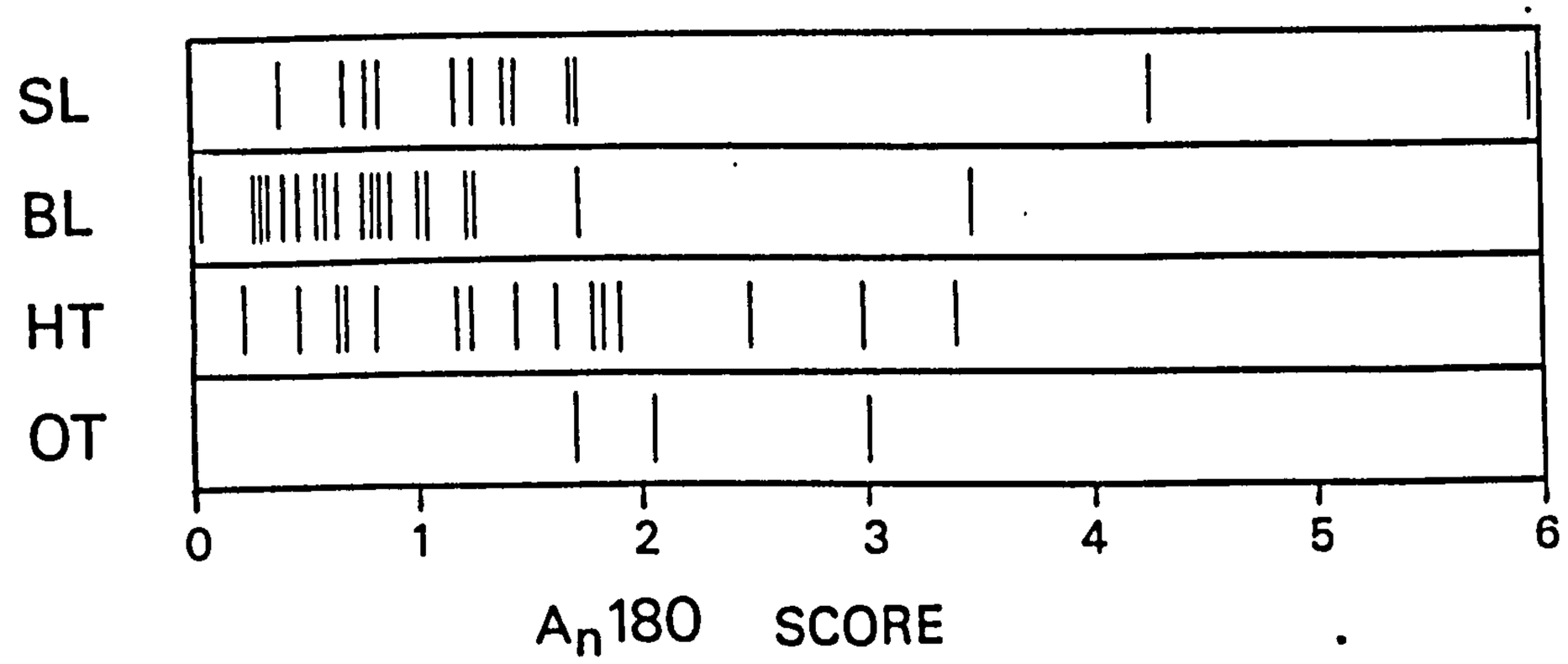


Figure 9.30: A_n360 and A_n180 scores for fabrics measured on solifluction lobes (SL), boulder lobes (BL), horizontal terraces (HT) and oblique terraces (OT). Each vertical line represents a single fabric measured on a sample of 50 clasts. The A_n360 statistic assesses strength of preferred downslope dip and the A_n180 statistic evaluates the strength of preferred downslope orientation (Dale and Ballantyne, 1980).

lobes are presented in figure 9.31. The grain size curves for solifluction lobes fall within a fairly narrow envelope: the clay-silt fraction (with one exception) constitutes only 6-8% of the total; the fine sand fraction (60-200 μm) makes up 19-38%; the medium sand fraction (200-600 μm) is generally dominant, comprising 32-51% of the total; and the coarse sand fraction ranges from 16-32%. Within this envelope regional variations are discernible, in that grain size is generally finer for the samples taken from lobes in an area of highly micaceous rocks in the Fannichs, and the fine sand fraction is somewhat depleted for lobes on the Torridon Sandstone of An Teallach. In all cases, however, fine and medium sand (60-600 μm) make up most (57-72%) of the fine fraction of the lobe. The granulometry of the sandy matrix underlying the openwork zone of a boulder lobe on Ben Wyvis falls within the envelope described by the solifluction lobe samples, but that for material under a boulder lobe on An Teallach exhibits a deficiency in the clay-silt and fine sand fractions, which together comprise less than 20% of the total.

The predominance of fine and medium sand in the Ben Wyvis and Fannich samples reflects a similar concentration of this size range in the <u>in situ</u> weathered detritus of these areas, and the curve for the An Teallach boulder lobe is similar to that for weathered detritus on Torridon Sandstone (figure 8.28). In general, therefore, the granulometry of both solifluction and boulder lobes appears to be inherited from that resulting from ~~microgelivation~~ granular disaggregation of the local bedrock. Only the samples from the An Teallach solifluction lobe are anomalous in this respect. It has been mentioned that on An Teallach most solifluction features occur at the foot of a slope around NH 057846, and the body of a lobe excavated in this area consisted entirely of fines (figure 9.22). From this it is reasonable to infer that these lobes formed in sandy material that had been washed downslope, so that the granulometry of the lobe does not reflect that of frost-weathered bedrock.

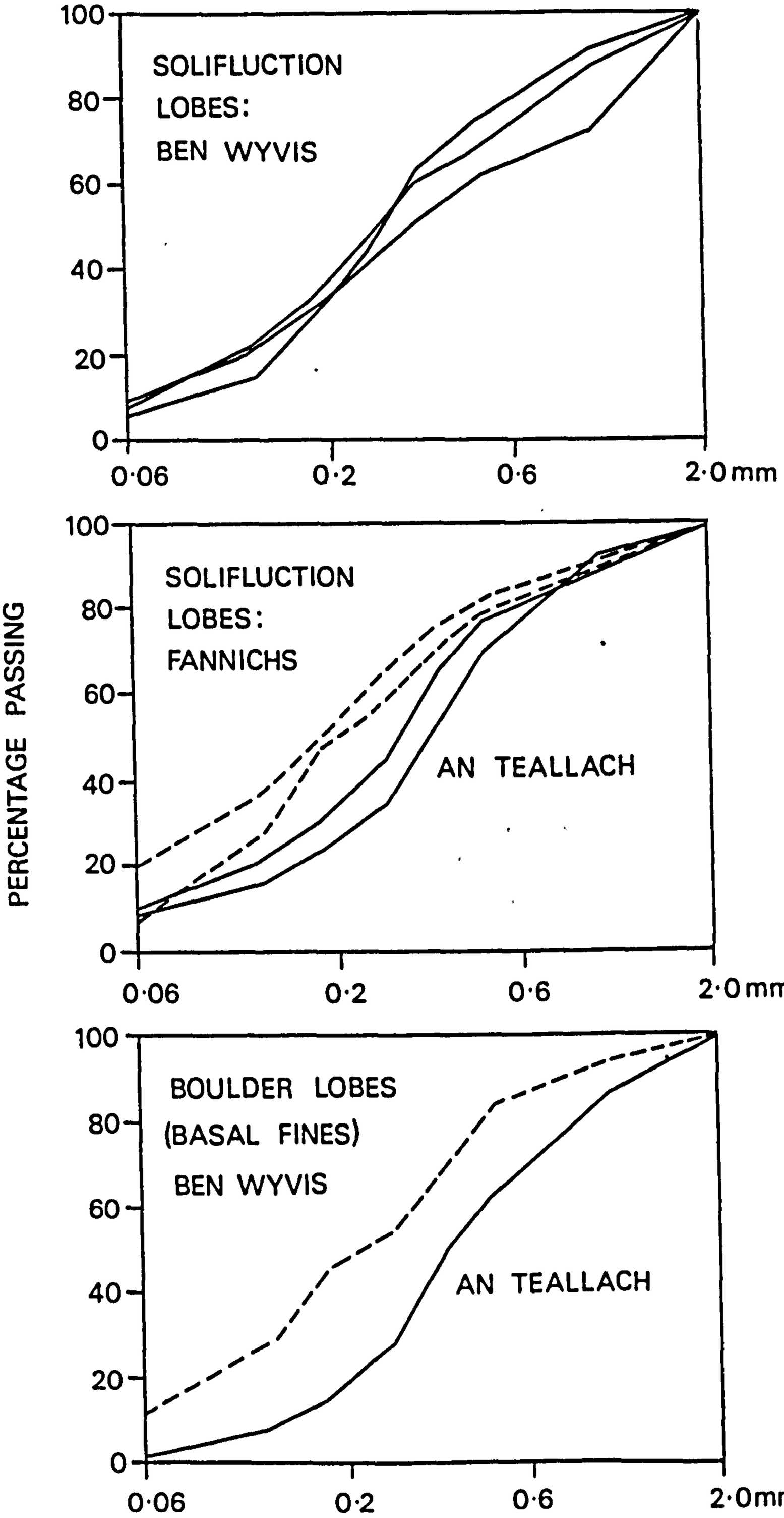


Figure 9.31: Grain-size distributions for samples of fine material collected from solifluction lobes and from below the openwork zone in boulder lobes.

9.2.5.5 Discussion

From the above descriptions it is evident that the sedimentological differences between boulder features and solifluction features extend beyond a difference in the coarseness of constituent material, the primary criterion employed here to distinguish the two types. The vertical sorting evident in the boulder lobe sections is similar to that observed on blockslopes (on which boulder lobes are frequently found) but there is no evidence for such sorting in the solifluction lobe sections. Buried soil horizons have survived (albeit in fragmentary form) under all the excavated solifluction lobes, but only one of the excavated boulder lobes overlies palaeosol fragments. A matrix of fines constitutes the bulk of all the trenched solifluction lobes, yet fines are rare throughout the upper portions of the boulder lobes. The strength of preferred downslope orientation and dip is generally greater for solifluction lobe fabrics than for boulder lobe fabrics, and the clasts themselves are significantly smaller. Lateral sorting, with the largest clasts at the lobe margins, is evident on some boulder lobes but not apparent on solifluction lobes.

The sedimentological characteristics of lobes reported above agree with the observations of previous authors. The correspondence between boulder lobes and areas of coarse detritus was noted by Galloway (1958, 1961a), and the concentration of large clasts in the upper layers of boulder lobes is similar to the "rubble layer" described by Ragg and Bibby (1966). King (1972) demonstrated lateral sorting of boulder lobes, and Birks (1973) noted that the fronts of solifluction sheets on Skye were " ... curling over to envelope the surface humus and vegetation", implying that the underlying soil is being buried by present-day movement. A section through a solifluction sheet on Ben Arkle illustrated by Mottershead and White (1969) closely resembles those shown in sections 1-4 in figure 9.22, but boulder sheets on a nearby blockslope they regarded as "genetically different". The problem of genesis is returned to in section 9.2.7.

9.2.6 Age and activity

Seven mass-movement sites of the type described in the appendix were set up across lobes with the marker poles inserted on either side of each lobe. Four of these were set up across solifluction lobes on Ben Wyvis, one across a boulder lobe on Ben Wyvis, and the remaining two across boulder lobes on An Teallach. No movement was recorded at either of the An Teallach sites over the period July 1976 - June 1978. The stakes at the Ben Wyvis boulder lobe site were disturbed by animals, but the relative position of the markers indicated that little or no movement had taken place. One of the four solifluction lobe sites was destroyed by animals before readings were taken, and only one year of movement (August 1976 - July 1977) could be recorded from the remainder before they too were destroyed by animals. Surface movement at the three intact solifluction sites over this period is shown on figure 9.32. Significant movement was recorded only at the two steeper sites, where the maximum rates of displacement were 17.4 and 10.7 mm y^{-1} on slopes of 29.5° and 27° respectively. The form of surface displacement suggests that movement reached a maximum at one point on the lobe and decreased towards the lobe margins.

The above evidence is consistent with interpretation of boulder lobes as Lateglacial features and solifluction lobes as of Flandrian age. Analysis of the pollen contained in buried soil fragments under the excavated solifluction lobe on An Teallach by M. Robinson (personal communication) indicated that this lobe was active in the early Flandrian, which is again consistent with this interpretation. Two samples from the humic material underlying the shallow boulder lobe excavated on the N.W. slope of Ben Wyvis (sections 6 and 9, figure 9.22) were however considered by Robinson to be of mid Flandrian age (7,000-5,000 B.P.) or younger, evidence that suggests that this lobe was active several millenia after the end of the Loch Lomond Stadial. It is difficult to assess the significance of this evidence, however; as Robinson pointed out, more recent pollen may well have been transported through the sandy body of these

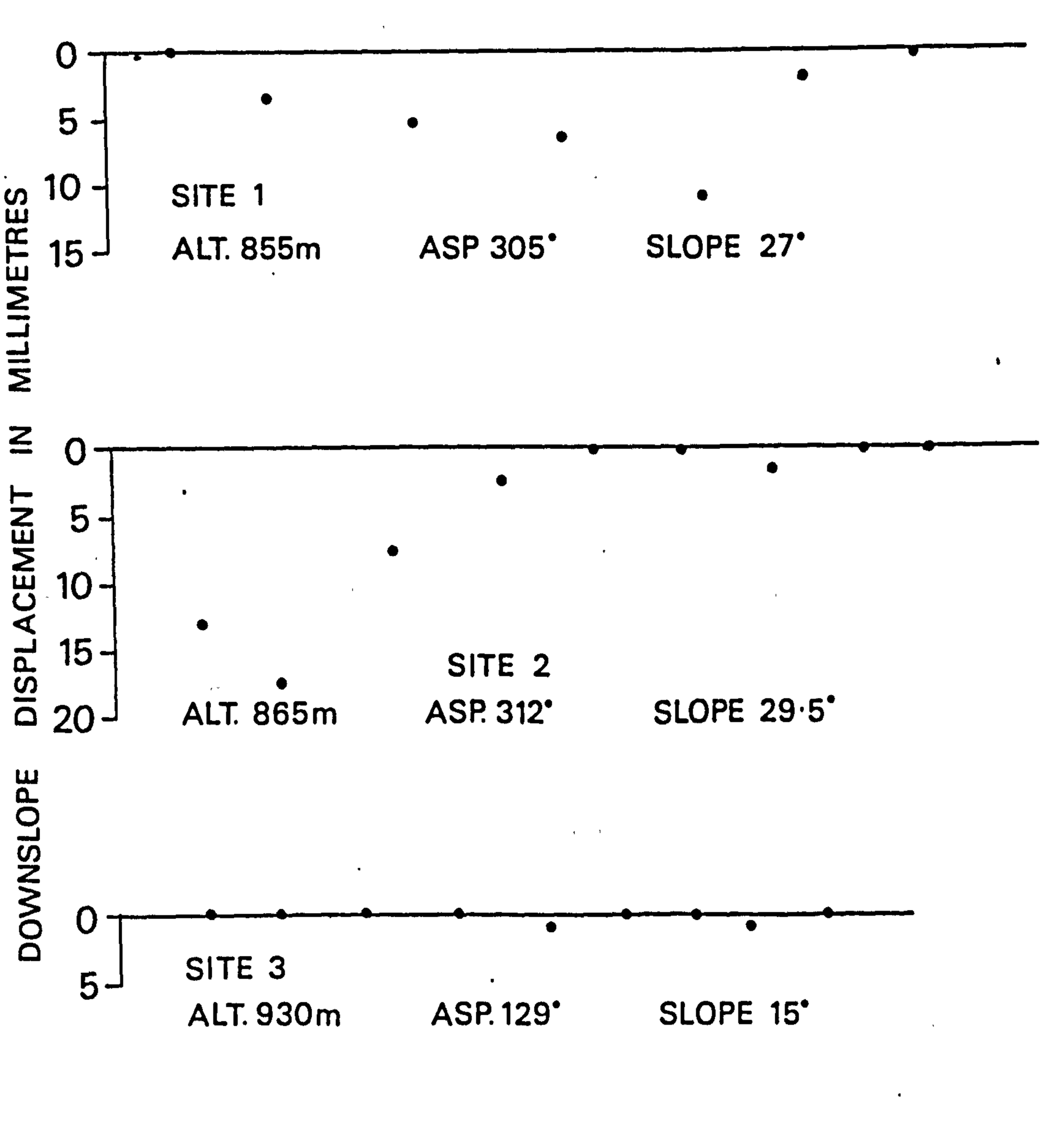


Figure 9.32: Downslope displacement of markers on the surfaces of three solifluction lobes at altitudes of 830-950 m on Ben Wyvis for the period August 1976 to July 1977.

lobes, giving a misleading assemblage of grains. It is safest to conclude that whilst the evidence for movement of solifluction sheets and lobes at present and throughout much of the Flandrian period is unequivocal, the majority of boulder sheets and lobes have apparently been immobile since the Loch Lomond Stadial, although some may have been active at some time during the Flandrian. It is perhaps noteworthy (i) that the openwork zone on the Ben Wyvis lobe that overlies the buried soil fragments is much shallower than those of the other two excavated boulder lobes; (ii) that the granulometry of this lobe resembles that for active solifluction lobes; and (iii) that it is located on a steep slope (c. 30°) on which solifluction lobes are currently active. On the other hand, the rounded nature of exposed clasts on this lobe contrasts with the angularity of buried clasts, suggesting that the surface stones (which are covered by mosses and lichens) have been subject to a long period of ~~microgelivation~~ granular disintegration during which disturbance has been minimal.

9.2.7 Processes

The formation and downslope movement of sheets and lobes of detritus has been widely attributed to "solifluction", a term coined by Andersson (1906) to represent " ... the slow flowing from higher to lower ground of masses of waste saturated with water" (p. 95). Later authors used the term to encompass frost creep and even patterned ground formation (Washburn, 1967, p. 11) but it is now generally employed sensu Andersson. As "solifluction" may refer to soil flow in any environment, Baulig (1956) proposed a related term, gelifluction, to indicate solifluction associated with frozen ground. Although the latter is now widely used (e.g. Washburn, 1973, 1979; Embleton and King, 1975), Andersson's original term is preferred here as it carries no genetic implications. Following Washburn (1967, 1979), the writer considers frost creep (see section 8.3.4) as a separate process that often accompanies periglacial solifluction rather than as a component of solifluction as envisaged, for example, by Högbom (1910), Sigafoos and Hopkins (1952), Williams (1959) and Jahn (1967).

The mechanics of solifluction are incompletely understood, and indeed the term may encompass more than one process (Carson, 1978). Translational sliding of the active layer under conditions of artesian pore pressures has been proposed by engineers working on low-angle clay slopes (Chandler, 1972; Hutchinson, 1974; Skempton and Weeks, 1976), but this explanation is inconsistent with the triangular displacement profiles of solifluction soils in alpine and sub-arctic environments (Williams, 1966; Benedict, 1970; Harris, 1977). These suggest that flow, rather than sliding, is the dominant form of movement. Liquifaction of the soil mass with moisture values exceeding or approximating the Atterberg liquid limit has been considered a probable cause of flow (e.g. Washburn, 1967), with the soil in a very loose condition following thaw and high pore pressures leading to reduced intergranular cohesion (shear strength) and thus flow. Williams (1966), however, carried out stability analyses on thawing soils, and found that theoretically stable soils had in fact experienced movement. Also, it is not clear why wholesale liquifaction does not cause mudflows (Price, 1969), unless vegetation acts to bind the liquified mass and thereby prevent failure (Embleton and King, 1975, p. 98), a possibility that engendered the idea of "bound" solifluction (Wilson, 1952). A more plausible explanation has been provided by Harris (1973, 1977) who argued that as ice-rich soil thaws there is short-term localized liquifaction at the thaw plane as ice lenses melt and consequent flow of saturated soil into the resulting voids. This idea of localized liquifaction avoids the difficulties raised by invoking liquifaction of the entire soil mass, and explains movement of soils that would be stable in terms of conventional stability analysis. Harris also suggested that this mode of movement may apply to non-cohesive (sandy) soils, but that shearing along distinct shear planes may characterise heavy plastic clays, a suggestion that accords with the conclusions (described above) of engineers working on clay slopes (Carson, 1978).

Washburn (1967, 1979) has proposed that four factors determine the susceptibility of soil to solifluction and the rate of soil movement by this process: moisture availability,

slope, vegetation cover and granulometry. There is unanimous agreement as to the fundamental importance of high moisture content (Sigafoos and Hopkins, 1952; Williams, 1957, 1959, 1966; Benedict, 1970, 1976; Chambers, 1970; Chandler, 1972; Furrer, 1972; Harris, 1972, 1973, 1977; Hutchinson, 1974). Movement is greatest where saturation is prolonged (Washburn, 1967; Harris, 1973). Latelying snow may therefore play an important role in sustaining solifluction (Williams, 1957, 1959; Kallander, 1967), although melting ice lenses may also act as an important source of moisture (Taber, 1943; Washburn, 1947). Harris (1973) has pointed out that snowcover during freezing is also of critical importance, as this slows the rate of freezing line penetration and thus favours the growth of ice lenses (Higashi, 1958; Everett, 1966) that not only yield water on thawing, but create discontinuities within the soil and promote heave. Harris also concluded that, for wet areas, the annual displacement is a function of slope, and for such slopes Washburn (1967) found a linear relationship between displacement and the sine of the slope angle.

The role of vegetation in promoting or hindering solifluction is less clear. Budel (1948, 1950) and Wilson (1952) distinguished "free" and "bound" solifluction, the principal characteristic of the latter being retardation by vegetation. Sigafoos and Hopkins (1952), however, considered that vegetation, by retaining water and retarding runoff, actually favours solifluction. Washburn (1967) concluded that the influence of moisture predominated over the impeding effect of vegetation, and that the distinction between "free" and "bound" types was misleading. In the same paper he proposed that silts, which are highly frost-susceptible but, unlike clays, lack cohesion and have low liquid limits, constitute the most favourable grain size for the operation of solifluction; sands tend to be well-drained and less frost-susceptible. In practice, solifluction operates over a broad range of grain sizes (Graf, 1973), although it is apparently favoured by soils containing a high proportion of silt size material.

The importance of high moisture content in promoting solifluction suggests that this process is likely to be widespread in the wet environment of upland Britain. The main

constraints on solifluction in the study areas are the granulometry of the regolith, the possible absence of an annual freezing cycle and the presence in some areas of an insulating cover of peat. The importance of grain size in limiting the distribution of solifluction features is illustrated by their absence on many of the slopes of An Teallach (where a regolith of medium to coarse sand with a low clay-silt fraction (figure 8.28) predominates) and by their comparative abundance on the micaceous rocks of the Fannichs, where similar climatic conditions prevail. A more general constraint is imposed by the lack of true annual freezing cycles in upland Britain, where periods of thaw may occur at any time during the winter (chapters 2 and 5), preventing the development of a deep frozen layer. Alternatively, the view may be adopted that the occurrence of several sustained freeze-thaw events each winter (events related to the alternating dominance of polar and tropical airmasses rather than annual or diurnal temperature regime) is likely to favour solifluction, albeit of a fairly superficial nature. The shallowness (low riser heights) of solifluction features in the study areas may reflect the infrequency of deep freezing.

The characteristics of the features identified in the present study as solifluction sheets and lobes are very similar to those reported from other periglacial areas. The properties of the latter have recently been summarized by Benedict (1976), and include the following.

(i) Vegetation covered ("turf-banked") sheets ("terraces") and lobes are the most abundant type and are characterised by a homogeneous structure and lack of vertical sorting.

(ii) Sheets and lobes are related features, the latter representing "channelization" (local acceleration) of movement; the fronts of sheets are often aligned oblique to slope.

(iii) Solifluction features are found on slopes of 2^{o}-35^{o}.

(iv) Virtually all solifluction lobes and terraces (sheets) contain at least one buried organic layer.

(v) Where movement is unimpeded solifluction fabrics display strong downslope orientation.

All of these characteristics are typical of the solifluction features of the study areas, implying that the latter move downslope in a manner similar to that of solifluction features elsewhere. Although such movement is normally dominated by solifluction sensu stricto (see above), frost creep may play an important, even dominant, role (Washburn, 1967; Benedict, 1970), although the lack of sorting in the excavated lobes indicates that flow (which tends to destroy frost sorting) is the principal mechanism.

Although not necessarily representative of long-term trends, the form and amount of surface displacement measured on the Ben Wyvis lobes (figure 9.32) resemble those recorded on similar features elsewhere. A decline in surface displacement towards the lobe margins has been described by several authors. (e.g. Washburn, 1967), and appears to be typical. Rates of surface displacement (summarized, for example, by Washburn (1973, p. 179), Embleton and King (1975, p. 103) and French (1976, p. 137)) generally fall within the range 5-100 mm y^{-1}, although Rapp (1962) and Williams (1959) recorded movements of 300 mm y^{-1} and 250 mm over three weeks respectively. Summarizing the results of a large number of studies, Benedict (1976) found that "median maximum" surface (?) velocity for sheets and lobes is 30 mm y^{-1}, greater than either of the maximum values measured on Ben Wyvis (11.7 and 17.4 mm y^{-1}), and based on rates recorded on generally gentler slopes. From the limited data available it therefore appears that present-day solifluction in the northern Highlands is less effective than in other periglacial environments. The validity of this generalization cannot be established, however, without further data on current rates of solifluction activity.

Whereas the movement of solifluction sheets and lobes is attributable to a widely-recognised if incompletely understood process, the explanation of boulder lobes is problematic. Several authors have suggested that periglacial boulder concentrations such as blockslopes and boulder lobes may represent relict

solifluction features immobilised through eluviation of fines (Andersson, 1906; Lozinski, 1912; Klatka, 1961, 1962; Jahn, 1967; Caine, 1968; Smith, 1968; Perov, 1969; Washburn, 1973, p. 427; Clapperton, 1975; Embleton and King, 1975, p. 118). Galloway (1958) adopted this idea to account for the stabilization of Lateglacial boulder sheets and lobes on Ben Wyvis and Lochnagar, an explanation accepted by Kelletat (1970b) in his account of boulder lobes at other localities in Scotland. This explanation is attractive in several ways. First, the relationship of boulder features with the limits of the Loch Lomond Advance glaciers indicates that they were formed under the severe conditions of the Lateglacial cold period, which could account for their being generally larger than presently-active solifluction features. Secondly, they occupy a range of slopes identical to that occupied by solifluction features. Thirdly, reduced strength of fabric may be accounted for by settlement of clasts concomitant with eluviation of fines. There is positive evidence for the latter in the shape of small fans of stratified sand downslope of the risers of some boulder lobes on Ben Wyvis (McMillan, 1978). Finally, lateral sorting (evident on some boulder lobes) has also been reported in association with some solifluction lobes (e.g. G. Lundqvist, 1949; J. Lundqvist, 1962; Rudberg, 1964), although the author is not aware of any examples of active stone-banked solifluction lobes in upland Britain.

Interpretation of Lateglacial boulder features as eluviated solifluction features fails, however, to explain all of their characteristics. It was established above that the thickness of solifluction lobes is related to slope, but that of boulder lobes reflects clast size and thus lithology. Also, the eluviation theory cannot explain why on the N.W. slope of Ben Wyvis a belt of boulder lobes lies between two belts of active solifluction features, although it is possible that the solifluction features downslope of the boulder lobe zone have formed in concentrations of fine material washed out of the boulder lobes (cf Jahn, 1975). A third difficulty lies in the coarseness of the fine fraction of some boulder lobes, such as that excavated on An Teallach. In terms of the curve of "frost

susceptibility" produced by Beskow (1935), the fines contained in the base of this lobe are too coarse to permit ice lens formation and are therefore unlikely to have experienced solifluction. A more serious objection to this explanation is that boulder sheets and lobes are commonly found on blockslopes, on rocks that have resisted ~~microgelivation~~ granular disintegration (e.g. microgranite (figure 9.11), granulite and fine-grained quartzite). In section 8.3.4 it was argued that the surface layers of such blockslopes have never contained fines and hence cannot have been subject to solifluction. Finally, it was shown above that vertical sorting is typical of all boulder lobes but does not appear in any of the solifluction lobe sections. This, however, may merely indicate that frost sorting and creep predominated over solifluction when these features were active, a suggestion made by McMillan (1978) and returned to below.

The deficiencies of the eluviation theory were recognized by Shaw (1977), who found that the upper layers of large granite boulder lobes on Lochnagar contained no fines whatsoever. He concluded that

> " ... this debris never had abundant fine material on these slopes. It is envisaged that the lobes moved by a process analogous to rock glacier creep, in which the interstices of the lobes were filled with ice ..." (p. 130).

The idea of boulder deposits creeping downslope through the deformation of interstitial ice has also been advanced by Brockie (1965) and Caine (1968) to explain the movement of stone streams and blockslopes respectively, but meets with apparently insuperable difficulties, as outlined in section 8.3.4. Such boulder deposits are insufficiently thick to allow deformation of a rock-ice mixture (Whalley, 1976), and lack features typical of rock glacier decay, such as kettle holes (Caine, 1968). Moreover, although rock glaciers of Lateglacial age have been identified in upland Britain (Dawson, 1977; Sissons, 1979b), these occupy talus-foot locations and are much more massive than even the largest boulder sheets or lobes. Finally, most of the ice contained in boulder features during the Lateglacial cold periods must have been subject to annual

melting; perennial ice could have occupied only the lowest layers of these features.

This last consideration leads to the suggestion that frost heave associated with annual thawing and freezing in the former active layer may have been the principal mechanism responsible for the Lateglacial activity of boulder sheets and lobes, and indeed the blockslopes on which they are developed (section 8.3.4). There are several points in favour of such an interpretation. It accounts for mass-movement on openwork blockslopes (in which fines are largely confined to the base of the regolith) and for the lateral sorting encountered on some boulder lobes. The fining-downward sequences in the upper zone of all excavated lobes also points to the operation of frost-heave and frost-sorting. Finally, the relationship between boulder features and the limits of the Loch Lomond Advance glaciers also accords with their interpretation as the products of large-scale frost creep under permafrost conditions, as deterioration of permafrost must have accompanied climatic amelioration, resulting in the immobilization of boulder sheets and lobes as the glaciers decayed.

Active "stone-banked" (boulder) sheets and lobes similar in structure and appearance to those in upland Britain have been described in the Front Range, Colorado, by Benedict (1970, 1976), who considered that

> " ... movement of blocky debris is characteristic of a frost creep environment ..." a hypothesis " ... supported by the strong vertical sorting of the features, and by the fact that they are best developed on steep upper slopes." (1970, p. 216).

He also believed that vertical frost sorting may have occurred before movement commenced, as inferred by Ragg and Bibby (1966) for the slope deposits on Broad Law, but as downslope creep inevitably accompanies frost sorting on slopes the two processes are likely to have occurred concurrently. Implicit in Benedict's explanation is the notion that frost creep is caused by heave of the fines that underlie the openwork zone (see also Rudberg, 1962, 1964). In some boulder lobes and blockslopes in the study areas the openwork zone is as much as one metre deep,

and in massive granite boulder lobes such as those of Lochnagar the base of the openwork zone apparently lies even deeper (Shaw, 1977). This implies that annual freezing and thawing must have extended to depths in excess of one metre for heave of underlying fines to have been effective. This is not an insuperable objection, however, as active layer depths in existing permafrost areas often exceed 2 m (Brown, 1970). A more serious difficulty arises from the coarseness of the basal fines under some block-slopes and boulder lobes. These lie close to or below Beskow's (1935) critical limits for the formation of lenses of segregated ice and are unlikely to have been susceptible to heave. It is arguable, however, that some fraction of the basal fine material (e.g. particles in the silt to medium sand size range) has been depleted through subsequent subsurface wash, evidence for which was described above.

The main deficiency in Benedict's explanation is that it fails to account for vertical sorting in the openwork zone as illustrated in figure 9.22. Upheaving of clasts contained in a matrix of fines, described by many authors (e.g. Högbom, 1910; Hamberg, 1915; Beskow, 1930, 1935; Schmid, 1955; Kaplar, 1965; Washburn, 1969), provides an adequate explanation for the formation of an uppermost openwork zone, and it is possible to envisage the downward migration of smaller clasts amongst a framework of boulders in this zone as heaving and resettling are repeated. Heave of the basal fines, however, cannot explain the concentration of large boulders at the top of the openwork zone (figure 9.26). This can only be explained in terms of frost heave of the entire deposit, as argued in section 8.3.4, with heaving occurring as interstitial ice formed between the clasts in the openwork zone and settling as this ice melted. It is therefore proposed that this mechanism was largely responsible for the downslope movement of boulder sheets and lobes during the Lateglacial.

There is no reason to suppose, however, that wholesale heave of the type described above constituted the only mechanism responsible for the movement of boulder lobes and terraces, although for massive features such as those of the Cairngorms and Lochnagar it could have been. For smaller boulder lobes such as

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the turf-banked boulder lobe excavated on Ben Wyvis (figure 9.22, sections 6 and 9) frost heave or solifluction of the frost-susceptible fines underlying the openwork zone may well have contributed to movement, although the well-sorted openwork layer in these lobes indicates that frost-creep rather than solifluction was the dominant form of movement (McMillan, 1978). One further reason for regarding this and other shallow turf-banked lobes as forms transitional between "true" boulder lobes and solifluction lobes is that their dimensions generally lie closer to those of the latter (table 9.2). One particularly revealing characteristic of this group of lobes is the steepness of their risers (mean = 51°). Steep risers are characteristic of solifluction lobes; those of more massive boulder lobes rarely exceed 35° (table 9.2; Benedict, 1970; Shaw, 1977).

One final point deserves consideration. It was shown above that solifluction features are active at present in the study areas, yet it was argued that relict boulder features here and elsewhere in upland Britain are not the product of solifluction. What, then, was the role of solifluction on high ground during the Lateglacial cold periods?

One possibility is that solifluction played little part in the evolution of the Lateglacial landscape in upland Britain. Several geomorphologists (e.g. Rapp and Rudberg, 1960; Lundqvist, 1962; Rudberg, 1962, 1964) have recognised a broad subdivision of periglacial landscapes in Scandinavia into a lower tundra zone (in which solifluction features are common) and an upper frost-shatter zone (dominated by blockfields and associated features, including boulder lobes). Under the more severe climatic regime of the frost-shatter zone there is little evidence of solifluction: the dominant processes are frost-heave and frost-creep. It is possible that the same may have been true of the mountains of Great Britain during the Lateglacial cold periods. A second possibility is that solifluction was active during the Lateglacial on regolith of suitable granulometric composition, and that solifluction features formed at this time have remained active up to the present. However, solifluction features during the Lateglacial would presumably have formed at

altitudes below those of presently-active sheets and lobes, but there is no evidence in the study areas (or, to the writer's knowledge, anywhere else in upland Britain) of inactive Lateglacial solifluction features downslope of active sheets and lobes. A final possibility is that shallow turf-banked boulder lobes such as those on the N.W. slope of Ben Wyvis represent the solifluction features of the Lateglacial period, even though frost creep was apparently the dominant mechanism responsible for their activity. Shallow boulder sheets and lobes of this type are in the author's experience more common than the massive boulder lobes of granite and quartzite areas. Apart from those on Ben Wyvis excellent examples of this type of feature occur in such widely-separated areas as the Merrick and White Cairn in the Southern Uplands, Ben a'Ghlo and Ben Lawers in the Grampians, the Mamores (Whyte, 1970) and Beinn Liath Mór Fannich (map 4). Given the shallowness of the openwork zone in these features and the apparent frost-susceptibility of the underlying fine matrix, the possibility of postglacial activity during exceptionally severe winters cannot be excluded. Indeed, if the Flandrian pollen assemblage found under the excavated lobe on Ben Wyvis is representative rather than inwashed, then postglacial movement of shallow boulder lobes of apparent Lateglacial origin is a certainty.

9.2.8 Conclusions

As suggested in the initial hypothesis (section 9.2.1), solifluction features and boulder features represent fundamentally different types, being different not only in age but in size, shape, distribution, structure and mode of activity. The former are active at the present day and closely resemble solifluction features in other periglacial areas. Movement is apparently a combination of solifluction sensu stricto (localized liquifaction resulting from the thaw of ice lenses) and frost creep. The unsorted structure of solifluction lobes suggests that the former process is dominant. Present rates of movement appear to be lower than those reported from other periglacial areas. The granulometry of the regolith exercises a

fundamental local control on the distribution of solifluction features, and across Scotland the lower limit of active solifluction features declines to the north and west.

For convenience, boulder features may be considered to be of two types, "shallow" and "massive", although it is likely that these represent the end members of a continuum. "Shallow" boulder features are those in which the surface openwork zone is underlain at shallow depth (< 50 cm) by a zone containing a matrix of fines; "massive" boulder features are those with a deep openwork zone. Both are apparently of Lateglacial origin and moved downslope as a result of large-scale frost creep under permafrost conditions, although frost heave and solifluction of the fine matrix underlying the openwork zone may have contributed to movement of the shallow type. It is possible that some features of the shallow type have experienced limited activity during the Flandrian. As with the solifluction features, the lower limits of boulder features (undifferentiated) decline northwards and westwards across upland Britain.

Criteria for distinguishing the three types without excavation are as follows.

(i) Solifluction features are usually vegetation-covered, sometimes turf-banked, with steep (typically around 45°) generally low (< 1 m) risers. Fine material is normally present at the ground surface.

(ii) "Shallow" boulder features are usually turf-banked or largely unvegetated, with generally steep (40° to 50°) and low (< 1.2 m) risers. They differ from (i) in that the treads (and sometimes the risers) consist, on the surface, of coarse detritus.

(iii) "Massive" boulder lobes may be unvegetated, stone-banked or vegetation-covered. Riser angles are much gentler than those of (i) and (ii), typically 20° to 30°, but riser height usually exceeds one metre.

The above criteria are generalizations, and do not hold in every case.

9.3 Terraces

9.3.1 Introduction and literature

By far the most common type of periglacial terrace feature in upland Britain is the turf-banked terrace, characterized by a steep, well-vegetated riser and a gently-sloping, bare tread (chapter 7). Features of this type have been described in many areas, including Ronas Hill, Shetland (Ball and Goodier, 1974), Ward Hill, Orkney (Goodier and Ball, 1975), Ben Klibreck (Kelletat, 1970a), Ben Arkle (Mottershead and White, 1969), the quartzite hills of Caithness (Crampton, 1911; Crampton and Carruthers, 1914), An Teallach (Godard, 1965; Ballantyne, 1977), the Fannichs (Peach et al., 1913a), Càrn Bàn and Ben Wyvis (Peach et al., 1912, Galloway, 1958, 1961a), the Applecross hills (Kelletat, 1970a, Robinson, 1977), Rhum (Ryder, 1968; Ryder and McCann, 1971; Mathieson, 1977), the Mamores (Whyte, 1970), the Cairngorms (King, 1968, 1971b), the Drumochter hills (Kelletat, 1970a), the Lake District (Hollingworth, 1934; Hay, 1937) and the northern Pennines (Tufnell, 1969). In some accounts other terms have been used to refer to these features, such as denuded steps (King, 1968, 1971b), parallel terraces (Tufnell, 1969) and solifluction terraces (Sissons, 1976a), but it is clear from the accompanying descriptions that the forms referred to are similar to those considered here. Similar features have also been reported from other periglacial mountain areas, for example the "turf-banked steps" of Mt. Washington (Antevs, 1932), the turf-banked terraces or "mud terraces" of Sweden (G. Lundqvist, 1949; J. Lundqvist, 1962) and the "miniature turf-banked terraces" of the Colorado Front Range (Benedict, 1970).

There is no concensus as to the way in which turf-banked terraces form, or the nature of terrace activity. Peach et al., (1912) judged such terraces to be the product of "earth creep", and Hollingworth (1934) proposed that movement took the form of frost creep, with the debris moving around "dams" of stationary turf, a suggestion adopted by Benedict (1970). Mottershead and White (1969), however, considered frost creep unimportant as they

judged the debris on which they found terraces to be too coarse to be susceptible to frost heave. Solifluction has been widely invoked as the mechanism responsible for terrace formation and/or movement (Antevs, 1932; Hay, 1937; Galloway, 1958, 1961a; Lundqvist, 1962; Ryder and McCann, 1971; Mathieson, 1977) though it is often not clear what the advocates of this mechanism mean to convey by the term "solifluction".

A completely different viewpoint was adopted by King (1968, 1971b), who noted that turf-banked terraces in the Cairngorms tend to be aligned either parallel to or normal to the prevailing wind direction and concluded that turf-banked terraces are deflation features. The bare treads he attributed to "turf exfoliation" (Rasenabschälung; Sapper, 1915), a combination of needle ice action and deflation; the risers on some terraces he considered to have been formed through aeolian deposition.

Many investigators have been puzzled by the fact that turf-banked terraces often dip steeply across the slope. Such oblique terraces were observed in Sweden by Högbom (1914) and Frodin (1918), who suggested that these represent elongate polygons or possibly structural control, but reviewing their work Lundqvist (1962) admitted that these explanations were unacceptable and that the origins of oblique terraces were "not clear". The difficulty of explaining oblique terraces is compounded by their sometimes occupying the same slopes as horizontal terraces, yielding the phenomenon of interconnecting terraces, illustrated by a number of authors (e.g. Hollingworth, 1934, figure 25; Lundqvist, 1962, figure 12) but not yet satisfactorily explained. Another feature of some terraces, also unexplained, is the formation of small lobes at the terrace fronts (Mottershead and White, 1969).

In Great Britain the most thorough work on turf-banked terraces has been carried out by Ball and Goodier (1974; Goodier and Ball, 1975) on the windswept slopes of Ronas Hill (Shetland) and Ward Hill (Orkney). Judging by "freshness of form", they considered those in the former area active, in the latter "less active". Like King, they recognised the affinities between turf-banked terraces and wind stripes (elongate deflation scars on otherwise vegetated terrain; chapter 10) and sought to explain the origin of terraces

in terms of both slope and the direction of dominant wind. Having found, like Lundqvist (1962) " ... that the orientation (of oblique terraces) is not that which would occur from downslope solifluction alone", they were, however, forced to admit that " ... how the interaction of wind and slope is effective in controlling this orientation is not at all clear" (1974, p. 96). Working in the Colorado Front Range, Benedict (1970) had no such reservations in explaining the orientation of oblique terraces, which he considered to be:

> " ... aligned either parallel or at right angles to the prevailing winter wind direction, whichever orientation comes closest to paralleling the contour of the slope" (p. 171)

The authority of Benedict's explanation belies the lack of evidence he supplied. Indeed, lack of detailed evidence is symptomatic of nearly all the abovementioned work, and has hindered the emergence of a general explanation of the development of turf-banked terraces. The formulation of such a general explanation represents the aim of the research reported in this section.

9.3.2 Characteristics of turf-banked terraces

9.3.2.1 General

The classification set out in chapter 7 recognizes four types of turf-banked terrace: horizontal, oblique, interconnecting and lobate (figures 9.1, 9.12, 9.14-9.18). It is arguable, however, that the last-mentioned is merely a morphological variant of the horizontal type, and that interconnecting terraces represent the intersection of two primary terrace types, horizontal and oblique. The relationship between these two primary types is interesting. The treads of horizontal terraces, which have a low (generally $< 10^{o}$) across-slope dip, often end in short oblique ramps that connect treads at different levels. On some slopes these ramps approach or exceed the horizontal treads in size, giving rise to interconnecting (horizontal and oblique) terraces. On other slopes turf-banked ramps dipping steeply across-slope are continuous for many metres, producing the impressive features here termed

oblique terraces. Even on the most fully-developed oblique terraces (those on the southern slope of Glas Mheall Mór, An Teallach; figure 9.17) horizontal terraces never completely disappear, as they take the form of shallow turf-banked steps cutting across the broad treads of oblique terraces. There is therefore a continuum from horizontal terraces through interconnecting terraces to oblique terraces.

Another continuum exists between sheets and terraces. Earlier (section 9.2.1; figure 9.1) these two classes of feature were distinguished on the basis of tread (downslope) gradient, as sheet tread angles tend to increase linearly with slope angle (equation 9.1) whereas terrace tread gradients increasingly depart from slope angle as the latter increases (equation 9.2). Thus, although the two classes of feature are distinctly different on steep slopes, they are very similar on gentle slopes, and indeed one type may merge into the other. Such a transition is well illustrated on the slopes south of Sgùrr Mór (map 4), where turf-banked terraces of exceptional length (up to 87 m) and width (up to 12.3 m) merge with large vegetation-covered terraces which in turn grade into an area of vegetation-covered solifluction sheets. A similar continuum exists on gentle slopes south of Coire Granda (map 6), where the distinction made between vegetation-covered solifluction sheets, large vegetation-covered terraces and large turf-banked terraces is of necessity arbitrary. The similarity between sheets and broad terraces on gentle slopes is heightened by the development of small solifluction lobes at intervals along the risers of such terraces.

The most striking feature of turf-banked terraces of all types is the contrast in vegetation cover between the sparsely-vegetated treads and the often thickly-vegetated risers. This contrast is particularly prominent on terraces developed on Moinian rocks, where riser cover is often complete, but less so on the Torridon Sandstone of An Teallach. On the latter mountain a quadrat study of 118 terraces was carried out using a 0.5 m^2 frame that was placed at random on the riser and then on the tread of each terrace. Measured tread vegetation cover ranged from 0-34% with a mean of 5.3% (2.5% Calluna vulgaris; 2.4% grasses

such as Festuca, Deschampsia and Juncus spp.). Riser cover ranged from 20-96%, with a mean of 54.0% of which Calluna vulgaris made up 32.8%, grasses 12.4% and Alchemilla alpina, Vaccinium spp., various mosses and the lichen Cladonia most of the remaining 8.8%. Despite the fairly sparse cover on some terrace risers on An Teallach, in no instance did tread cover exceed 50% of riser cover.

9.3.2.2 Morphology

Variation in terrace morphology and the relationship between morphology and possible controls such as slope, altitude, aspect, vegetation cover and lithology were investigated through surveying 358 terraces at 26 sites on An Teallach (11 sites, 198 terraces), Ben Wyvis (6 sites, 60 terraces), the western Fannichs (3 sites, 30 terraces), Sc\`urr M\'or in the eastern Fannichs (1 site, 20 terraces) and S\`ail Mh\'or, a 767 m Torridon Sandstone hill 5 km north-west of the summit of An Teallach (5 sites, 50 terraces). Details of these sampling sites are given in table 9.4. The variables measured on each terrace are illustrated on figure 9.33 and defined in table 9.5. With the exception of 25 oblique terraces surveyed at site 11 (the south slope of Glas Mheall M\'or, An Teallach) all the terraces surveyed were of the horizontal type, including horizontal lobate terraces and the horizontal components of interconnecting terraces. For completeness, the sample of horizontal terraces includes examples of features described above as transitional to solifluction sheets on Sc\`urr M\'or (site 26) and south of Coire Granda, Ben Wyvis (sites 18-21).

Figure 9.34 summarizes some of the results of the terrace survey. Horizontal terraces display great variation in size, with lengths ranging from 2.8 m to 87.1 m and tread widths from 0.5 m to 12.3 m. Nearly all the largest terraces, however, are the features described above as transitional to solifluction sheets on Ben Wyvis and Sc\`urr M\'or. If such transitional features are discounted, 98% of horizontal terraces have lengths of less than 20 m and tread widths of less than 4 m, and approximately 70% have lengths less than 10 m and widths less than 2 m. In contrast, only 36% of oblique terraces have lengths of less than 10 m and only 16% have widths of less than 2.0 m.

Table 9.4: Turf-banked terrace survey sampling sites

Area	Site no.	N.G.R. (NH)	Approx. Altitude	No. of Terraces
An Teallach (map 2)	1	073868	700 m	20
	2	072868	710 m	18
	3	073865	740 m	19
	4	063862	740 m	19
	5	068854	800 m	20
	6	072867	735 m	20
	7	070858	775 m	22
	8	069860	775 m	10
	9	068856	800 m	5
	10	071854	795 m	20
	11	075850	690 m	25
Sàil Mhór	12	035887	745 m	10
	13	036887	715 m	10
	14	037885	675 m	10
	15	038883	685 m	10
	16	039881	580 m	10
Ben Wyvis (maps 5 & 6)	17	501709	855 m	10
	18	502710	860 m	10
	19	500703	810 m	10
	20	501710	870 m	8
	21	498722	760 m	12
	22	472677	850 m	10
W. Fannichs (map 3)	23	148711	910 m	10
	24	149710	870 m	10
	25	151710	900 m	10
E. Fannichs (map 4)	26	205715	1000 m	20

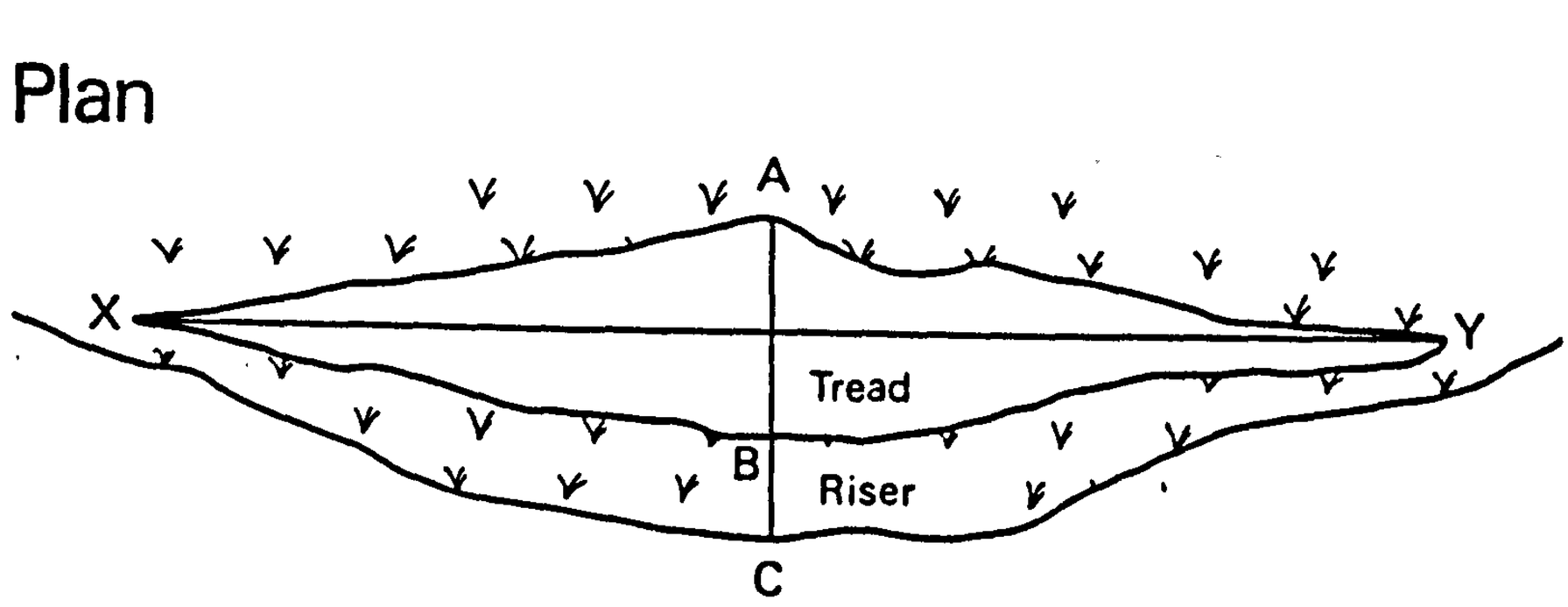


Figure 9.33: Definition of morphological variables measured on terraces (see table 9.5).

Table 9.5

Turf-banked terrace survey: definition of measured and derived variables

(See figure 9.33)

Variable	Symbol	Definition
Length	ℓ	XY
Width	w	AC (maximum width)
Tread width	w_t	AB
Riser width	w_r	BC
Height	h	CC'
Tread height	h_t	BB'
Riser height	h_r	CC' - BB'
Tread angle	α_t	$\tan^{-1} (w_t / h_t)$
Riser angle	α_r	$\tan^{-1} (w_r / h_r)$
Dip across slope	γ	Across-slope dip of tread along line XY, measured by Abney level.
Orientation	θ	Orientation of XY, measured down-dip with respect to magnetic north.
Aspect	θ'	Orientation of AC, measured downslope with respect to magnetic north.

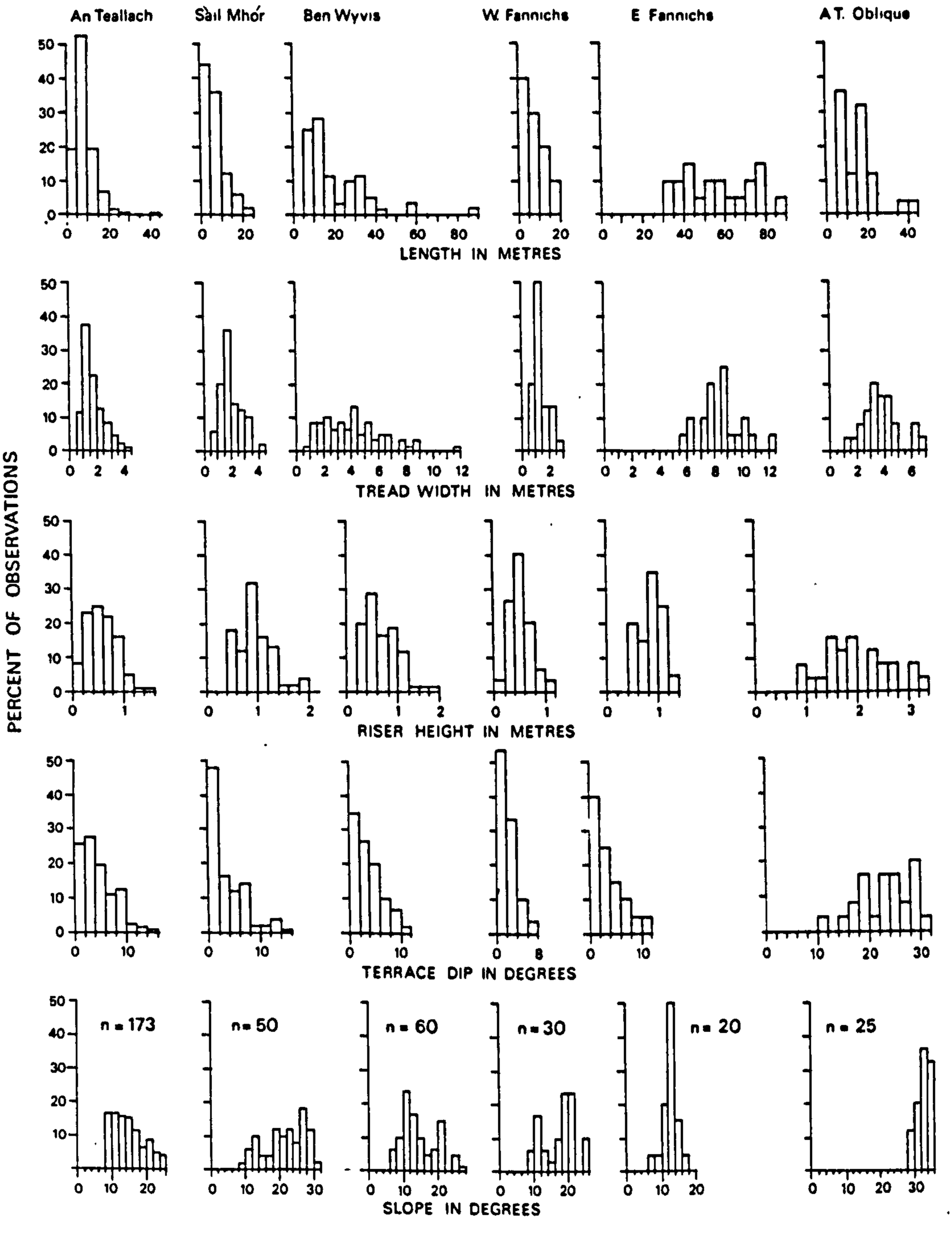


Figure 9.34: Percentage frequency distributions summarizing some of the data obtained in the survey of turf-banked terraces.

An even greater distinction exists between horizontal and oblique terraces in terms of riser height. Only 4% of all 333 surveyed horizontal terraces have risers exceeding 1.2 m in height, yet this figure is exceeded by the risers on 22 (88%) of the 25 oblique terraces surveyed. There is, however, no distinction between "transitional" and other horizontal terraces in terms of riser height. This is not surprising, as few solifluction sheets have risers exceeding 1.2 m. Similarly, the distribution of dip values is similar for both "transitional" and other horizontal terraces, for which only 2% have across-slope dips exceeding 12°, but very different (as would be expected) for oblique terraces (96% with dips exceeding 12°). Finally, although the data show that horizontal terraces occupy slopes in the range 6°-31°, oblique terraces are confined to steeper slopes (28°-36°). In part, however, the smallness of the latter range reflects sampling from a single steep slope.

For the sample of all 333 horizontal terraces certain relationships exist between morphological variables (table 9.6). By far the strongest relationship is the positive linear correlation between length and tread width: since wide terraces are less likely to be truncated by oblique ramps or the risers of other terraces they tend to be longest. The relationships between the other morphological variables reflect their common response to a number of independent controls, of which slope, aspect and clast size are the most important.

9.3.2.3 Factors controlling terrace morphology

Possible relationships between terrace morphology (variables ℓ, w, w_t, w_r, h, h_t, h_r, α_t, α_r, and γ) and likely controlling factors (altitude, slope, aspect, tread and/or riser vegetation cover) were explored by generating scattergrams for the former against each of the latter. This exercise was confined to the 333 horizontal terraces, as the sample of oblique terraces was considered to have been measured over too small a range of slopes, aspects and altitudes to give meaningful results. In addition, the influences of lithology on the morphology of horizontal terraces was examined by comparing those developed on Torridon Sandstone

Table 9.6

Correlations between selected morphological variables (Horizontal terraces only).

		ℓ					
	r^2	.784					
w_t	p	.00001					
	rel.	L +	w_t				
	r^2		.086				
h_r	p	N.R.	.00001				
	rel.		NL +	h_r			
	r^2	.102	.103				
γ	p	.00001	.00001	N.R.			
	rel.	L -	L -		γ		
	r^2	.097	.125		.053		
α_t	p	.00001	.00001	N.R.	.00001		
	rel.	NL -	NL -		L +	α_t	
	r^2	.049	.040	.068			
α_r	p	.00002	.0002	.00001	N.R.	N.R.	
	rel.	L +	L +	NL -			

n = 333

L = linear

NL = nonlinear

N.R. = no relationship

(An Teallach and Sàil Mhór) with those developed on Moine rocks (Fannichs, Ben Wyvis). Finally, the effect of particle size was investigated by measuring the intermediate axes of 50 clasts at the fronts of the treads of 20 horizontal terraces at 10 different sites on An Teallach, and generating scattergrams of ℓ, w and h against mean intermediate axis length ($\overline{b}$).

Scattergrams of the morphological variables specified above against altitude and vegetation cover revealed no meaningful relationships, even when the samples from different mountains were analysed separately. Slope, however, is a major control of terrace morphology (table 9.7). Terrace form adapts to increase in slope in various ways, of which the most important is a rapid decrease in tread width (figure 9.35) and thus total terrace width and length. Riser height and tread slope also tend to increase as slope steepens, as does across-slope dip. If the sample of oblique terraces is plotted on the same scattergrams, as in figure 9.36, then differences in response between the two types emerge strongly, providing further evidence that horizontal and oblique terraces are fundamentally different landforms.

The influence of aspect on terrace morphology cannot be assessed in the same way, as orientation data are measured on a circular scale with an arbitrary origin. The influence of aspect may, however, be tested by transforming the data so that they reflect departure (through 180°, both clockwise and anticlockwise) from a selected origin. Thus if 50° is selected as origin, then 340° becomes 70°, 310° becomes 100°, 180° becomes 130° and so on. The data were transformed in this way for origins at 10° intervals (0°, 10°, 20°, ... 350°), and the resulting scattergrams inspected for trends. Certain positive relationships emerge when the origin is located at or near 270°, and equivalent negative relationships when it is around 90°. No relationships are evident for origins of 310°-50° or 130°-230°. Regression relationships for morphological variables against departure in aspect from 270° are given in table 9.7. The analysis was restricted to 108 terraces on the north plateau of An Teallach in order to minimise the influence of other controls. The results indicate that with departure from a westerly aspect terrace risers increase in width and height but decrease in slope, so that terrace width and height tend to be largest at easterly aspects, as does tread width.

Table 9.7

Relationships between terrace morphology and slope and transformed aspect

	Slope (n=333)		Transformed aspect (n=108)	
	Regression	r^2	Regression	r^2
Length	$\ell = 526.6\alpha^{-1.441}$	.375	No relationship	
Width	$w = 1951.3\alpha^{-0.646}$	.129	$w = 215.7\, e^{0.010\beta}$	.200
Tread width	$w_t = 4973.5\alpha^{-1.142}$	.316	$w_t = 146.1\, e^{0.008\beta}$	.093
Riser width	No relationship		$w_r = 60.2\, e^{0.014\beta}$	.150
Height	No relationship		$h = 65.3\, e^{0.007\beta}$	.147
Tread height	No relationship		No relationship	
Riser height	$h_r = 1.621\alpha + 38.7$	.073	$h_r = 42.9\, e^{0.008\beta}$	.128
Tread slope	$\alpha_t = 0.868\alpha^{0.779}$	.336	No relationship	
Riser slope	No relationship		$\alpha_r = 36.8 - 0.146\beta$	.095
Dip	$\gamma = 0.253\alpha - 0.465$	.175	No relationship	

Notes: 1. Significance assessed as $p < .001$.

2. Aspect transformation given by

$$\beta = \begin{cases} \theta' + 90 & \text{if} \quad \theta' \leq 90 \\ 270 - \theta' & \text{if} \quad 90 < \theta' \leq 270 \\ \theta' - 270 & \text{if} \quad \theta' > 270. \end{cases}$$

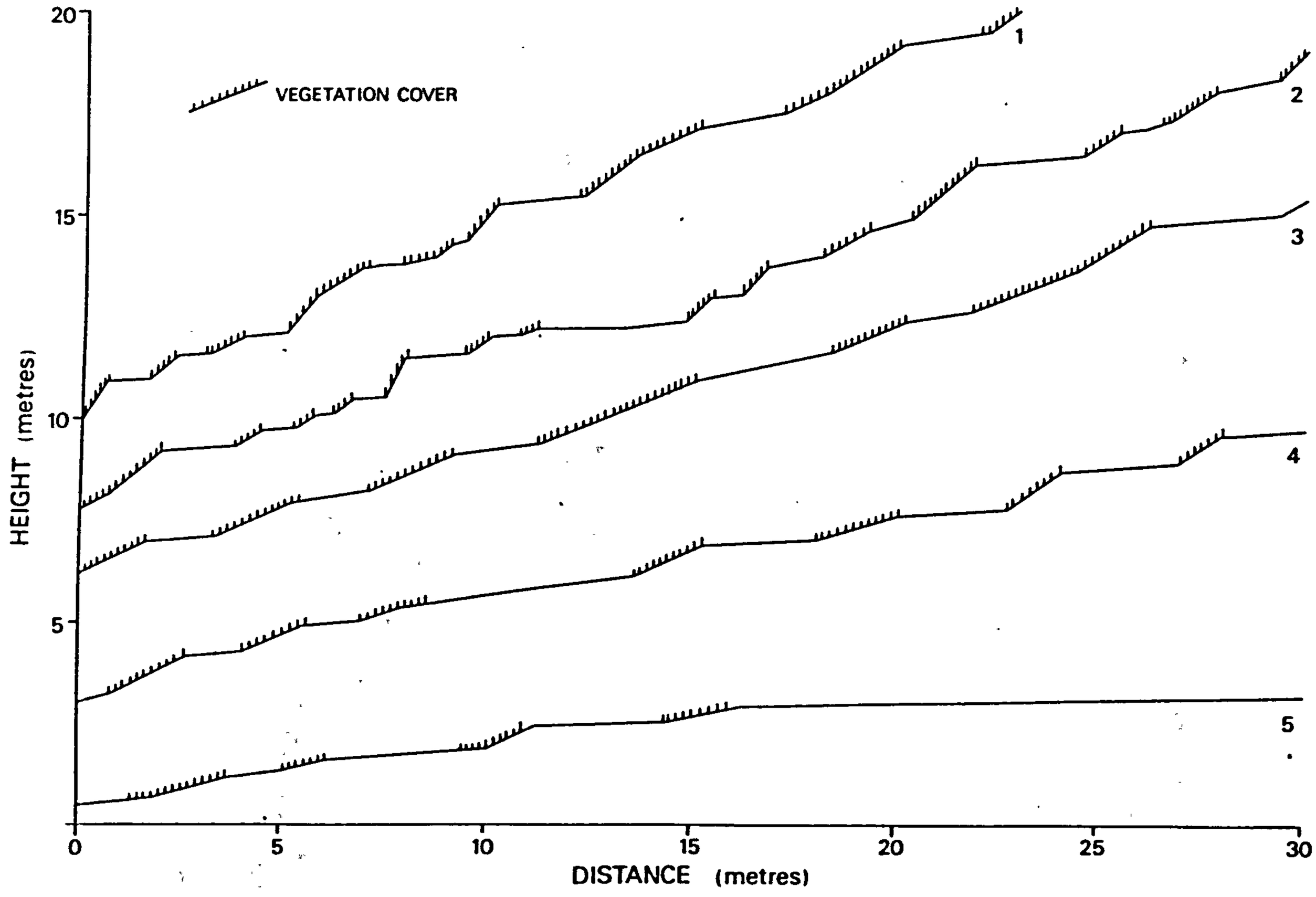


Figure 9.35: A transect levelled from the northern plateau of An Teallach at 073866 along a bearing 70° east of north showing variations in terrace morphology with increasing slope. (The transect continues from each line to the next).

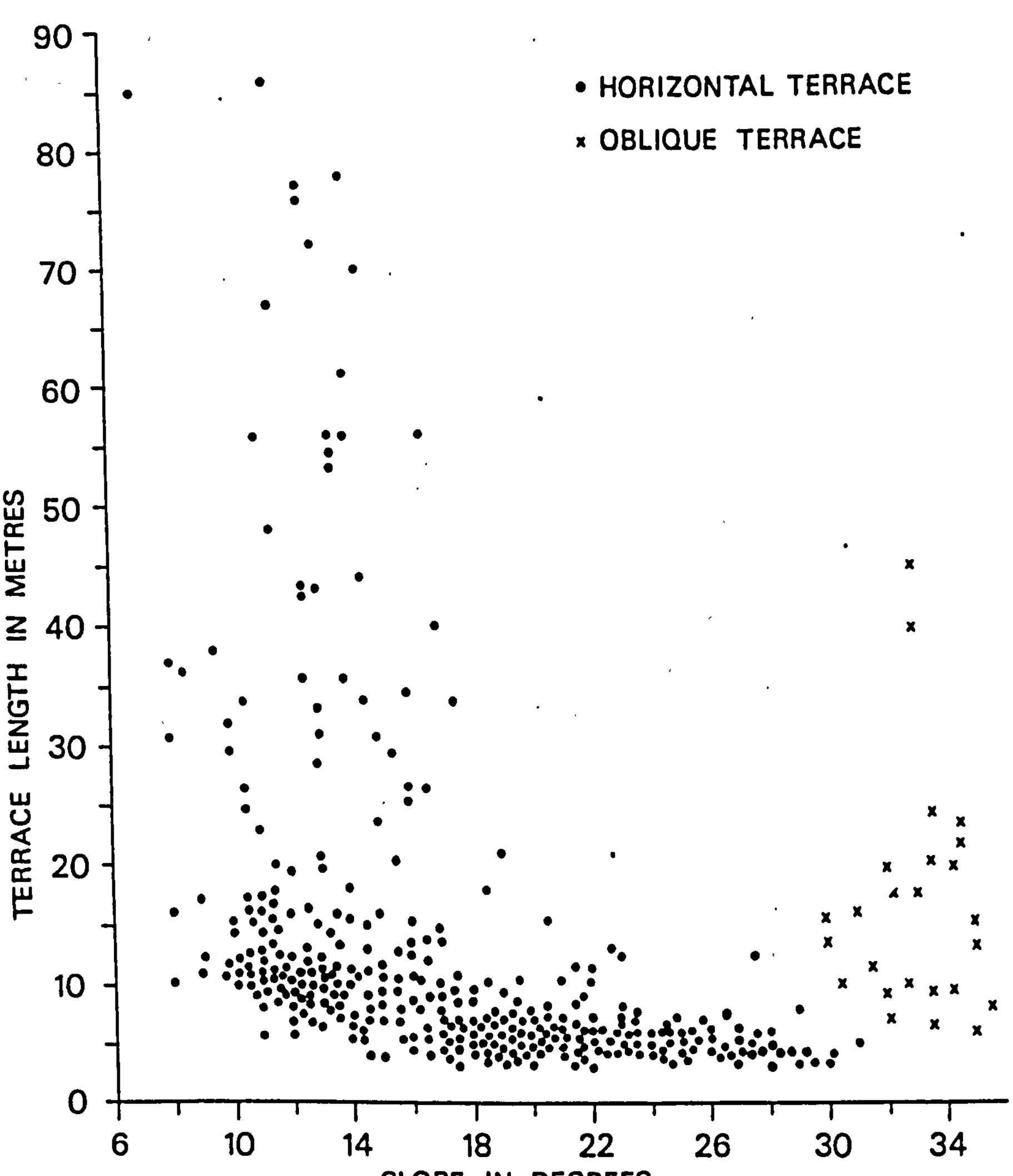


Figure 9.36: Scattergram of terrace lengths plotted against equivalent slope angles for 333 horizontal turf-banked terraces and 25 oblique turf-banked terraces.

These relationships suggest optimal terrace development on lee slopes, a point returned to later.

Comparison of the dimensions of terraces developed on Torridon Sandstone with those on Moine rocks (figure 9.34) shows that terraces surveyed in the western Fannichs have similar size and dip distributions to those on An Teallach and Sàil Mhór. Differences in length and tread width between these distributions and those of terraces surveyed on Ben Wyvis and the eastern Fannichs is attributable to the inclusion of transitional "terrace-sheet" features in the latter areas (the eastern Fannichs sample consisting entirely of transitional features). It therefore appears that lithology does not exert a systematic influence on the dimensions of horizontal turf-banked terraces, except that transitional terrace-sheet forms are apparently restricted to regolith that is prone to solifluction. It was established earlier (section 9.2.5.4) that solifluction features on An Teallach are found only on washed sand deposits as Torridon Sandstone regolith is normally too coarse to favour solifluction.

The size of horizontal terraces is nonetheless strongly related to the mean size of clasts at the front of the terrace tread (figure 9.37): the larger mean clast size, the larger the length, width and height of the terrace. It therefore is possible to have terraces of greatly differing sizes on slopes with similar gradient and aspect, as indeed occurs on An Teallach.

9.3.2.4 Structure and sedimentology

The structural and sedimentological characteristics of turf-banked terraces were investigated by trenching terraces at seven sites (table 9.8). The stratigraphy at each site (except 7) was recorded, samples of matrix material were withdrawn for mechanical analysis, and size, shape and fabric measurements were carried out at all sites except number 6.

Sections cut through terraces 1-6 revealed considerable variability in structure (figure 9.38). A conspicuous feature of the An Teallach examples (1, 2, 3, 5) is the crude tripartite stratigraphy of these terraces, with a stony surface layer (usually

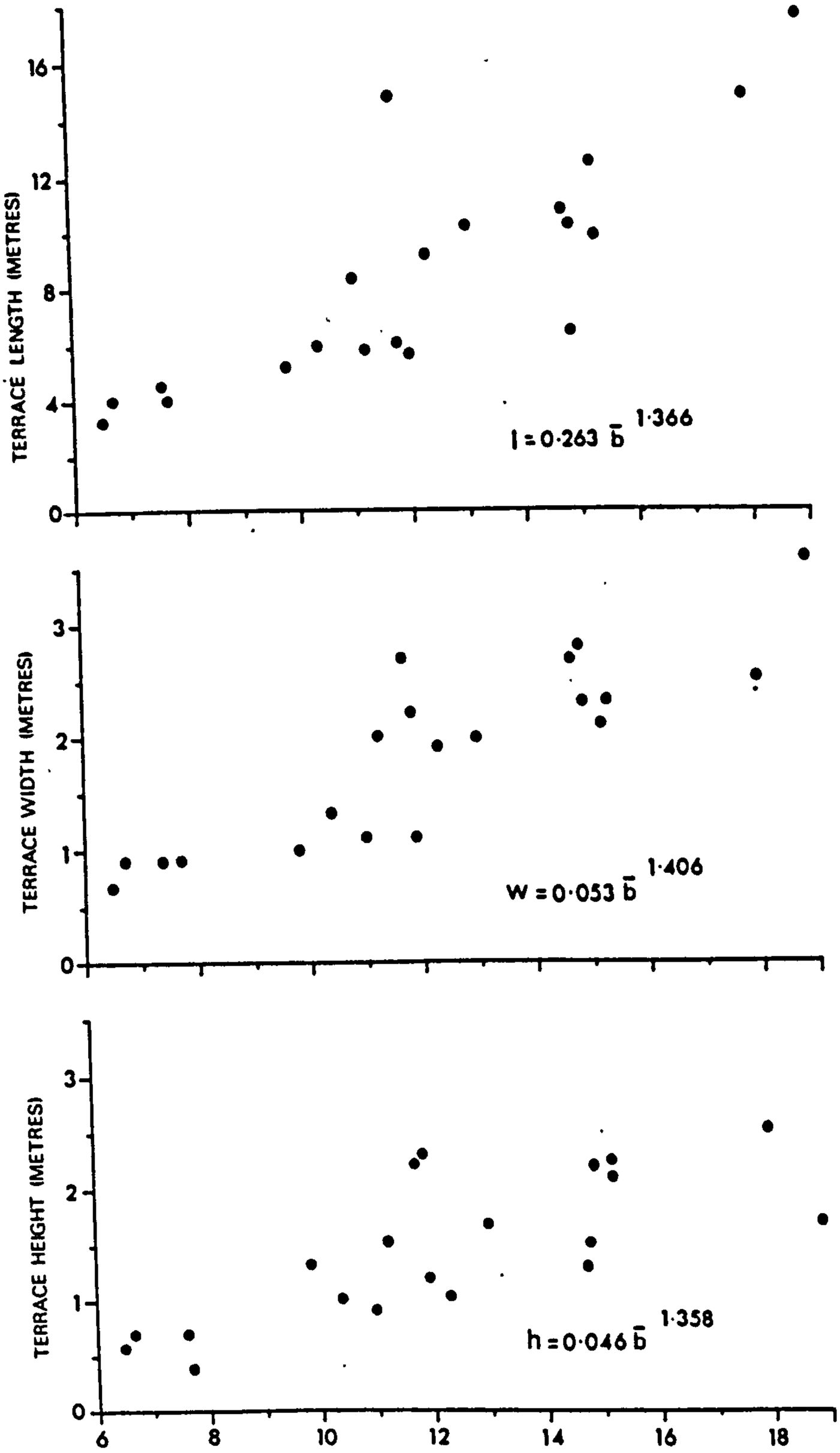


Figure 9.37: The lengths, widths and heights of 20 horizontal turf-banked terraces on An Teallach plotted against the mean intermediate axis lengths of 50 clasts sampled from the front of each terrace.

Table 9.8

Locations of trenched turf-banked terraces.

Site	Mountain	N.G.R. (NH)	Type	Altitude	Aspect	Slope
1	An Teallach	064868	horizontal	780 m	35°	20°
2	An Teallach	073867	interconnecting	720 m	50°	24°
3	An Teallach	077862	horizontal	600 m	90°	15°
4	Ben Wyvis	498721	interconnecting	760 m	9°	21°
5	An Teallach	073863	horizontal	720 m	147°	20°
6	E. Fannichs	205717	horizontal	1000 m	75°	19°
7	An Teallach	077849	oblique	720 m.	170°	34°

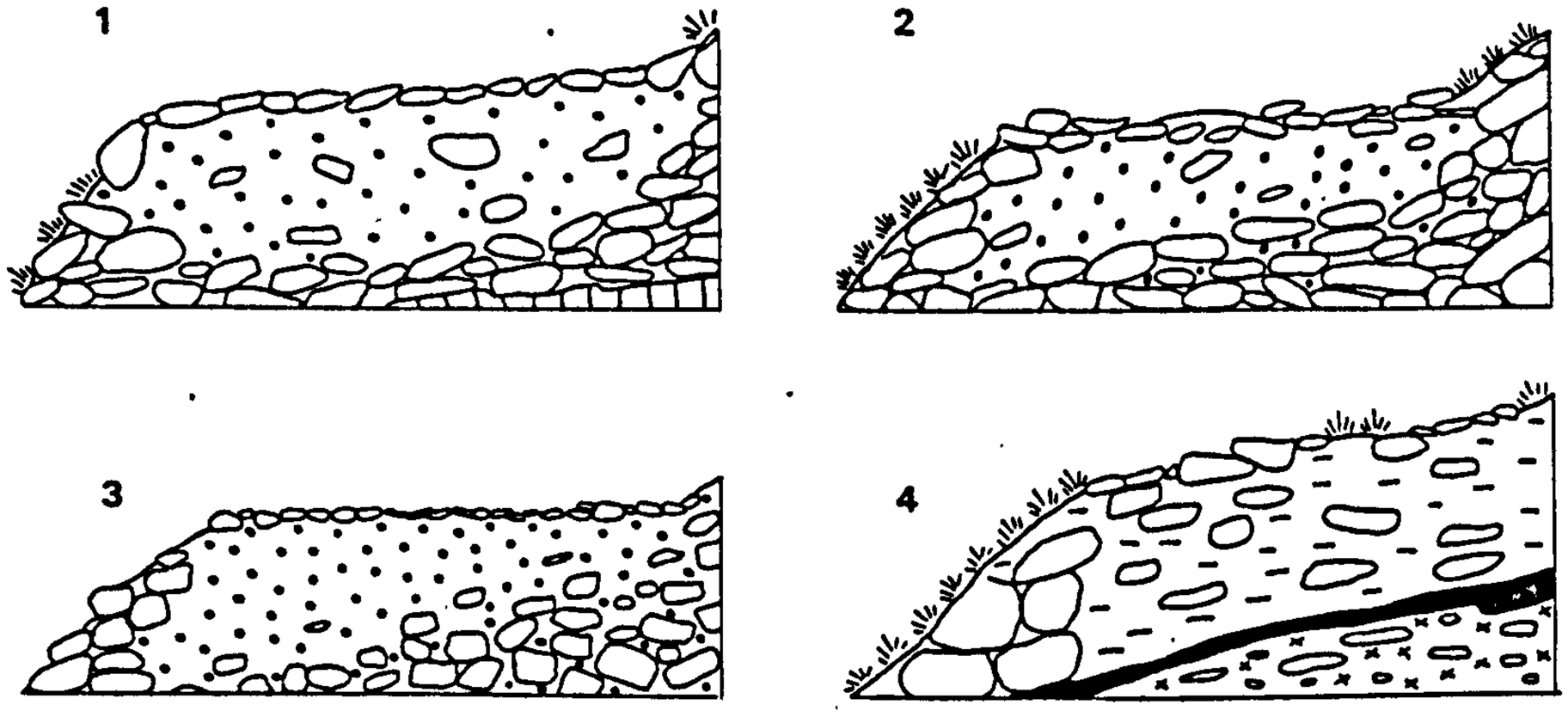


Figure 9.38: Sections cut through six turf-banked terraces (see table 9.8).

only one clast thick) underlain by a sandy zone 20-70 cm deep (in which clasts are comparatively rare) and a basal regolith of abundant clasts embedded in a sand matrix. This structure recalls that of debris surfaces on An Teallach (figure 8.24) and suggests the operation of vertical frost-sorting in the upper layers and/or deflation of fines from the terrace treads. A similar though less pronounced pattern is evident for the Fannich and Ben Wyvis examples (4, 6).

Another striking feature of some terraces (1-4) is a concentration of clasts under the riser. This may also reflect frost-sorting but is not common to all terraces; the riser of terrace 6 is almost clast-free. A final point of interest concerns the layers of organic material unearthed in terraces 4 and 5. It seems unlikely that these represent over-ridden soil horizons, as that in 5 rests directly on bedrock and that in 4 on a relatively impermeable yellow C horizon containing little organic material. The most probable explanation for this layer in both cases is that it represents the downward limit of illuviation of humic material through the friable terrace sands.

Analysis of the size, shape and fabric of samples of clasts taken from the risers, treads and interiors of terraces 1-5 and 7 revealed further interesting patterns. In all cases but one (terrace 2) the mean diameters ($\overline{(a+b+c)/3}$) of surface clasts exceeded those of clasts sampled within the terraces (figure 9.39; table 9.9). A pronounced concentration of large stones immediately above the vegetated riser is evident on many terraces and is reflected in the large mean diameters of clasts sampled from this position on terraces 1, 2, 3 and 5. On oblique terraces, the largest stones often underlie the riser, a pattern reflected by terrace 7.

Mean clast sphericity ($\overline{(a+b)/2c}$) does not apparently vary in any systematic fashion on or within individual terraces, but does vary between terraces. Terrace 3 is located on glacial deposits (the Loch Lomond Advance end moraine in Coire a'Mhuilinn) and hence is composed of much blockier material (low sphericity values) than terraces on Torridon Sandstone debris slopes. Conversely, the slabby detritus of terrace 4 (on Ben Wyvis) reflects macrogelivation

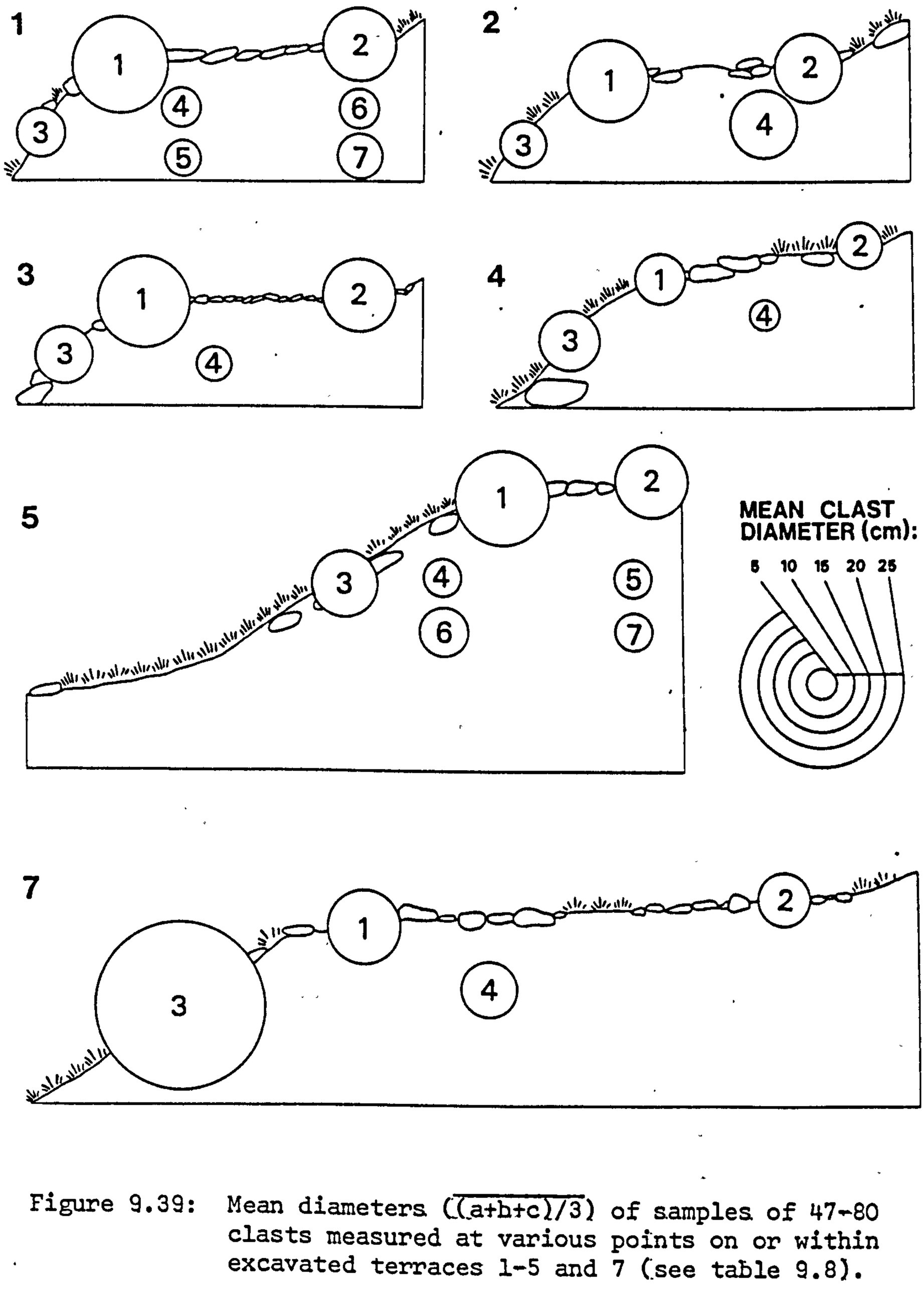


Figure 9.39: Mean diameters ($\overline{(a+b+c)/3}$) of samples of 47-80 clasts measured at various points on or within excavated terraces 1-5 and 7 (see table 9.8).

Table 9.9

Size, shape and fabric characteristics of clasts sampled on or within Turf-banked terraces

Sample	n	((a+b+c)/3)		(2000r/a)	((a+b)/2c)	A_n 360		A_n 180	
		Mean	S.D.	Mean	Mean	Score	Signif.	Score	Signif.
1.1	57	14.7	8.8	129	3.9	0.40	N.S.	3.38	.001
1.2	58	10.6	5.3	107	3.9	0.27	N.S.	0.23	N.S.
1.3	54	7.0	3.7	140	3.4	0.15	N.S.	2.98	.001
1.4	55	5.6	2.3	76	3.2	-	-	-	-
1.5	50	4.9	1.8	59	3.3	-	-	-	-
1.6	50	5.5	2.9	76	2.6	-	-	-	-
1.7	50	5.5	2.6	72	3.2	-	-	-	-
2.1	59	11.8	7.4	152	2.6	0.39	N.S.	0.66	.05
2.2	59	9.4	4.2	186	2.4	0.36	N.S.	0.68	.05
2.3	66	7.0	2.9	169	3.4	0.37	N.S.	1.19	.005
2.4	50	10.2	8.2	79	2.6	-	-	-	-
3.1	76	13.1	7.6	117	2.0	0.33	N.S.	0.49	N.S.
3.2	76	10.0	3.8	132	1.3	1.50	.001	1.83	.001
3.3	56	8.0	3.4	173	1.9	0.18	N.S.	2.49	.001
3.4	44	5.2	1.5	180	1.8	-	-	-	-
4.1	62	7.8	3.9	-	4.2	0.82	.025	1.25	.005
4.2	61	6.8	2.6	-	4.9	0.34	N.S.	1.63	.001
4.3	63	8.7	4.1	-	3.4	0.27	N.S.	0.83	.025
4.4	50	4.7	1.5	-	3.7	-	-	-	-
5.1	59	14.4	8.6	131	3.6	3.21	.001	1.45	.001
5.2	53	10.2	4.8	114	3.0	0.18	N.S.	1.90	.001
5.3	60	9.4	3.6	116	2.9	4.25	.001	1.84	.001
5.4	36	5.5	2.6	83	2.3	-	-	-	-
5.5	52	5.2	2.0	86	2.6	-	-	-	-
5.6	47	6.9	3.4	64	2.9	-	-	-	-
5.7	48	5.8	2.4	84	2.6	-	-	-	-
7.1	80	10.4	5.8	-	2.6	5.89	.001	2.05	.001
7.2	76	8.3	4.1	-	2.5	5.58	.001	2.98	.001
7.3	64	26.0	22.1	-	3.9	5.53	.001	1.68	.001
7.4	50	7.7	6.8	-	2.3	-	-	-	-

of the underlying pelitic schists. For terraces developed on debris slopes on An Teallach, mean roundness ($\overline{2000r/a}$) is invariably greater for surface than for subsurface clasts (table 9.9), which suggests that surface detritus has been subject to prolonged ~~microgelivation~~ granular disintegration and that there has been little exchange of surface and subsurface clasts in recent times. The lack of such a pattern for terrace 3 reflects the generally sub-rounded character of glacially-transported detritus.

Fabric analyses were carried out on samples of 50 clasts with a:b axis ratios exceeding 1:1.5 measured at the tread front, tread rear and riser of terraces 1-5 and 7 (table 9.9; figure 9.40). They reveal that, for horizontal terraces, clast orientation strongly reflects slope aspect, with only two of the fifteen samples not significantly different from a uniform distribution at the .05 level (A_n 180 statistic). The fabrics measured on horizontal turf-banked terraces differ from those measured on solifluction lobes, however, in that reverse (upslope) imbrication is common on the former; downslope inclination is significant at the .05 level (A_n 360) for only 4 of the 15 terraces (figure 9.30). In contrast, fabrics measured on an oblique terrace (number 7) all showed strong downslope dip and orientation.

Samples from the seven trenched terraces exhibited considerable variation in grain size (figure 9.41). Much of this variation reflects the difference in granulometry of frost-weathered regolith on Moine and Torridon rocks (compare figures 8.28 and 9.41). Terraces developed on Torridon Sandstone regolith (1, 2, 5 and 7) contain little material in the clay-silt size range ($<4\% < 63\ \mu m$), but samples taken from terraces on Moine rocks (4 and 6) have up to 14% clay-silt content. Three of the samples taken from terrace 3 (on the Coire a'Mhuilinn end moraine) have clay-silt fractions slightly in excess of 8%, presumably reflecting comminution of Torridon Sandstone fines during glacial transport. In all terraces sand (0.06-2.00 mm) is the dominant fraction, often with a large component in the medium sand (0.2-0.6 mm) range, irrespective of parent material.

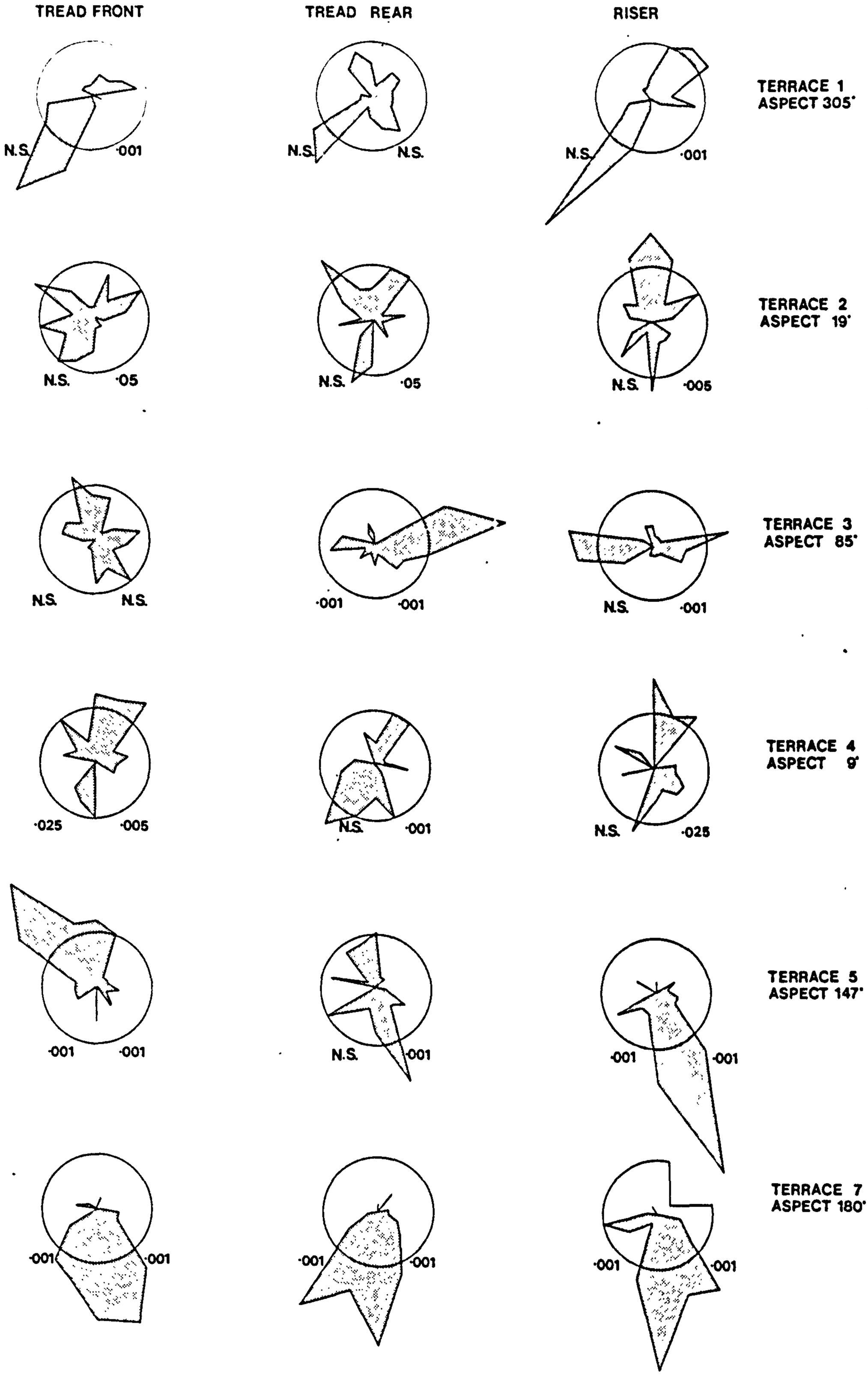
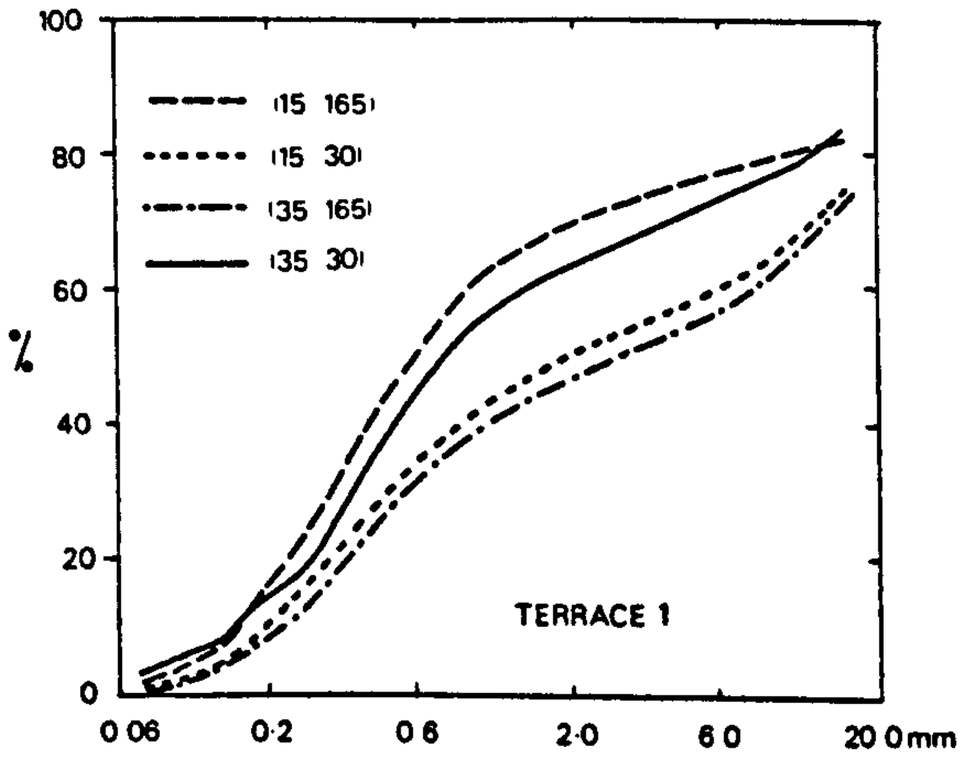
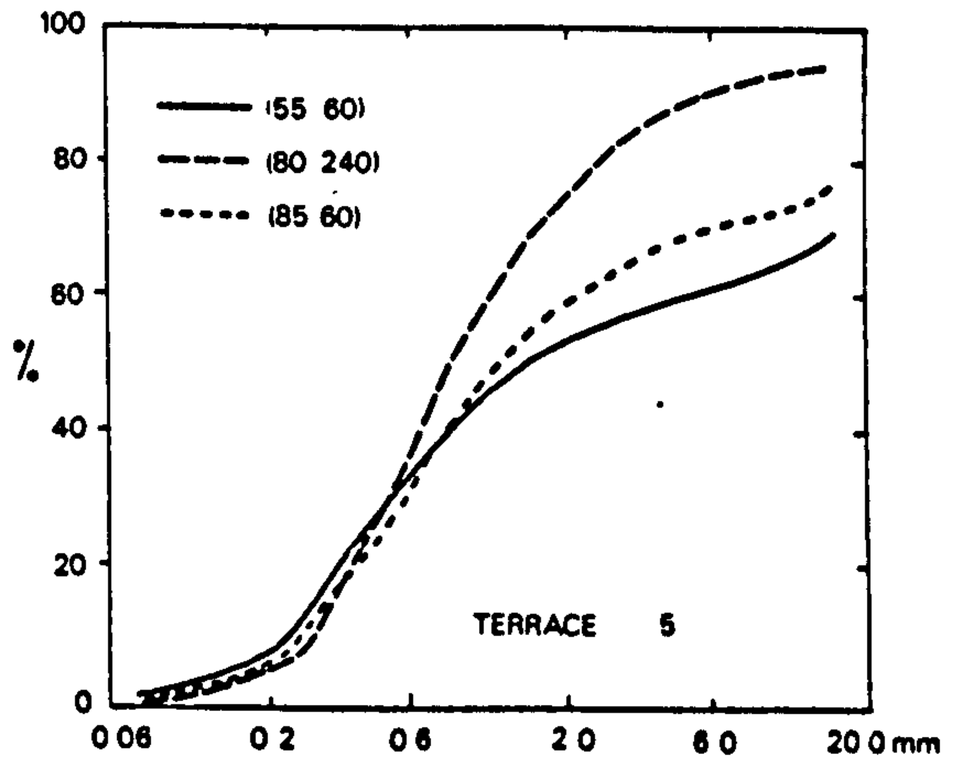
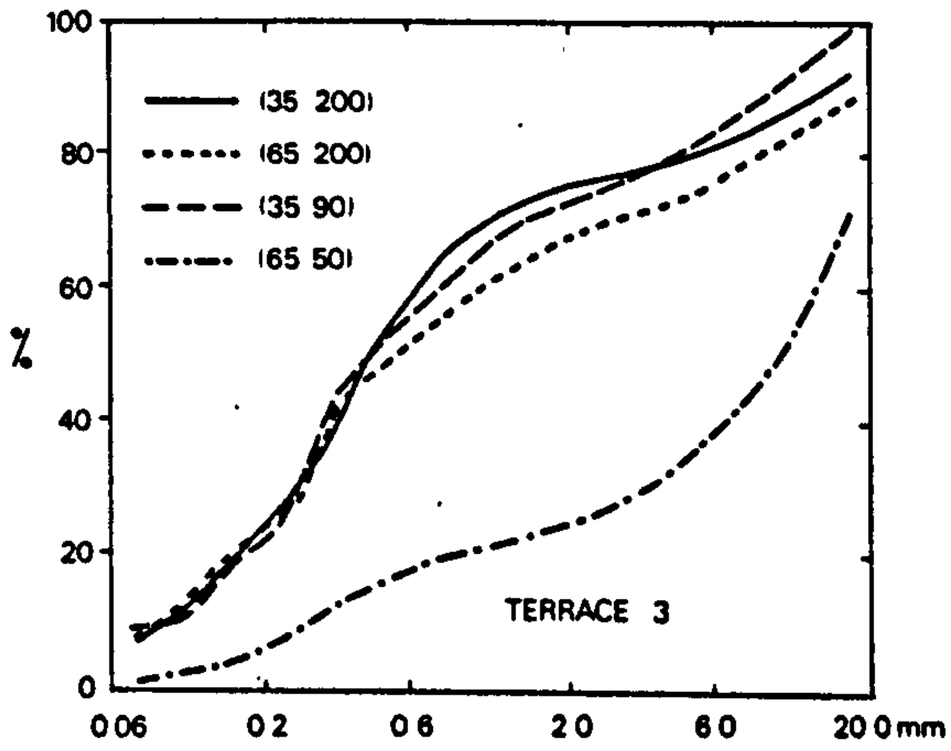
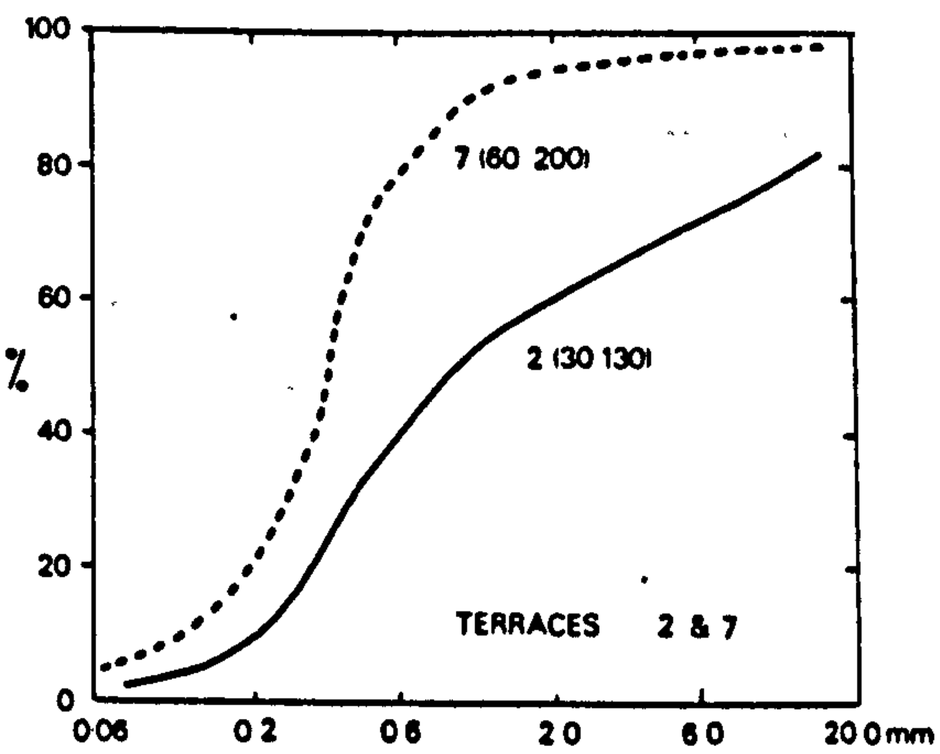


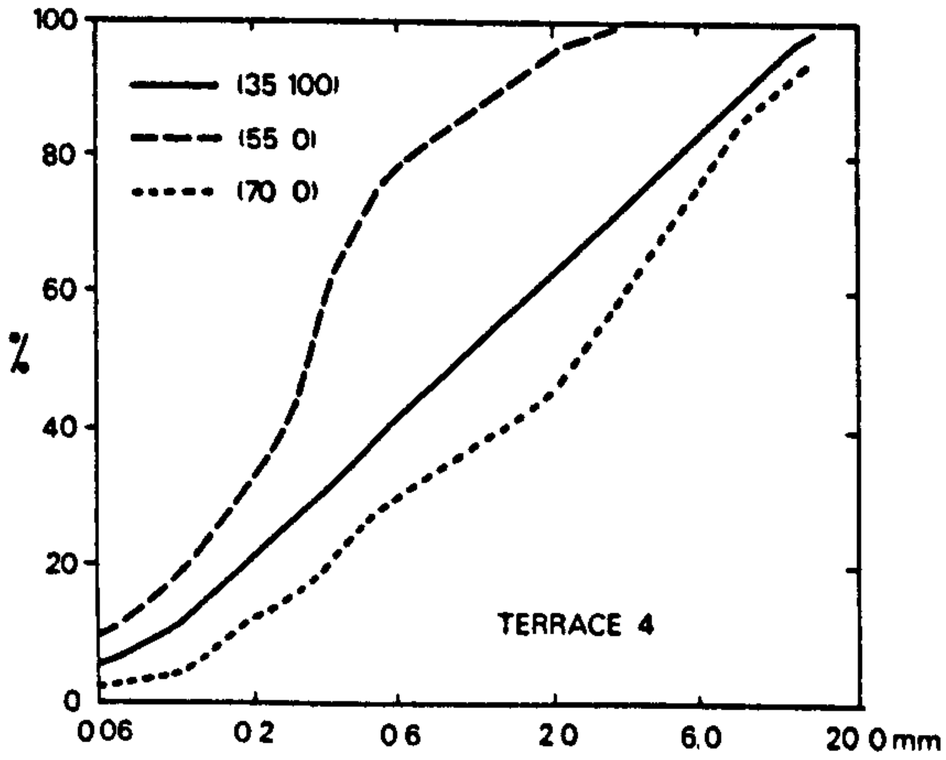
Figure 9.40: Fabrics measured on samples of 50 clasts on terraces 1-5 and 7 (see table 9.8). The figures represent significance level under the null hypothesis of circular uniformity in terms of inclination (left-hand figure; A_n360 statistic) and orientation (right-hand figure; $^nA_n180$ statistic).











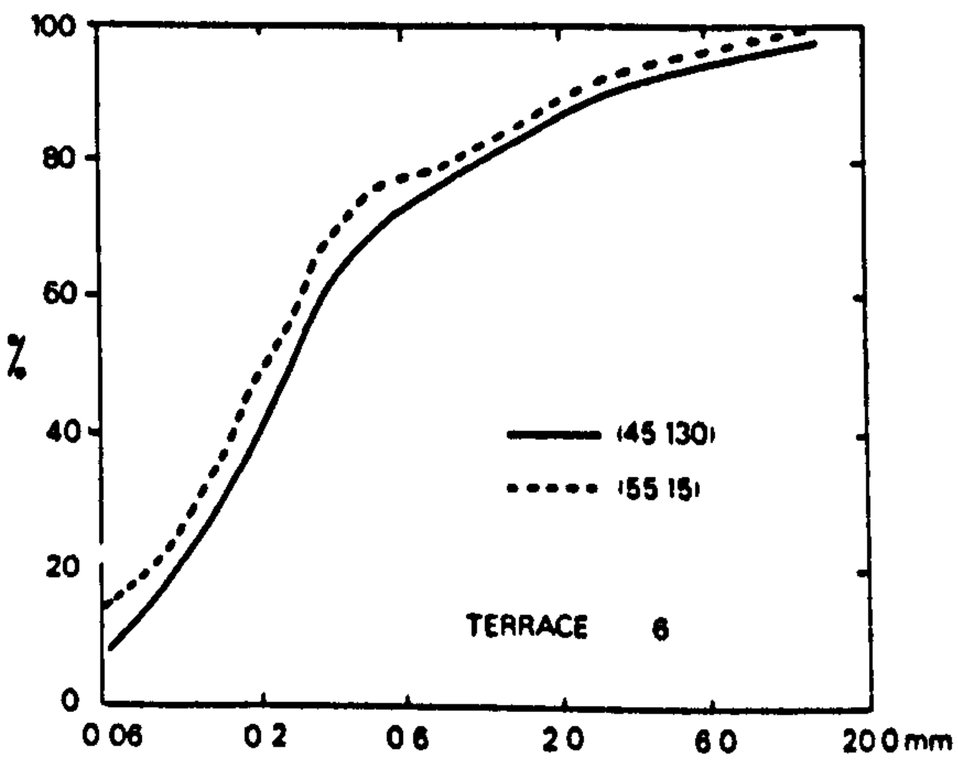


Figure 9.41: Grain-size distributions for sample of fine material collected from turf-banked terraces 1-7 (see table 9.8). The co-ordinates within brackets refer respectively to the depth and distance from the rear of the terrace of the sampling sites (both in centimetres).

The relatively high clay-silt component in terraces developed on Moine rocks suggests that these are subject to the formation of segregated ice lenses on freezing, and thus to pronounced frost heave and possibly solifluction. Lack of clay and silt in terraces developed on Torridon Sandstone, however, militates against ice lens formation and therefore solifluction, as induced suction potential (which causes the migration of water to the freezing plane) is generally low in sandy deposits (Washburn, 1979, pp. 67-8). This distinction is borne out by the presence of active solifluction features on Moine rocks, and their absence on Torridon Sandstone regolith. Also, excavations in terraces on An Teallach when the ground was frozen to a depth of 35 cm revealed interstitial ice filling pore spaces, but no ice lenses. The absence of ice segregation in Torridon Sandstone regolith probably reduces, but does not preclude, frost heave (see below).

9.3.3 Terrace distribution

The distribution of turf-banked terraces on maps 2-8 is considered in terms of six possible controls, namely regolith, slope, glacial limits, peat, altitude and aspect.

Regolith exerts a fundamental influence on terrace distribution. Terraces are entirely absent from blockslopes, where fines are absent from the upper layers of the deposit, yet are found on virtually all debris slopes except where other factors inhibit their formation. Thus the southern slope of Glas Mheall Mór (map 2), a Torridon Sandstone debris slope, supports superbly developed oblique terraces (figure 9.17) but the nearby quartzite blockslope of Glas Mheall Liath is devoid of terraces even though gradient and aspect are virtually identical. A surface or near-surface matrix of fines therefore appears a prerequisite for terrace formation.

The terrace survey data reported above give the range of slopes on which horizontal turf-banked terraces occur as 6°-31°. If shallow horizontal terraces developed across the treads of oblique terraces are included, then the range of slopes on which horizontal terraces are developed extends from 6° to about 36°, identical to that for solifluction and boulder sheets and lobes (section 9.2.4.2).

The small sample of 25 oblique terraces surveyed on Glas Mheall Mór cannot be considered representative in terms of slope, as the gradients measured are the steepest on which oblique terraces are developed. The lowest gradient on which oblique terraces occur is not known, but some around the small hill at 068858 (map 2) have formed on slopes of less than 15°.

Few turf-banked terraces are found within the limits of the Loch Lomond Advance glaciers, although their rarity in such areas is often attributable to lack of suitable slopes or regolith. In Glas Tholl (075849; map 2), however, oblique terraces cut across the former glacial limit, implying formation since the disappearance of glacier ice. In Coire a'Mhuilinn, horizontal terraces are found on the distal (east) side of the Loch Lomond Advance moraine, implying formation since the moraine was deposited. Elsewhere (e.g. 060843) rather poorly-developed horizontal terraces lie within the supposed limit, and turf-banked terraces on ground previously occupied by Loch Lomond Advance glaciers in the hills south of Glen Torridon have been described by Robinson (1977, pp. 94-5). It therefore appears that although terraces are much more common and generally better developed on slopes outside the Loch Lomond Advance limit, some terrace development took place either during the period of Loch Lomond deglaciation or during the Flandrian.

The lowest turf-banked terraces on maps 2-8 occur at 450 m on the Western Hills of Rhum, 500 m north-west of the summit of Ard Nev (map 8). Terraces are also found just below 500 m on the Trollaval-Askival col in the Rhum Cuillin (map 7). On An Teallach, terraces descend to almost 600 m but on the western Fannichs they are not found below 750 m and in the eastern Fannichs their downslope distribution is delimited by the 800 m contour. Few terraces are formed below 800 m on Ben Wyvis, although a group E.S.E. of Glas Leathad Beag (map 6) descend to 700 m. In general, however, the lower limit of terrace development appears to rise eastwards, although this trend is complicated by availability of suitable slopes and regolith. Peat is apparently inimical to terrace development, and limits the distribution of terraces in parts of the Fannichs and Ben Wyvis.

The most striking feature of terrace distribution is the relationship between terraces and aspect. Figure 9.42 depicts the aspects of the 333 horizontal terraces surveyed in the field, and reveals a very strong concentration of terraces in a semicircular range from 350° through 90° to 170°. That this concentration does not merely reflect selection of sampling sites is confirmed by the distribution of terraces on maps 2, 4, 5 and 6. Horizontal turf-banked terraces are abundant on all these maps, yet in every case are absent from slopes with a westerly aspect. The same is true for horizontal terraces on Sàil Mhór. Elsewhere, however, there are exceptions to this rule, the best examples being the west-facing terraces on the A'Chailleach-Scùrr Breac ridge (map 3) and the amazingly regular terraces on the Ruinsival ridge in the Rhum Cuillin (369947, map 7; figure 9.12). A consistent feature of such exceptions is that they run parallel to and grade into wind stripes (elongate deflation scars).

Oblique terraces are also strikingly distributed, being restricted to slopes with aspects falling between north-west and north and between south-west and south-east. On north and south to south-east facing slopes oblique and horizontal terraces sometimes coexist, giving rise to inter-connecting terraces. The transition from horizontal terraces (with easterly aspects) through inter-connecting terraces (facing north or south to south-east) to oblique terraces (with north-westerly and south to south-westerly aspects) is complete at several sites on An Teallach (e.g. 066869; 064863; 069859; 068854) and elsewhere (e.g. 149711, map 3). The significance of these relationships is discussed in section 9.3.5 below.

9.3.4 Present activity

Present rates of downslope movement of material on the treads of turf-banked terraces on An Teallach were assessed by installing mass-movement sites of the type described in the appendix across the treads of nine terraces. The results (figure 9.43) show that movement is very irregular across the tread surfaces, with individual markers moving up to 37 mm in a

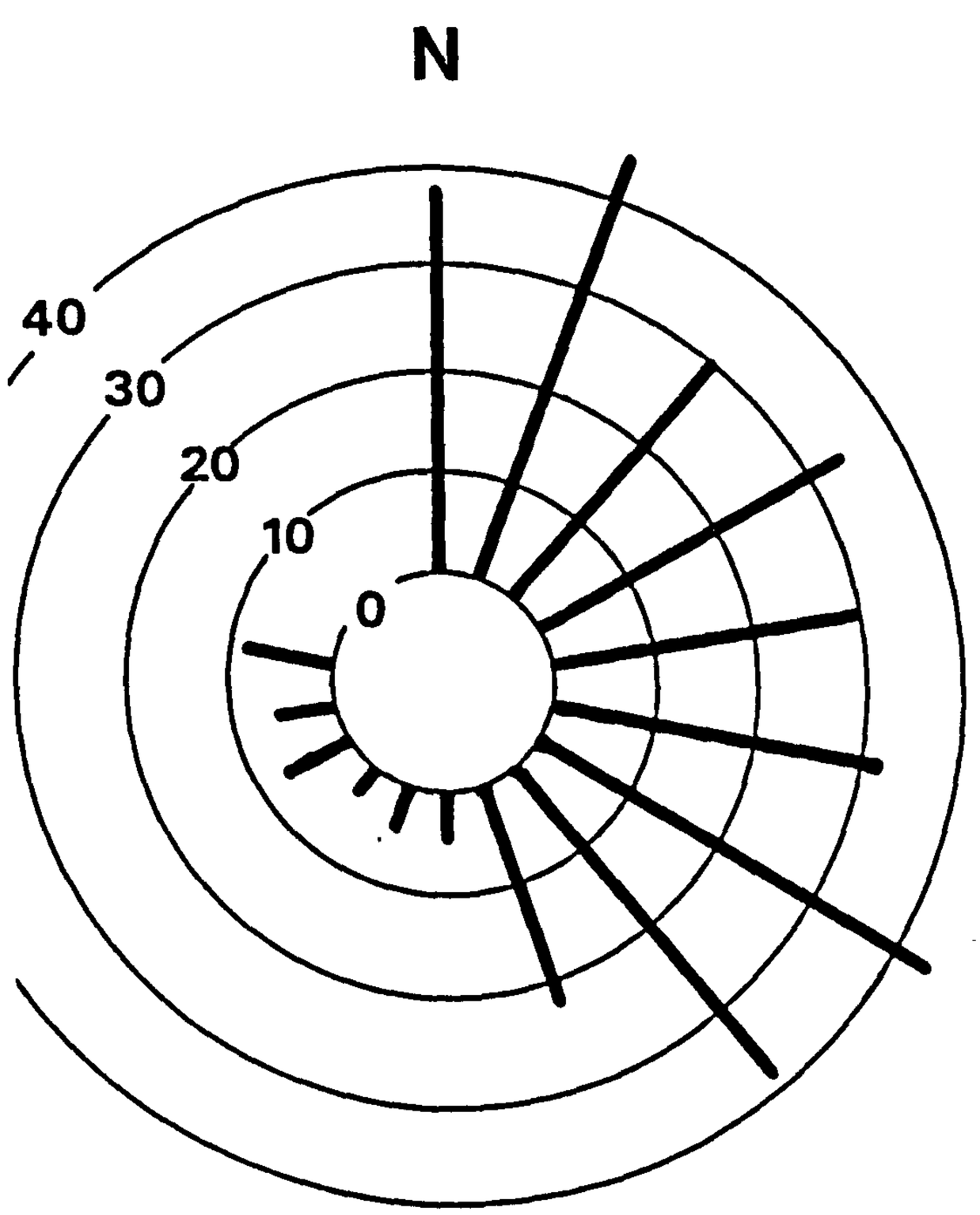


Figure 9.42: Aspects of the survey sample of 333 horizontal turf-banked terraces. Figures refer to number of observations.

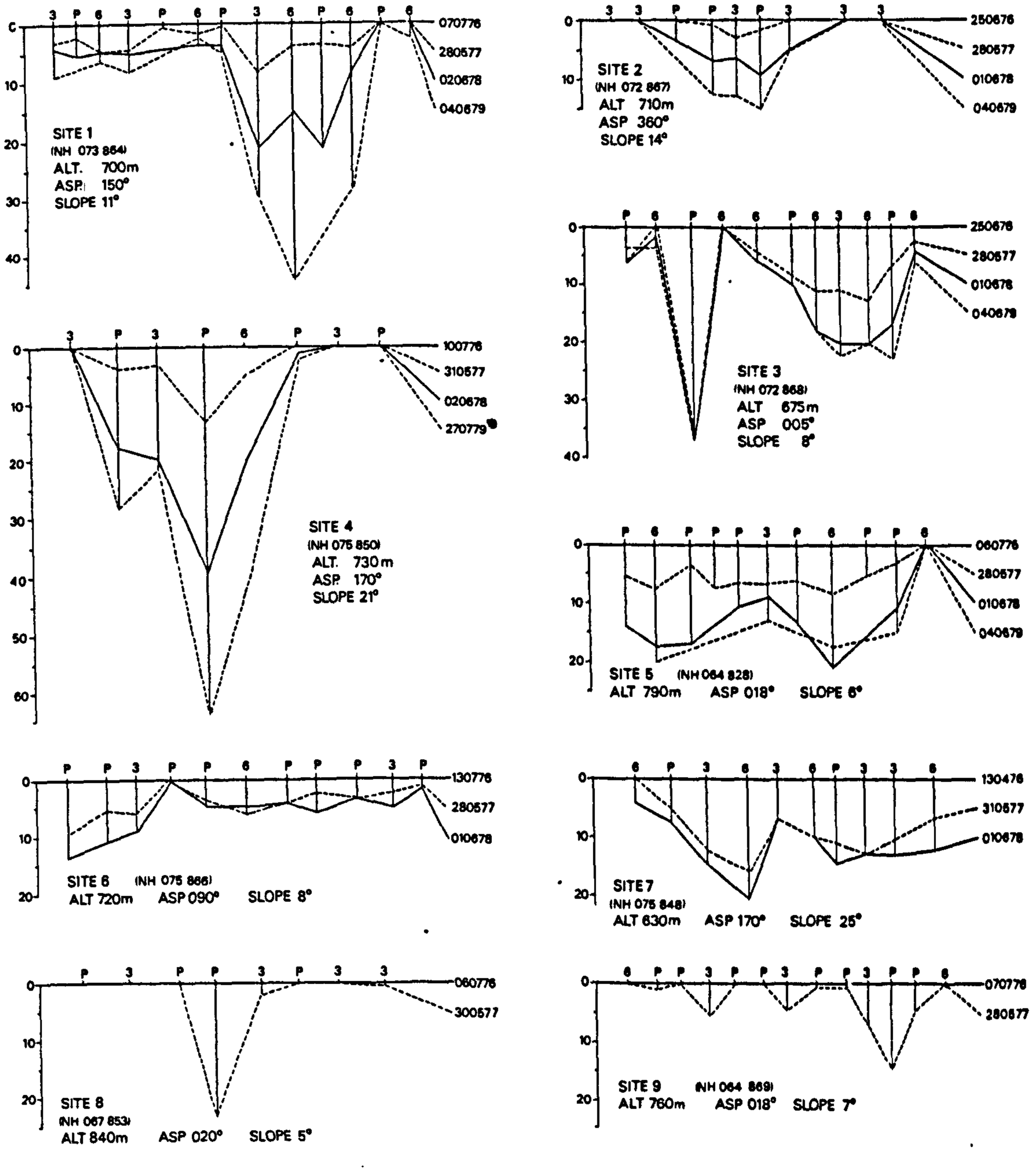


Figure 9.43: Downslope displacement of markers placed across the treads of nine turf-banked terraces on An Teallach. Sites 1 and 2 are interconnecting terraces; 4 and 7 are oblique terraces; the remainder are horizontal terraces. P - pins adhered to surface clasts; 3 - 3" nails; 6 - 6" nails (appendix 1). "Slope" refers to the downslope gradient of the terrace tread. The vertical axes represent downslope displacement in millimetres.

single year whilst adjacent markers remained stationary (site 3, 1976-7). It is instructive to compare rates of movement on turf-banked terraces with those measured on unterraced debris slopes on the same mountain (table 9.10). In both cases movement was generally greatest over the severe winter of 1978-9 and least over the winter of 1976-7. For debris slopes, the mean annual displacement of surface clasts markedly exceeds that for 3" nails, which in turn exceeds that for 6" nails, a trend that was earlier (section 8.4.5) attributed to rapid decline in displacement with depth. This pattern does not hold on terrace treads, where the mean annual displacement of all three markers is similar, probably reflecting impedence of clast movement by vegetation at the riser crest. As on unterraced debris slopes, movement is apparently the result of frost creep: measurements made of marker displacement on terraces 1, 2, 3, 5 and 9 in September 1976 and September 1977 indicated that displacement of markers over 2-3 summer months was generally zero and in no case exceeded 1.5 mm; in spring, individual clasts often appear displaced from enclosing "sockets" of sand; and needle ice was observed at shallow depth during periods of winter freezing, heaving sand particles and stones up to 3 cm normal to the tread surfaces.

Movement by frost creep is also indicated by the profiles of rubber tubes and columns of glass beads inserted vertically into the treads of two inter-connecting terraces in October 1976 and excavated in June 1979 (figure 9.44). These show an approximately exponential decline in displacement with depth, consistent with Davison's (1889) prediction that displacement by frost creep at depth z may be expressed as

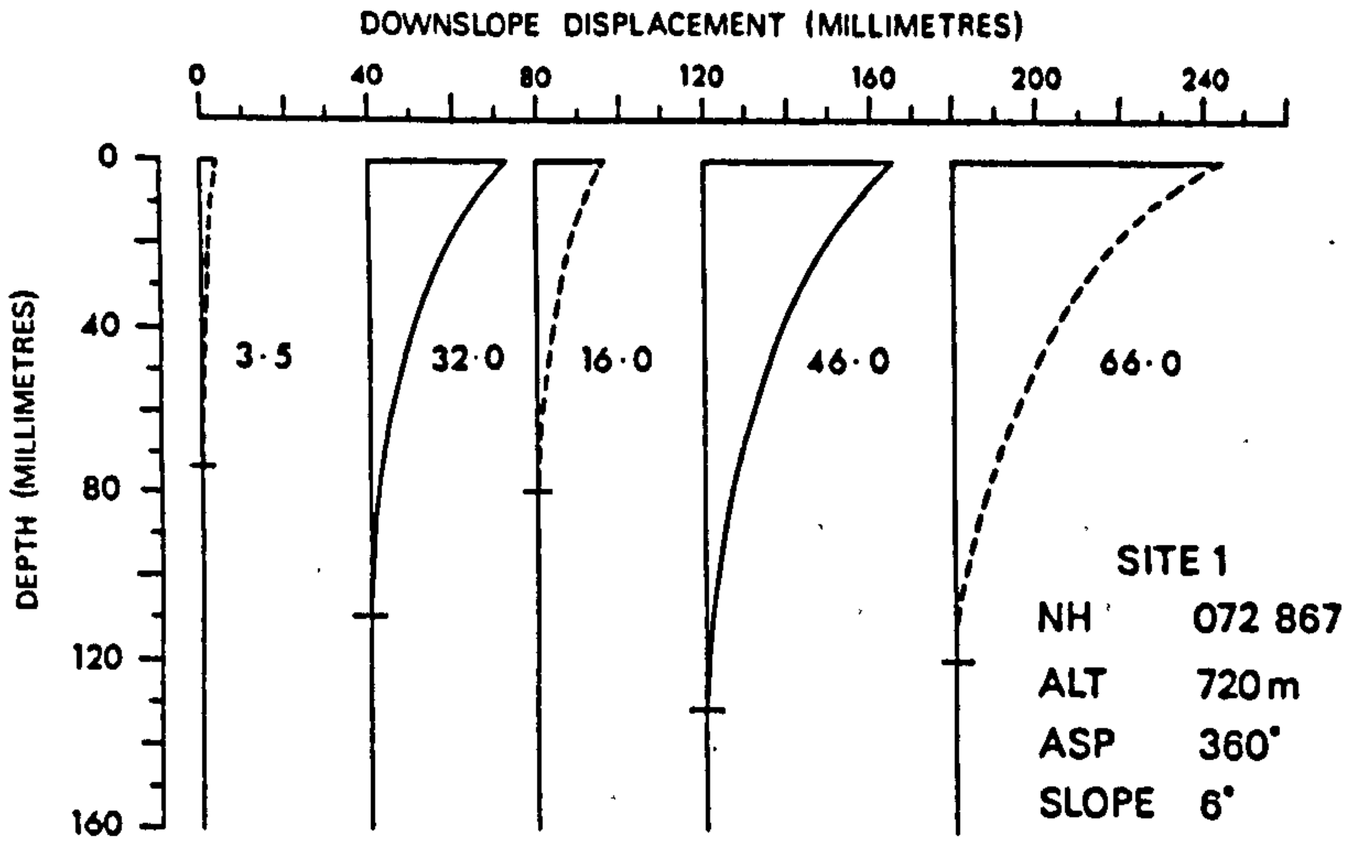
$$d_z \propto \frac{e^{-kz}}{k}$$

where k is some constant. Kirkby (1967) modified Davison's model to take into account the weight of soil overburden, but empirical observations on the form of frost-creep profiles (e.g. Benedict, 1970) indicate that the exponential form is typical, probably because the force of soil overburden (acting vertically downward)

Table 9.10

Mean downslope displacement in millimetres of surface clasts, 3" nails and 6" nails on vegetation-free debris slopes and the treads of turf-banked terraces, An Teallach.

Year	Debris slopes			Terrace treads		
	clasts	3" nails	6" nails	clasts	3" nails	6" nails
1976-7	9.8	4.0	3.1	3.9	3.0	4.3
1977-8	10.3	4.4	3.5	4.0	3.5	3.0
1978-9	11.2	5.0	3.9	4.8	2.5	4.8



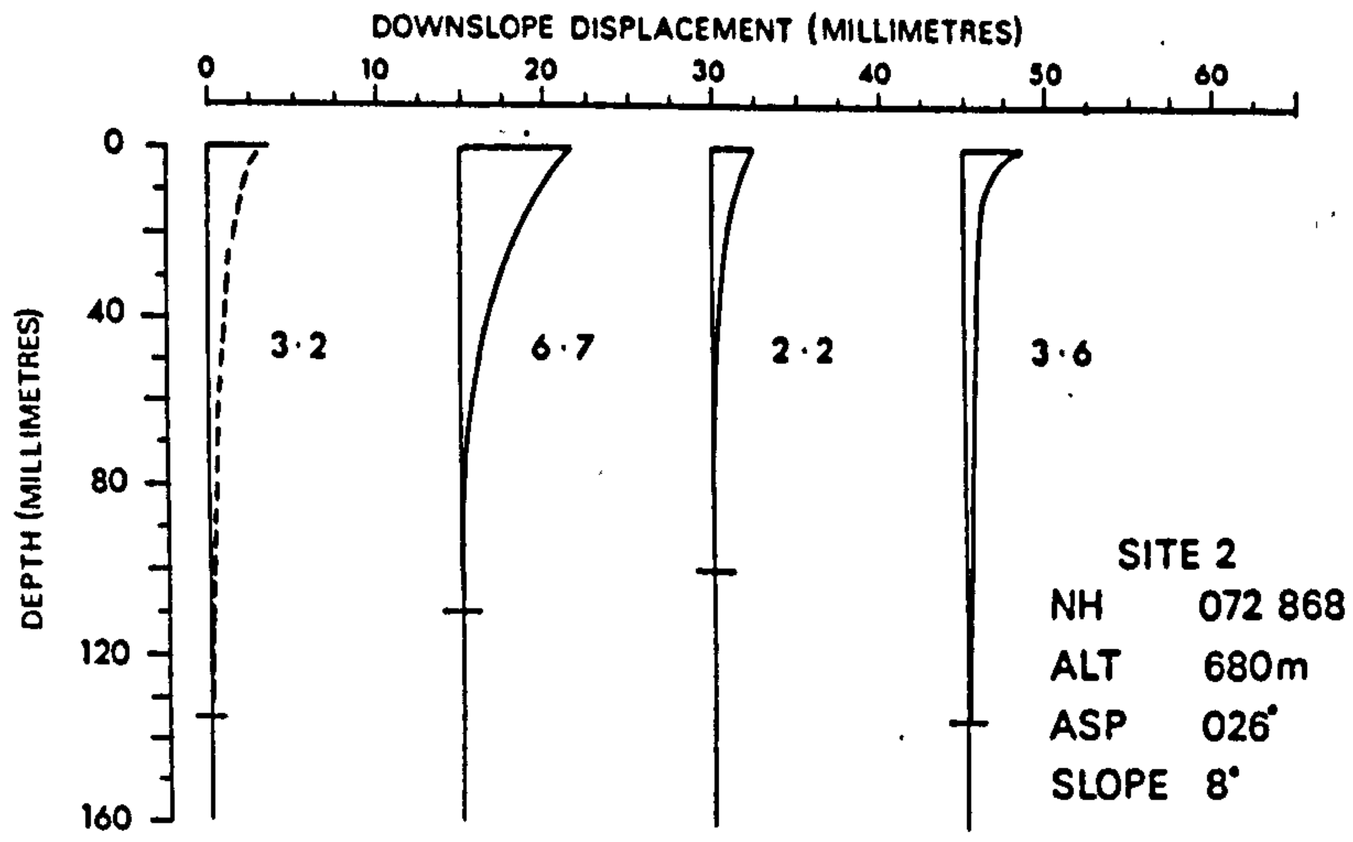


Figure 9.44: Displacement profiles of rubber tubes (continuous lines) and columns of glass beads (dashed lines) inserted vertically into the treads of two interconnecting terraces in October 1976 and e::cavated in June 1979. The horizontal dash indicates the greatest depth at which displacement was evident.

is small compared with the force generated by frost heave (acting normal to the soil surface). Davison suggested that the exponential form of such curves is the consequence of an exponential decline in the frequency of heave-effective freeze-thaw cycles with depth. On the profiles shown in figure 9.44, $d_z = 0$ where z lies in the range 7.5-14.0 mm, suggesting that frost creep is significant only in the uppermost few centimetres of soil.

The operation of frost heave over the period October 1976 to June 1979 is demonstrated by the heave of rubber tubes from the terrace treads (table 9.11). The amounts of heave shown may, however, be regarded only as minimum values of cumulative annual heave, as the tubes may have slipped back into the soil during thaw. Some individual values for heave were also recorded between 1976 and 1978 using a frost-heave frame (Gerlach, 1972, p. 66) installed across a nearby terrace tread (figure 9.45). The maximum heave recorded was 20 mm (November 1977), although most values are less than 10 mm. It should be noted that the pattern of points on figure 9.45 does not imply the operation of an annual freeze-thaw cycle, as periods of thaw separated most of the readings.

The value of these data is that they indicate that even though the weathered regolith of the An Teallach terraces is too coarse to allow ice lens formation, cumulative surface heave of up to (and possibly much more than) 7.3 cm y^{-1} occurs, resulting in the operation of frost creep in the top 10 cm or so of the soil. There can be little doubt that this is the mechanism underlying the movement of surface and near-surface debris on the treads of turf-banked terraces.

9.3.5 Formation

One of the most intriguing features of the distribution and alignment of turf-banked terraces is their relationship to dominant wind direction. Several researchers have suggested that terraces tend to be aligned either parallel to or normal to dominant wind direction (Benedict, 1970; Kelletat, 1970a; Whyte,

Table 9.11

Vertical heave of rubber tubes inserted in the treads of turf-banked terraces (in centimetres).

Period	Tube 1	Tube 2	Tube 3	Tube 4	Tube 5
1076-0677	3.0	1.2	3.8	3.4	7.3
0677-0678	3.1	1.3	2.8	3.9	5.1
0678-0679	2.5	1.7	2.5	2.9	3.8

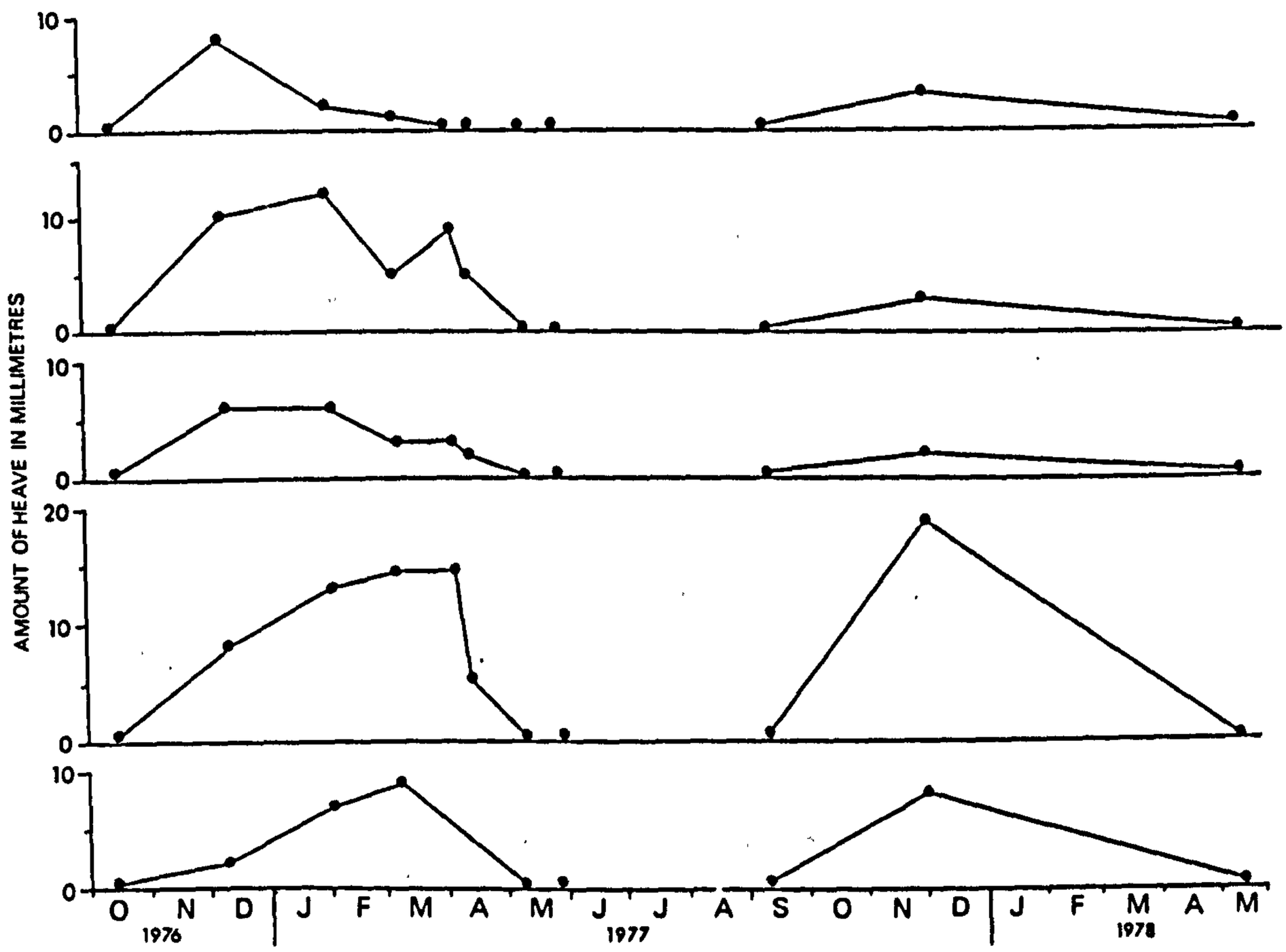


Figure 9.45: Measurements of vertical frost heave of five rods in a frame installed across the tread of a horizontal turf-banked terrace at c. 720 m altitude on An Teallach, October 1976-May 1978.

1970; King, 1971b) although as Ball and Goodier (1974; Goodier and Ball, 1975) have pointed out, the relationship is not always simple, and the alignment of oblique terraces in particular appears to reflect a complex interaction between wind and slope. Moreover, no convincing explanation of this relationship has yet been suggested, although it obviously provides a vital clue as to the mode of terrace formation.

In the Northern Highlands, the dominant winds blow from the west (section 5.6) and the majority of turf-banked terraces are found on lee slopes with an easterly aspect, aligned at right angles to dominant wind direction. Much of the vegetation on terraces on An Teallach consists of Calluna, which is a useful indicator of the direction of locally dominant winds as the stems tend to grow in a downwind direction (Crampton, 1911). The alignment of Calluna stems indicates that the dominant winds blow downslope on easterly slopes, so that terrace treads are relatively exposed to high winds but risers are sheltered. On more exposed N.E.- and S.E.-facing slopes wind stripes often cut right across the riser vegetation, again demonstrating that the terraces are aligned across the direction of dominant wind. Horizontal terraces aligned across the dominant wind direction constitute the most common terrace type in all study areas except possibly Rhum and are hereafter referred to as "normal" terraces.

Distinct from these are horizontal terraces that are aligned parallel to wind stripes, such as those on the A'Chailleach-Scùrr Breac ridge (map 3) and at 369947 (map 7). Horizontal terraces that appear to be aligned parallel to dominant wind direction (hereafter referred to as "deflation" terraces for reasons described below) differ from normal terraces in certain respects. They are usually more regular, with straight-fronted, totally-vegetated risers and flat vegetation-free treads (figure 9.12). Moreover the concentration of large stones behind the riser that is evident on most normal terraces is rarely evident on deflation terraces.

Oblique terraces invariably dip westwards on both north- and south-facing slopes and are aligned across the direction of locally dominant wind. As on normal terraces, the alignment of Calluna stems shows that dominant winds blow eastwards and slightly downwards

across such slopes, often cutting wind stripes through the risers of oblique terraces (Ball and Goodier, 1974; Goodier and Ball, 1975).

In terms of relationship with dominant winds, therefore, three types of turf-banked terrace may be distinguished: normal terraces (horizontal terraces aligned at right angles to dominant winds), deflation terraces (horizontal terraces aligned parallel to dominant winds) and oblique terraces (aligned across the dominant wind direction). For convenience, the formation of each type is discussed separately.

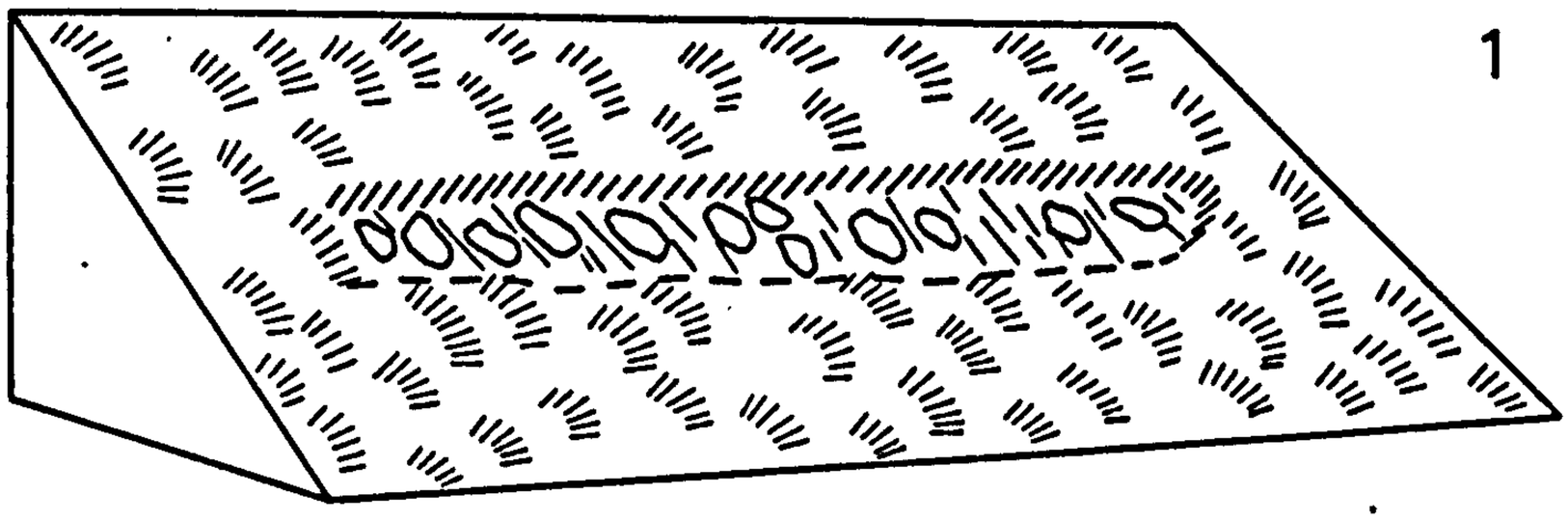
The development of deflation terraces is easiest to envisage. As noted above, such terraces are closely associated with wind stripes, and indeed grade into elongate deflation scars where slope angle drops below c. 10°. Like wind stripes they tend to be located on exposed spurs or cols through which westerly winds are funnelled. Deflation terraces are therefore not found on lee slopes, but have aspects that reflect the direction of dominant wind (often locally deflected in its passage through cols) plus or minus 90°. For example, in the col west of Scùrr Breac (151711, map 3) where the orientation of wind stripes indicates dominant winds blowing from 220°, deflation terraces on the west side of the col face E.S.E. and those on the east side face W.S.W. to west. Similarly, north-facing deflation terraces in Coire Mór, An Teallach (056846, map 2) are associated with east-west aligned wind stripes and W.N.W.-facing deflation terraces on the Ruinsival ridge, Rhum (369947, map 7) are associated with S.S.W. to N.N.E.-aligned wind stripes.

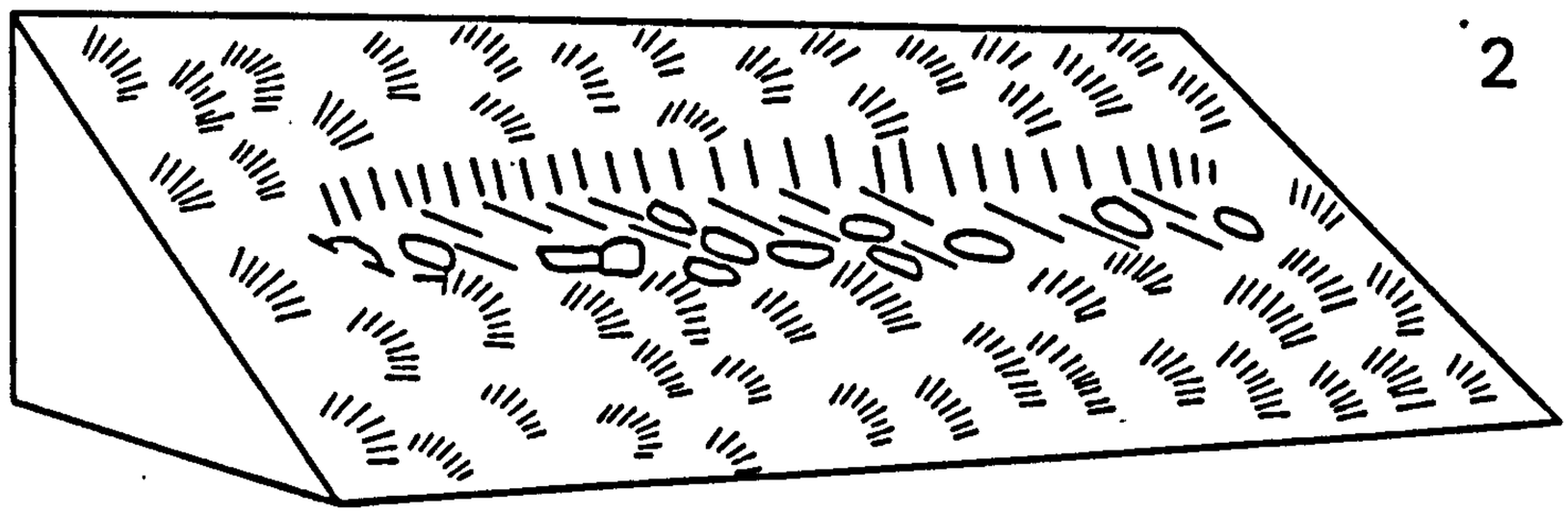
From such associations and the similarity in appearance of wind stripes and terrace treads it may be inferred that deflation terraces are modified wind stripes on slopes exceeding approximately 6°. The initiation and development of wind stripes is discussed below (chapter 10), but basically appears to involve the stripping of turf cover by a combination of frost action and deflation (King, 1971b). The production of elongate deflation scars on hillslopes exposes bare ground to the operation of surficial frost creep, which was earlier demonstrated to be the principal mechanism responsible for the downslope transport of debris across

terrace treads. However, as envisaged by Hollingworth (1934), the vegetation bordering the downslope margin of a sloping deflation scar probably traps downslope-creeping debris, which therefore accumulates at the edge of the scar. In this way (figure 9.46) a low-angle vegetation-free tread is created by downslope transfer of material across the scar, and a turf-banked terrace parallel to dominant wind direction is formed. This explanation is supported by the existence in several areas of features transitional between wind stripes and mature terraces. At two sites on An Teallach (054846 and 066864) and at one in the Western Hills of Rhum (342991) wind stripes cut across the slope have reached stage 2 of figure 9.46, in which a horizontal tread is being formed by frost creep.

Although this explanation accounts for the formation of horizontal terraces aligned parallel to dominant wind direction and the close association of such terraces and wind stripes, it cannot account for the formation of normal or oblique terraces. The latter occupy aspects similar to those on which deflation terraces are found yet are radically different in morphology and alignment and clearly represent a completely different response to the operation of wind action and frost creep.

This difference in response can be explained in terms of differences in initial vegetation cover. The explanation of deflation terraces given above assumes an initially vegetated slope. If a vegetation-free slope is assumed, then it is possible to deduce a sequence of events that leads to the formation of oblique terraces. On a vegetation-free slope the entire surface is subject to surficial frost creep under present conditions, as demonstrated in section 8.4.5. Vegetation colonization is inhibited by strong winds sweeping across such slopes, and is only likely to occur in the sheltered lee of surface irregularities or boulders. Colonizing vegetation (usually Calluna) is likely to slow or arrest the downslope creep of surface debris, which therefore accumulates immediately upslope, thereby forming shelter in which further colonization may take place (figure 9.47). As this process is repeated a bank of vegetation extends diagonally upslope, joining with others and





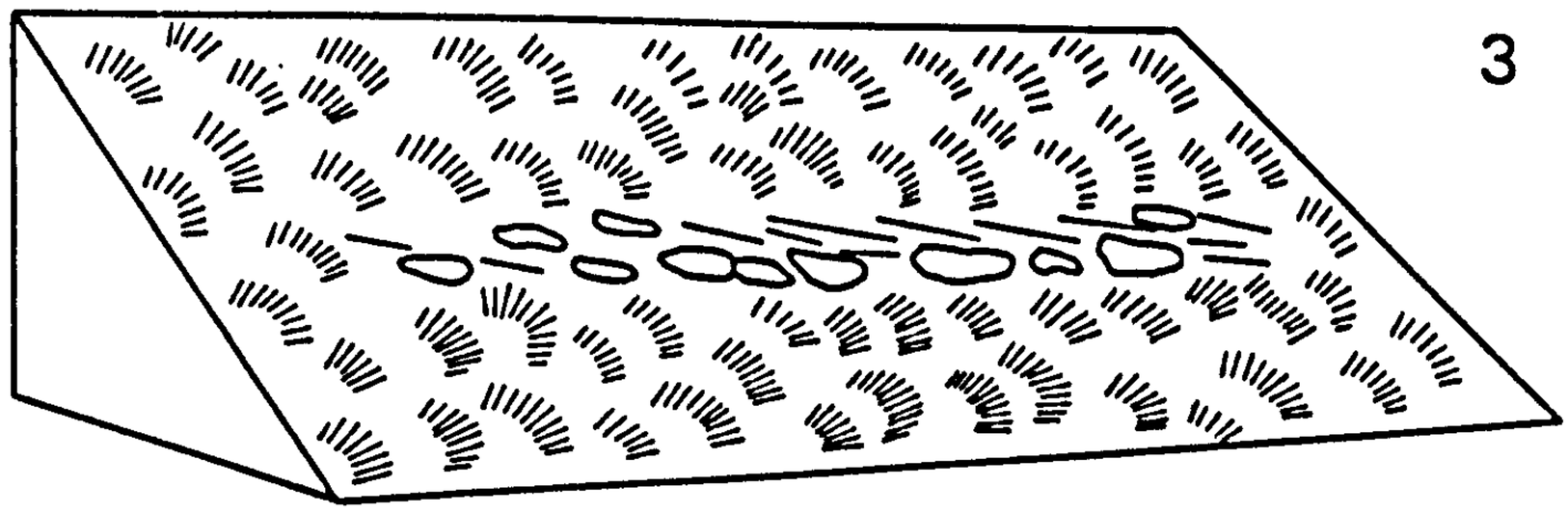
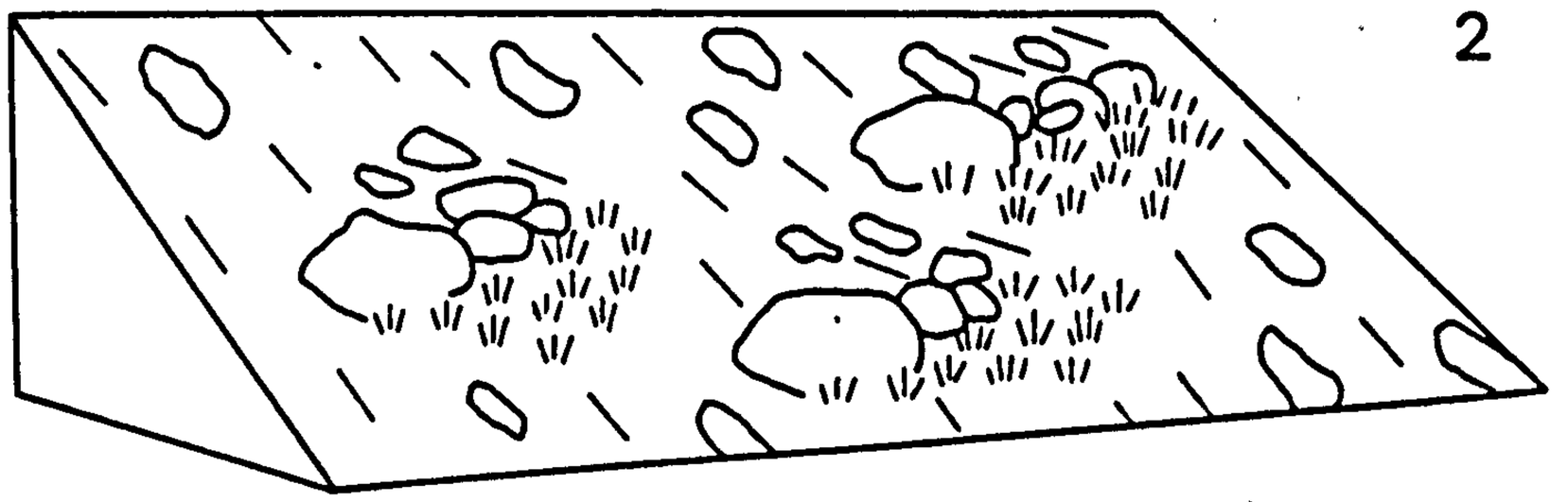


Figure 9.46: Formation of a deflation terrace.

1. Wind stripe cut across a vegetation-covered slope.
2. Frost creep transports surficial debris downslope across scar until it is trapped by vegetation.
3. Formation of a low-angle tread, across which further creep will be slow. Vegetation colonizes ground at rear of tread.





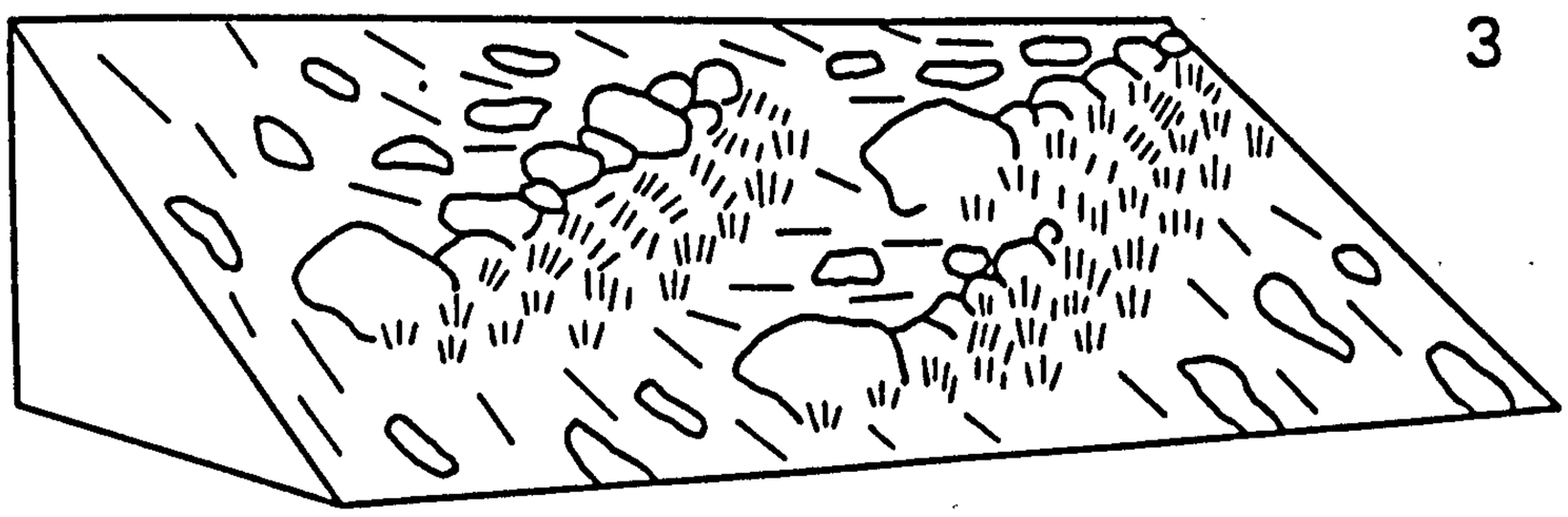


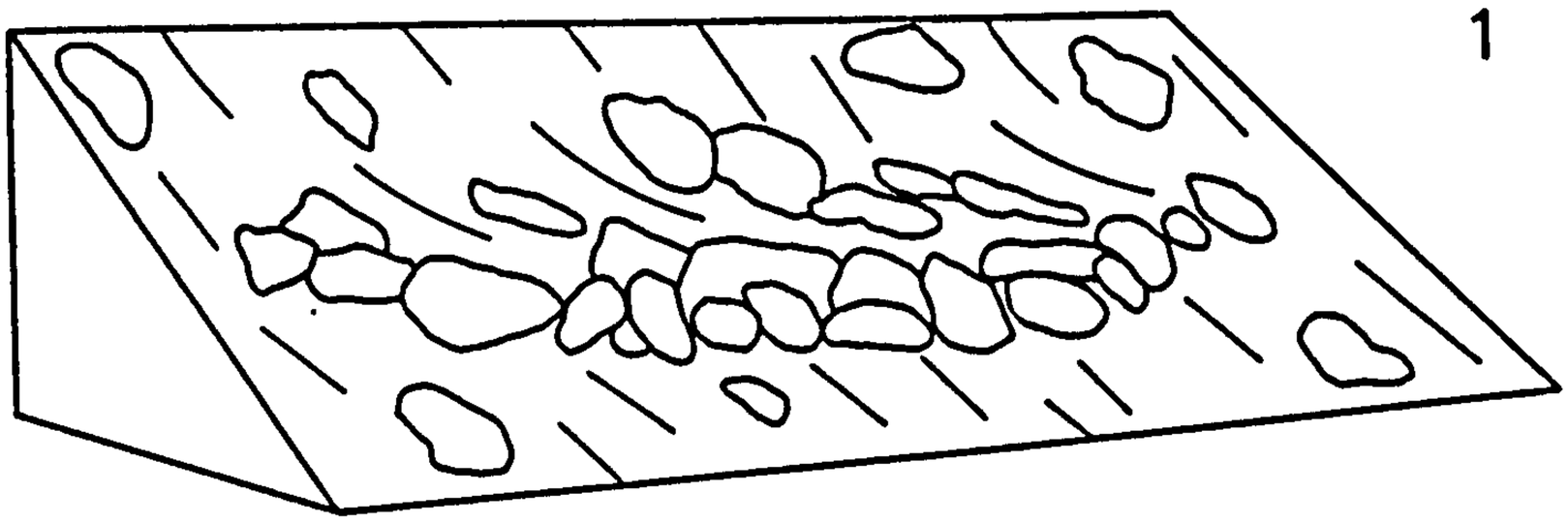
Figure 9.47: Formation of oblique terraces.

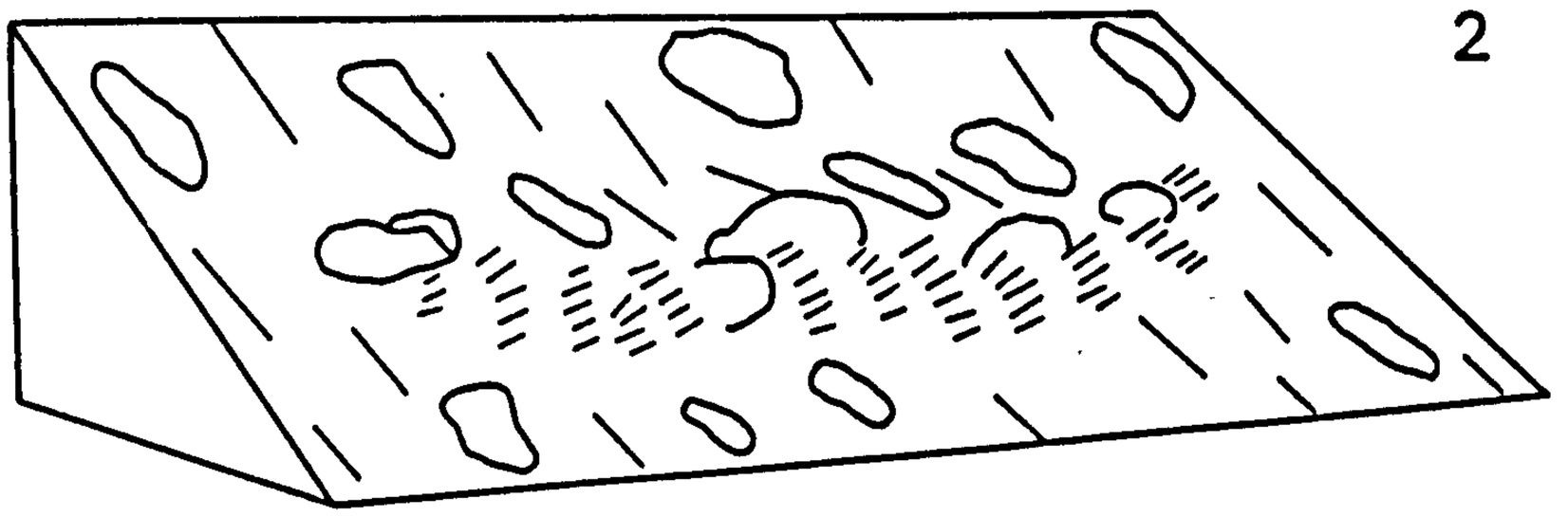
1. Vegetation colonization in lee of boulders (in response to dominant winds blowing from left to right).
2. Trapping of debris upslope of vegetation; further colonization in lee of trapped debris.
3. Progressive development of vegetated risers and build-up of treads through surficial frost creep.

through time forming the riser of an oblique terrace that is aligned across dominant wind direction. The development of oblique or deflation terraces on slopes with similar gradients and aspects may therefore be attributed to the initial presence or absence of vegetation cover on the slope. As deflation and surficial frost creep are the only processes necessary to form these terrace types it would appear that formation may take place under present conditions, and indeed both types are well developed within the limits of the Loch Lomond Advance glaciers (e.g. deflation terraces at 060845, map 2; oblique terraces at 078848, map 2).

Normal terraces pose greater problems of interpretation. As these are best developed on lee slopes over which dominant winds blow in a downslope direction (as evidenced by the orientation of Calluna stems) neither of the above explanations offers a satisfactory answer. Indeed, it is difficult to envisage the formation of horizontal turf-banked terraces on lee slopes unless the terrace form develops independently of wind direction, by mass-movement alone. Given the presence of terrace forms on lee vegetation-free debris slopes, it is easy to envisage colonization of the sheltered risers by vegetation that will tend to trap further detritus creeping down the treads, so that risers grow steadily higher and treads become progressively flatter (figure 9.48). This explanation begs two questions: is there evidence for the formation of terrace forms by mass-movement alone? and if so, when and how did such terraces develop?

Evidence for the development of terraces by mass-movement alone is found, paradoxically, on the windward western slopes of An Teallach, where at two localities (067858 and 069848) debris terraces occur. These are poorly developed compared with most turf-banked terraces, with rather steeper treads and lower (20-40 cm high) risers, both tread and riser being vegetation-free as the terraces face directly into the dominant wind. In fact the differences between these west-facing debris terraces and east-facing turf-banked terraces (the latter having generally higher risers and flatter treads and being altogether better defined) are exactly those that might be expected to follow vegetation colonization of the terrace riser. It therefore appears plausible that normal terraces represent debris terraces modified by vegetation colonization on the terrace riser.





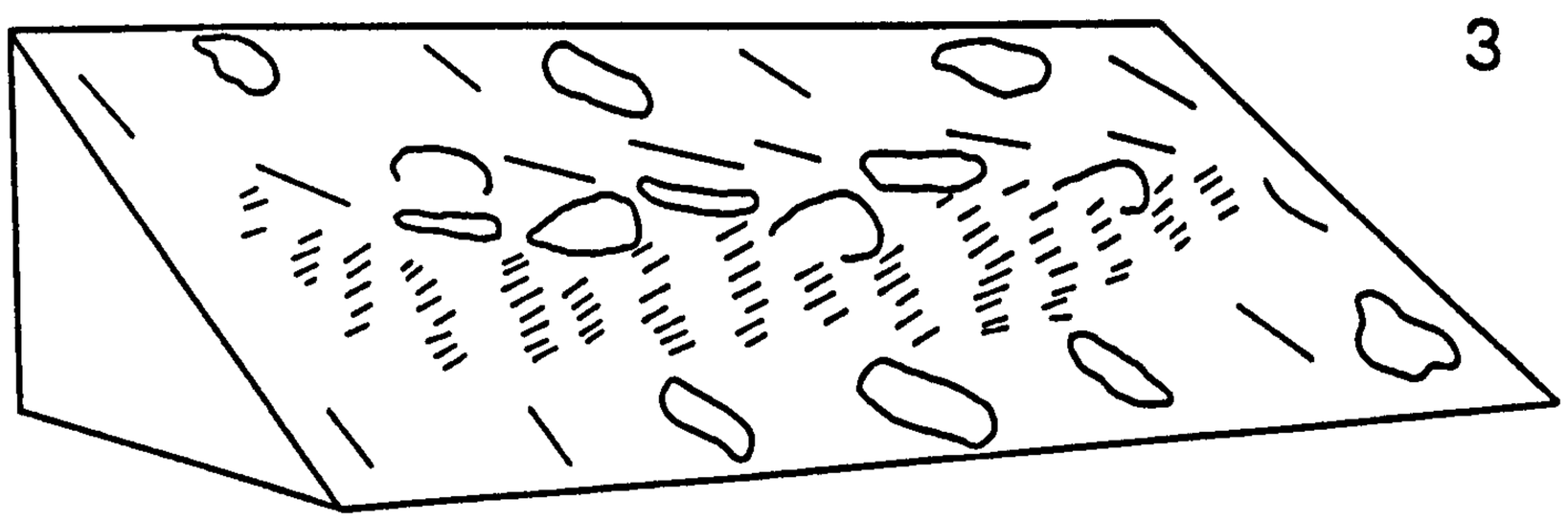


Figure 9.48: Formation of a normal terrace.

1. Debris terrace, probably of Lateglacial origin.
2. Vegetation colonization in shelter of riser.
3. Surficial frost creep causes lowering of tread gradient and build-up of riser.

The question of debris terrace formation is less easily answered. As demonstrated above, frost creep is presently effective only within a few centimetres of the ground surface and is therefore unlikely to form terraces up to 40 cm thick. The formation of such terraces would appear to require much deeper frost penetration, hence a more severe climate, which suggests that although surficial creep is active on terrace treads at present, debris and normal terraces may be of Lateglacial origin. Three pieces of evidence support this suggestion:

(i) the greater roundness of surface as opposed to subsurface clasts (table 9.9) which indicates little if any exchange of surface and subsurface clasts in recent times;

(ii) the tripartite stratigraphy of many normal terraces (figure 9.38) is similar to that of debris surfaces (section 8.4.2) and appears to reflect vertical frost sorting acting to depths of up to 70 cm; and

(iii) the absence of normal turf-banked terraces or debris terraces within the Loch Lomond Advance limits.

The last point is particularly significant. As illustrated above, both deflation and oblique terraces occur inside the Loch Lomond Advance limits, and must therefore have formed since the disappearance of the last glaciers. The author knows of no site, however, where normal or debris terraces occur inside these limits (even though normal terraces are found on the distal side of the Coire a'Mhuilinn and moraine) which suggests that their formation took place during the Loch Lomond Stadial even though they are still superficially "active" at present. The evidence for vertical sorting in the terrace profiles suggests that frost creep was the main formative mechanism.

9.3.6 Discussion

The success of the three explanations offered above may be judged by the extent to which they account for the characteristics of turf-banked terraces described in section 9.3.2. It should be

noted that the sample of 333 surveyed horizontal terraces includes both normal and deflation types, though the former predominate.

The existence of two morphologically distinct terrace types (horizontal and oblique) is accounted for by invoking different modes of formation for each type. Similarly, the phenomenon of inter-connecting terraces may be explained in terms of the Flandrian development of oblique terraces across normal terraces of essentially Lateglacial origin. The presence of lobes at the front of normal terraces may be attributed to their origin through frost creep. The gentleness of tread angles appears to result in all cases from the accumulation of creeping surface debris behind vegetation at the tread front so that horizontal terraces have a much more step-like profile than solifluction features on steep slopes yet resemble solifluction sheets on low angle slopes. All the explanations offered explain the contrast in vegetation cover between the treads and the risers, either in terms of wind-stripping (deflation terraces) or shelter (normal and oblique terraces). The strong relationships between the morphology of horizontal terraces and slope are explicable in terms of the low tread angles produced by surficial frost creep: terraces on steep slopes tend to develop higher risers and narrower treads (figure 9.35). The modification of debris terraces through vegetation colonization of risers explains the positive relationships between terrace size and transformed aspect (with origin 270^{o}) for the sample of 108 normal terraces on An Teallach (table 9.7). The interpretation of normal terraces as modified debris terraces also explains the relationship obtained between terrace size and clast size at the terrace front in that larger clasts presumably allow the development of higher, more stable risers, as demonstrated for more massive frost creep features (boulder lobes) by Shaw (1977).

The explanations offered also successfully account for the sedimentological and distributional characteristics of turf-banked terraces. The crude tripartite stratigraphy within some normal terraces (figure 9.38) is attributable to vertical frost sorting

during the Loch Lomond Stadial, and the absence of such sorting during the Flandrian explains the roundness of surface as opposed to sub-surface clasts. The concentration of large clasts under the risers of normal terraces may indicate either that the original debris terraces possessed a stony riser or may result from the trapping of larger clasts behind a vegetated riser. The grain size composition of terraces allows for the possibility of solifluction in some cases (terraces 3, 4 and 6, figure 9.41) but indicates, along with the evidence presented in section 9.3.4, that frost creep is the dominant formative mechanism. Finally, the three proposed explanations of terrace formation accord with observed aspects of terraces and the relationships of different types to the limits of Loch Lomond glaciation. As the distribution of all types of turf-banked terrace is related to dominant winds, the observed westerly drop in the altitude of the lowest terraces can be explained by increasing exposure.

The three models of terrace formation outlined above are therefore considered to offer a satisfactory general explanation of the formation of turf-banked terraces. In several areas, however, vegetation-covered terraces similar in size and morphology to turf-banked types are found. These are sometimes horizontal (e.g. 467702, map 5; 500711, map 6) sometimes oblique (e.g. 472663, map 5; 497721, map 6) and sometimes inter-connecting. As these types occupy aspects typical of turf-banked equivalents, vegetation-covered terraces are interpreted as "fossilized" turf-banked terraces, immobilized by the development of vegetation cover on the terrace treads. The distribution of vegetation-covered terraces is restricted to the Fannichs and Ben Wyvis, probably because the coarse sandy regolith of An Teallach is less favourable for vegetation colonization on exposed surfaces.

9.4 Ploughing Boulders

9.4.1 Introduction

Ploughing boulders, often referred to as ploughing blocks or gliding boulders and defined in chapter 7 as "boulders located

at the downslope end of furrow-like depressions" are amongst the most common manifestations of present periglacial slope activity in upland Britain (figure 9.49). First noted in the Lake District by Hollingworth (1934) and Hay (1937, 1942), ploughing boulders have subsequently been described in many mountain areas, including Ben Wyvis (Galloway, 1958), the Mamores (Whyte, 1970), the Cairngorms (Galloway, 1958; King, 1968; Sugden, 1970b), Lochnagar (Shaw, 1977), the Southern Uplands (Galloway, 1958, 1961a; Tivy, 1962; Ragg and Bibby, 1966), northern England (Johnson and Dunham, 1963; Tufnell, 1966, 1969, 1972, 1976) and Wales (Goodier and Ball, 1969; Ball and Goodier, 1970). In his survey of active periglacial features in the Scottish Highlands, Kelletat (1970a) observed "wanderblöcke" on most of the mountains he visited, including Ben Lawers, the Drumochter Hills, Ben Klibreck and Ben More (Mull) and the present author has observed them in Argyll (Ben Starav, Beinn Achaladair) and in the south-east Grampians (Ben a'Ghlo).

Outside Great Britain, most descriptions of ploughing boulders relate to the Alps or Pyrenees (e.g. Poser, 1954; Schmid, 1958; Stingl, 1969) where they have been assumed to represent the lowest limit of present-day periglacial slope activity (Furrer, 1965, a, b; Höllermann, 1964, 1967). Despite the apparently widespread distribution of ploughing boulders in Alpine environments, however, little research has been carried out on the nature, rate and significance of boulder movement, and these features receive undeservingly scant mention in textbooks of periglacial geomorphology; only Embleton and King (1975, pp. 118-9), French (1976, p. 189) and Washburn (1979, p. 223-4) even mention ploughing boulders.

Although most of the British references cited above make only brief mention of ploughing boulders, detailed studies of their characteristics have been made by Tufnell (1972, 1976) and Shaw (1977). The former summarized and classified the morphological characteristics of a large sample of ploughing boulders in northern England, recorded the movement of five blocks over a ten-year period, discussed distribution and terminology at length and suggested possible causes of boulder movement. The latter also investigated the morphology and distribution of a large sample



Figure 9.49: Two ploughing boulders on a gradient of c. 15° at 940 m altitude on the S.W. shoulder of Scurr Breac, western Fannichs.

of ploughing boulders (on Lochnagar) and measured the movement of twelve boulders over three or four years, but like Tufnell reached no firm conclusions as to the mechanism of boulder movement. Elucidation of the nature of ploughing boulder movement therefore constitutes the prime aim of the research reported below.

9.4.2 Characteristics

Figure 9.50 summarizes some of the main morphological characteristics of 151 ploughing boulders, 75 of which were sampled from the Eastern Fannichs (59 from the south-west slope of Scùrr Breac alone), 50 from the north-west slope of Ben Wyvis and 26 from Scùrr na Clach Geala (map 4). Furrow depths were not measured at the last-mentioned, so that for this variable sample size is 125.

Boulder size varies greatly: the largest boulder recorded was 301 cm long, 161 cm across and 91 cm above the ground surface; the smallest was 32 cm by 27 cm and only 13 cm above the surface. The range of boulder sizes is similar to that given by Shaw (1977) for Lochnagar, although the ploughing boulders measured are generally larger than those measured by Tufnell (1972) in northern England ($\bar{b}$ = 49.5 cm) yet smaller than those measured by the same author in the Alps ($\bar{b}$ = 116 cm). It appears that the size distribution of ploughing boulders is largely dependent on the size distribution of available boulders on suitable slopes. On the Fannichs and Ben Wyvis most boulders with intermediate axes exceeding 30-40 cm have produced furrows, but most smaller boulders have not, which suggests that "ploughing" requires that the boulders involved exceed a critical mass or volume.

Maximum furrow lengths measured by Tufnell and Shaw were 8.0 m and 8.9 m respectively, and although values of 10.1 m and 18.9 m were recorded in the Fannichs, the distribution of furrow lengths sampled in the Northern Highlands resembles those sampled in northern England and Lochnagar in that all three samples possess a strong positive skew, reflecting the rarity of long furrows. Furrow depth (measured 10 cm upslope of each boulder) is more normally distributed, with >91% of furrows having depths of 10-35 cm, occasional larger values being associated with large boulders.

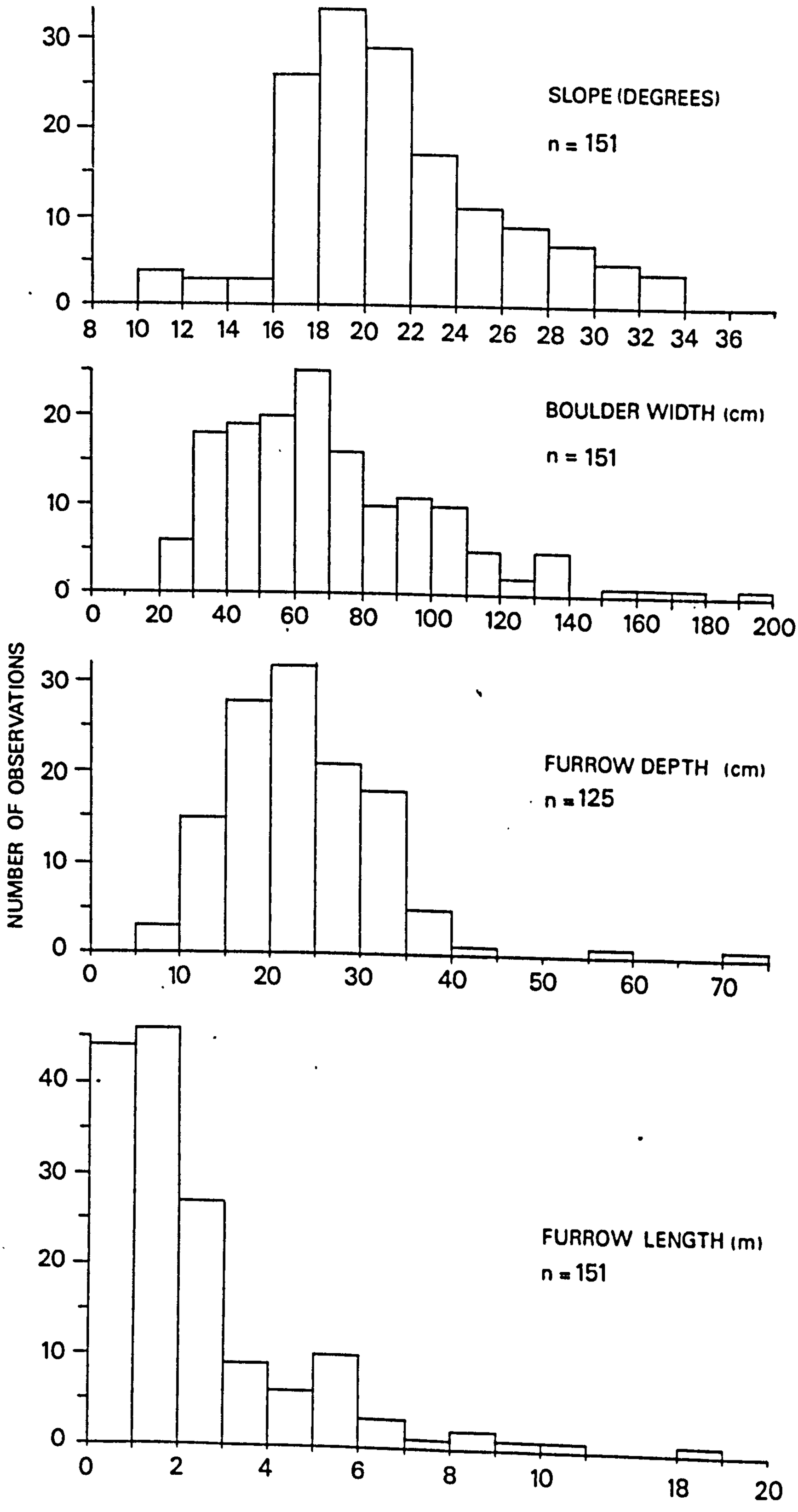


Figure 9.50: Frequency distributions summarizing some of the morphological characteristics of ploughing boulders surveyed on the Fannichs and Ben Wyvis.

Upslope of many boulders there is a much deeper niche a few centimetres wide in a downslope direction. This often extends to the base of the boulder and, unlike the furrows (which usually support a total vegetation cover) is unvegetated. The presence of such a niche probably indicates present activity.

Excavation of the ground around five ploughing boulders in the Fannichs revealed that four were embedded in the soil to a depth of 35-60 cm, so that in each case over 70% of the boulder lay below ground level. The fifth, a large slab, was embedded to a depth of 20 cm. In all cases the enclosing soil was structureless, azonal and fairly stony, closely resembling the homogeneous brown soils of active solifluction lobes (figure 9.22).

As might be expected, furrow orientation usually lies within a few degrees of general slope aspect (figure 9.51), although occasionally deflected by as much as 20^{o} as a result of local slope irregularities. Boulder orientation measured along the principal axis shows a strong downslope alignment with a very small secondary mode across-slope.

9.4.3 Morphometric relationships

It is arguable (Shaw, 1977) that furrow length provides a rough measure of rate of boulder movement or at least of the difference in the rate between boulder movement and movement of the surrounding regolith (which tends to infill the furrow). Whyte (1970) and Shaw (1977) attempted to relate furrow length to factors likely to control rates of boulder movement, such as boulder size, slope, altitude and aspect, but with little success. The apparent lack of relationships does not necessarily imply that boulder size, slope, aspect and orientation do not affect rates of activity, as factors not taken into account (e.g. depth to which boulders are embedded, granulometry and rate of regolith movement) may obscure the influence of the controls considered. For the sample of 151 ploughing boulders surveyed in the Northern Highlands, no meaningful relationships were found between furrow length (ℓ_f) or depth (d_f) and slope aspect or altitude, but weak to moderate nonlinear correlations exist between these variables and variables

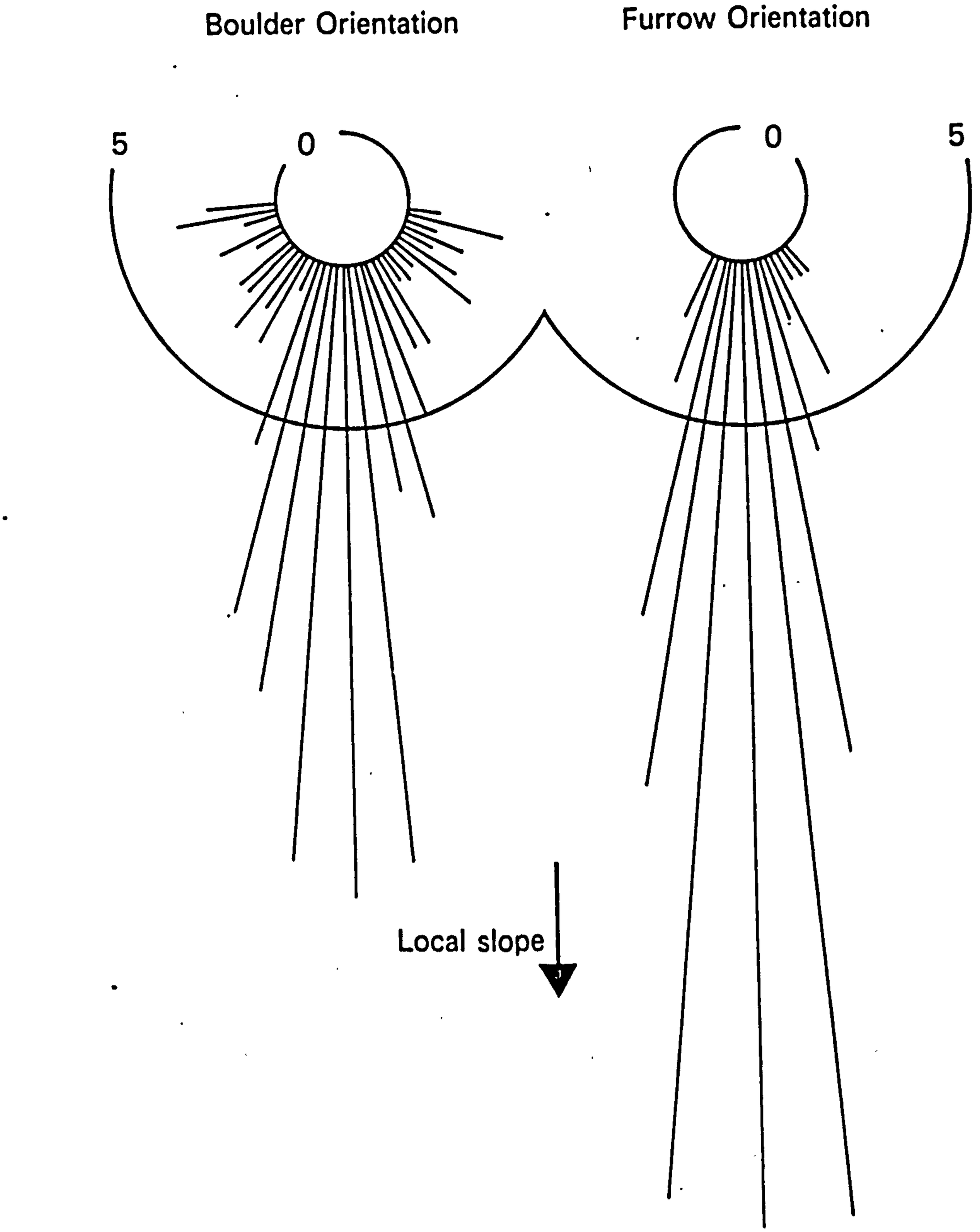


Figure 9.51: The orientation of ploughing boulders and their associated furrows with respect to local slope. Sample size = 151.

indicative of boulder size, namely length (a), width (b) and maximum height above the ground (h) (figure 9.52).

The simplest explanation of these relationships is that the widest and deepest boulders leave the best developed furrows. Wide, deep furrows are more likely to survive encroachment and therefore to be preserved for a greater distance upslope. Alternatively, larger boulders may move generally faster downslope than smaller boulders, giving longer furrows. Three points of evidence support the latter explanation. First, it was noted above that ploughing appears to require that boulders exceed some critical mass or volume, and it might be expected that the greater the mass or volume of a boulder, the more effective the ploughing mechanism. Secondly, if preservation of the furrow is the main consideration affecting furrow length, it would be expected that the strongest relationships would be between ℓ_f and b and possibly ℓ_f and h, yet in fact the strongest relationship is between ℓ_f and a, suggesting that boulder mass rather than width or depth is the most important control. Finally, although the relationship obtained between ℓ_f and d_f is statistically significant at the .05 level, it is much weaker than those obtained between ℓ_f and the size variables, again suggesting that furrow preservation alone is not the only factor influencing furrow length.

9.4.4 Activity

Present downslope movement of ploughing boulders in the study areas was monitored by setting up sites of the type described by Tufnell (1972, pp. 256-8) on a south-facing slope at 800 m on Scùrr Breac (165710, map 3). Three sites were set up in August 1976 and four in September 1977. The characteristics of the seven monitored boulders are given in table 9.12, along with displacements measured in September 1977 and September 1979.

Mean annual rates of movement for the two or three year measurement period are highly variable, ranging from 6 mm y^{-1} to 34.5 mm y^{-1}. This range is intermediate between the rather low rates recorded by Shaw (1977) on Lochnagar (0.3-8.7 mm y^{-1}) and the generally faster movement measured by Tufnell (1976) on the Pennines (0.4-64 mm y^{-1}). The most interesting feature of the

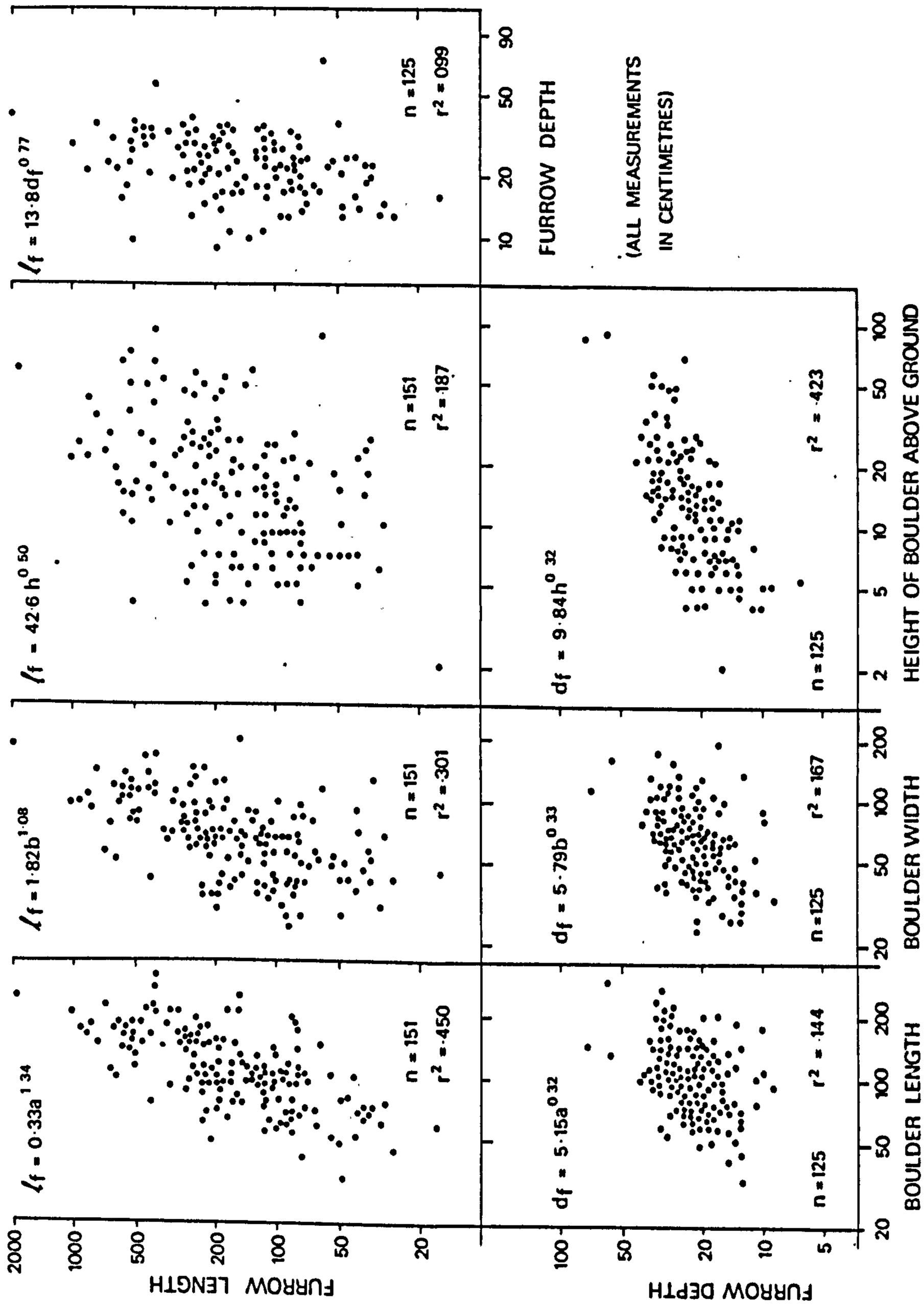


Figure 9.52: Logarithmic plots illustrating the relationships between furrow length (ℓ_f), furrow depth (d_f), boulder length (a), width (b) and height above the ground surface (h) for ploughing boulders surveyed on the Fannichs and Ben Wyvis.

Table 9.12

Ploughing boulder displacement at 800 m altitude on Scùrr Breac

a (cm)	b (cm)	h (cm)	ℓ_f (cm)	d_f (cm)	Slope (degrees)	Displacement (mm) 310876-150977	Displacement (mm) 150977-090979
301	161	91	392	59	17.0	9	11
144	114	87	61	75	20.5	11	14
183	82	12	81	32	13.5	9	9
145	77	70	522	24	20.0	NM	28
111	63	7	230	15	20.0	NM	27
146	140	6	259	13	23.0	NM	47
177	95	22	216	36	24.5	NM	69

NM - Not measured.

rates of displacement on Scùrr Breac is the strong control exerted by slope (figure 9.53). This was not evident in the data collected by Shaw or Tufnell, but is sufficiently pronounced to suggest that gradient is the dominant control of movement at this site. Boulder size, however, had no apparent influence on rates of displacement. This finding does not necessarily conflict with the morphometric evidence presented above, where the relationship between boulder size and furrow length indicates that boulder size influences not absolute rates of movement, but displacement relative to the surrounding regolith. Together, these data indicate that although slope is an important determinant of rates of movement, the amount by which a boulder outstrips surrounding mobile regolith is at least partly a function of boulder size.

9.4.5 Distribution

The lowest altitude at which ploughing boulders are found on the Fannichs and Ben Wyvis is about 700 m, but elsewhere they descend much lower. On Orval (map 8) they occur at 500 m, and Kelletat (1970a) reported ploughing boulders at a similar altitude on Ben More, Mull. King (1968) considered the lower limit of ploughing boulders to be 700 m in the Cairngorms, but Shaw (1977) and Tufnell (1972) recorded ploughing boulders as low as 450 m on Lochnagar and in the Pennines respectively. The lower limits reported for Wales vary considerably, from 610 m (Moelwyn) to 885 m (Carneddau) (Goodier and Ball, 1969; Ball and Goodier, 1970). No country-wide trend emerges from these figures.

Reports of the gradients on which ploughing boulders occur, however, show considerable concordance. Gradients measured during the survey of 151 ploughing boulders on the Fannichs and Ben Wyvis ranged from 10° to 34° (figure 9.50). This accords with Tufnell's (1972) observation that few ploughing boulders occur on slopes of less than 10° or greater than 30°, and is similar to the 9°-38° range measured by Shaw (1977). The lowest reported gradient on which ploughing boulders are found in upland Britain is 6°, measured by Galloway (1958) in the Southern Uplands.

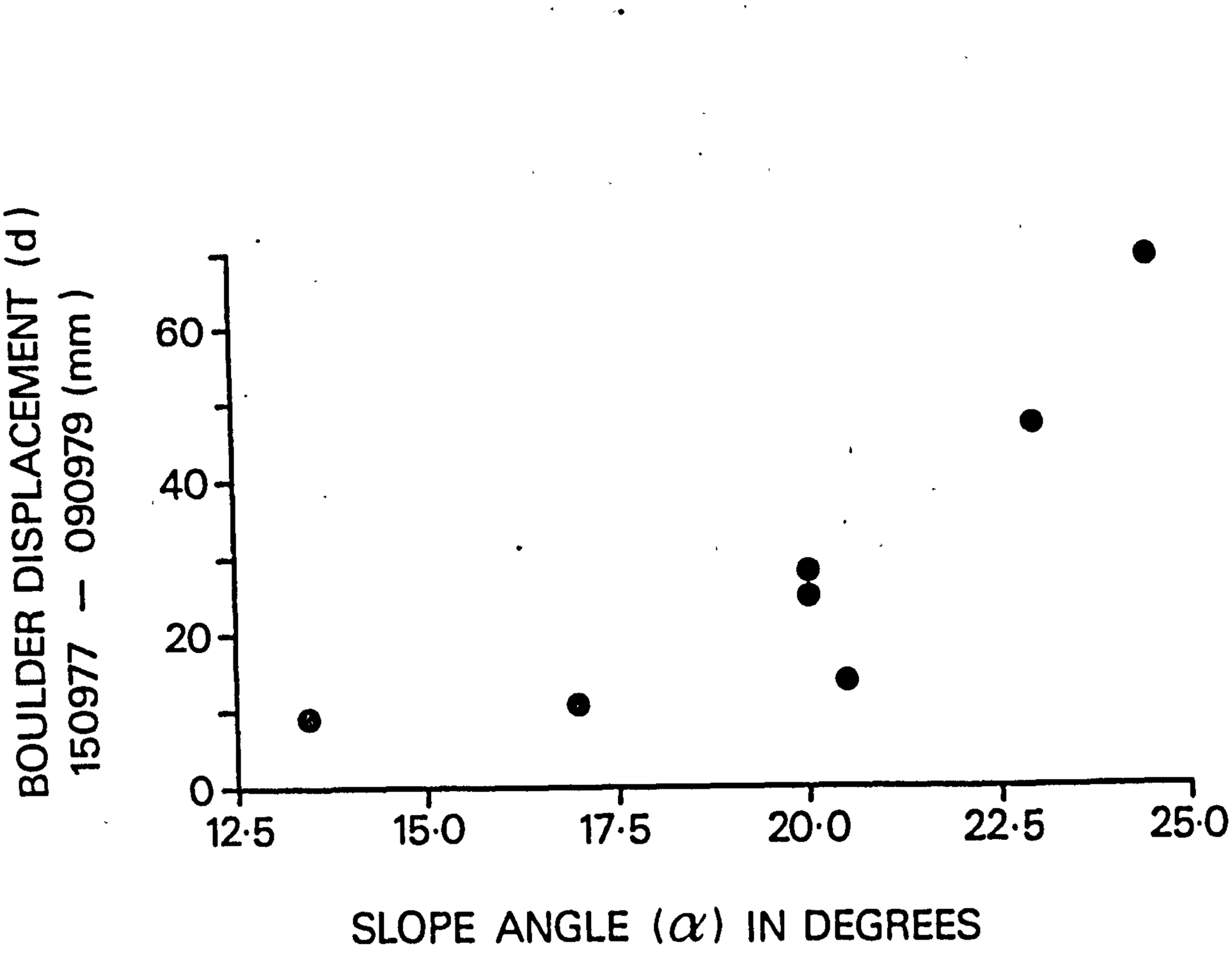


Figure 9.53: Downslope displacement of seven ploughing boulders at 800 m altitude on Scurr Breac (western Fannichs) plotted against gradient.

It therefore appears that ploughing boulders occupy a similar range of slopes to that on which sheets, terraces and lobes are found (6^{o}-36^{o}).

Two aspects of the mapped distribution of ploughing boulders are striking. The first is their absence on certain rock types, most noticeably the Torridon Sandstone of An Teallach and, less surprisingly, blockfield-forming rocks such as quartzite (An Teallach), granulite (Càrn Gorm, Ben Wyvis) and microgranite (Sron an t-Saighdeir, Rhum). The second is the close association of ploughing boulders with vegetation-covered solifluction sheets and lobes. In the Western Fannichs and on the north-west slope of Ben Wyvis ploughing boulders extend downslope only as far as the lowest solifluction sheets and elsewhere are rarely found in the absence of solifluction features. This association appears to be of general significance. It is evident in diagrams drawn by Kelletat (1970a, figures 10, 12, 14, 17 and 19) that show the distributions of "rasenloben" and "wanderblöcke" to be virtually identical on Ben Lawers, the Drumochter hills, Ben Klibreck and Ben More, Mull. Tufnell (1976) also mentions the close association of some ploughing boulders with what he terms "gelifluction terraces". The significance of these two aspects of ploughing boulder distribution is discussed in the following section.

9.4.6 Discussion: the nature of ploughing boulder activity

Tufnell (1972) suggested four mechanisms of boulder movement, namely insolation creep, frost creep, freezing of water trapped upslope of the boulder and " ... sliding of a block on the upper surfaces of frozen ground which lies directly beneath a rather wet and fluid active layer" (p. 260). He implied (p. 258) that all four mechanisms contribute to displacement. Others, such as Poser (1954) and French (1976) have suggested that solifluction is the main agency responsible for boulder movement. The latter wrote (p. 139) that

> " ... solifluction sheets ... are capable of transporting large erratic boulders. The latter, which are sometimes referred to as 'ploughing boulders', are rafted on the surface ..."

The adequacy of these proposed mechanisms is here examined in the light of the evidence presented earlier.

Insolation creep (sensu Statham, 1977) is the slow movement of clasts down inclined surfaces as a result of temperature fluctuations. Expansion and contraction of clasts on heating and cooling act with gravity to produce net downslope displacement (Moseley, 1869; Davison, 1888a, b). However, this mechanism has only been demonstrated to be effective where clasts rest on regular surfaces, and is unlikely to act where boulders are deeply embedded in the soil, and where diurnal temperature changes within the buried portion of the boulder are liable to be slight. Moreover, this mechanism fails to explain the absence of ploughing boulders on low ground (where diurnal temperature variations are often greater than on mountain slopes) and the predominance of boulder movement in the winter months (Tufnell, 1972) when the amplitude of diurnal temperature fluctuations is much reduced (section 5.2). It therefore seems unlikely that insolation creep plays a significant role in ploughing boulder movement.

Frost creep was considered a plausible mechanism of boulder movement by Washburn (1979) who defined ploughing boulders as "a special kind of frost creep and/or gelifluction deposit" (p. 223). Although it is possible that frost creep causes boulder movement in more severe periglacial environments (Schmid, 1958), it is unlikely that this process is effective in upland Britain, where deep frost penetration is infrequent and where significant frost creep is apparently confined to the upper few centimetres of regolith (figure 9.44), well above the base of most ploughing boulders. This mechanism may therefore also be discounted as an effective cause of boulder movement.

Tufnell's third suggestion (that freezing of shallow pools of water trapped in furrows upslope of ploughing boulders results in the boulder being pushed downslope by expansion of ice on and after freezing) lacks conviction. Despite his assertion that "it is not uncommon to find that water has collected to a depth of several centimetres just behind a block" (p. 260), the author has found this to be a rare occurrence. Moreover, the stresses set up by freezing of such water are more likely to be absorbed by the

vegetation and soil bordering the furrow than by displacement of a heavy boulder deeply embedded in the regolith. Finally, this mechanism fails to explain the greater displacement of boulders on steeper (and presumably better drained) slopes (figure 9.53).

Sliding of boulders on a frozen ground surface when the regolith above is "rather wet and fluid" is equally unlikely. The angle of plane friction for boulders on a rock surface normally exceeds 30°, and it is likely that frozen soil offers similar resistance to initiation of sliding. Most ploughing boulders rest on slopes well below 30° (figure 9.50), on which sliding cannot take place. Moreover, for sliding to occur the resistance offered by the soil in which the boulder is embedded must be overcome.

In view of the close association of ploughing boulders and active solifluction features, the interpretation of boulder movement as a form of solifluction seems much more plausible than any of the four explanations offered by Tufnell. This explanation also accounts for the absence of ploughing boulders on An Teallach, where large surface boulders and slopes in the range 6°-36° are abundant but the granulometry of the Torridon Sandstone regolith apparently precludes ice-lens formation and therefore solifluction. The mechanism proposed by French (1976), namely the "rafting" of boulders on the surface of solifluction sheets is, however, inadequate in that it fails to account for the furrow left behind the boulder and the mound or "bow wave" that many ploughing boulders have pushed up as they advanced. These features clearly indicate that ploughing boulders move <u>through</u> the soil at a rate exceeding that of the associated solifluction mantle. Although the distributional evidence strongly favours solifluction as the mechanism underlying ploughing boulder movement, the manner of operation of this process requires further explanation. This is attempted below.

9.4.7 A simple model of ploughing boulder movement by solifluction

Consider a boulder of length (downslope) a, width (across-slope) b and mass m embedded to a depth z on a slope of angle α.

The vertical weight of the boulder is equal to m.g, where g is acceleration due to gravity. This may be resolved into a downslope component of shear stress (τ) given by

$$\tau = m\ g \sin \alpha / (b\ z \cos \alpha) \qquad (9.10)$$

and a perpendicular component of normal stress (σ_n) given by

$$\sigma_n = m\ g \cos \alpha / (a\ b) \qquad (9.11)$$

Resistance to boulder movement may be expressed as shear strength (τ_f) using a modified form of Coulomb's failure criterion (Carson, 1971):

$$\tau_f = (\sigma_n - u) \tan \phi_\mu' + s \qquad (9.12)$$

$$= \left[(m\ g \cos \alpha / (a\ b)) - u\right] \tan \phi_\mu' + s \qquad (9.12a)$$

where u is pore water pressure, ϕ_μ' is effective angle of plane sliding friction and s is the shearing resistance of the soil downslope of the boulder. This is given by

$$s = c' + (\sigma_z - u) \tan \phi' \qquad (9.13)$$

where c' is effective soil cohesion, σ_z is the normal stress at depth z in the soil and ϕ' is effective angle of internal shearing resistance of the soil.

Boulder movement will occur when $\tau \geq \tau_f$, in other words when shear stress exceeds shear strength. As τ is effectively constant, this can only happen through a lowering of the value of τ_f. This can occur in two ways. First, an increase in pore water pressure (u) may result in a lowering of the effective normal stresses $(\sigma_n - u)$ and $(\sigma_z - u)$ to a point where $\tau_f < \tau$. However, stability analyses of solifluction slopes (Williams, 1966; Harris, 1977) have indicated that such slopes are stable even under saturated conditions when the water table is at the soil surface. Moreover, operation of this mechanism would be expected to result in shallow slab failure, and could occur at any time of year (following rainstorms) yet Tufnell convincingly demonstrated that movement is largely if not entirely confined to the winter months.

The second mechanism involves localized liquifaction and flow of soil layers that are physically separated by thawing ice lenses

(Carson and Kirkby, 1972; Harris, 1977). If this occurs under the boulder, then the shear strength of the contact between the boulder and the underlying soil would be drastically reduced as

$$(\sigma_n - u) \tan \phi'_\mu \longrightarrow 0 .$$

If all strength is lost at the contact, boulder movement will occur when

$$\tau \geqslant s .$$

As the value of c' (equation 9.13) is likely to be negligible under such circumstances, movement will occur when

$$m\ g \sin\alpha / (b\ z \cos\alpha) \geqslant (\sigma_z - u) \tan \phi' .$$

Some idea of the stresses involved may be obtained by considering a "typical" boulder of density 2700 kg m^{-3}, with a = 1.0 m, b = 0.6 m and height (h) above ground 0.2 m, embedded to a depth (z) of 0.4 m on a slope of 20°. In this case

$$\tau = m\ g \sin\alpha / (b\ z \cos\alpha)$$

$$= \frac{[a\ b\ (h + z) \cos\alpha\ \rho] g \sin\alpha}{b\ z \cos\alpha}$$

$$= \frac{(1.0 \times 0.6 \times 0.6 \times 0.940 \times 2700) \times 9.8 \times 0.342}{0.6 \times 0.4 \times 0.940}$$

$$= 13.574 \text{ kN m}^{-2}$$

$$\simeq 1.36 \text{ kgf cm}^{-2} .$$

Stresses of this order are unlikely to cause shearing of the soil downslope (Carson and Kirkby, 1972, p. 93). However, as this soil is also presumably subject to localized liquifaction and loss of shear strength, the stress exerted by the boulder will tend to accelerate its downslope movement so that the boulder moves faster than the surrounding regolith. That shearing of the soil downslope of ploughing boulders does occur under such circumstances is evidenced by the formation of frontal mounds or "bow-waves" pushed up by the shear stresses imposed by the advancing boulder.

Although not directly testable (e.g. by measuring the shear strength of soil downslope of the boulder under "solifluction" conditions) this model offers an explanation of all the characteristics of ploughing boulders described earlier.

(i) The model accounts for the movement of boulders through the regolith, leaving furrows upslope and sometimes producing a frontal mound or "bow-wave" downslope.

(ii) The model explains the close association of ploughing boulders and solifluction features, and the absence of ploughing boulders on soils that are too coarse to permit the formation of segregated ice lenses.

(iii) The relationship between furrow length and boulder size (mass) is explicable in terms of the greater shear stresses imposed by larger boulders, particularly those that are not deeply embedded. The larger the boulder, the greater the downslope shear stresses hence the faster boulder movement relative to movement of the surrounding regolith.

(iv) The model accounts for the absence of small ploughing boulders below a critical mass. Small boulders exert insufficient shear stress to cause accelerated movement.

(v) The preferred downslope orientation of ploughing boulders probably occurs because this orientation is the most "efficient", allowing the application of shear stress against the minimum area of soil. Moving boulders will therefore tend to rotate and thereby adopt the orientation of least resistance.

(vi) The nonlinear relationship between boulder displacement and slope is explicable in terms of increasing shear stress ($\sin \alpha$) but decreasing normal stresses ($\cos \alpha$) with increasing gradient.

It is therefore concluded that the movement of ploughing boulders in upland Britain is largely if not entirely attributable to solifluction (sensu stricto) in the form of localized liquifaction and flow of soil layers physically separated by thawing ice lenses underneath the boulders.

9.5 Landforms produced by rapid mass-movement

9.5.1 Introduction

Within the areas depicted by maps 2-8 slope failure and consequent rapid mass-movement of surficial debris has produced two distinct morphological responses. The first, described in the map key as "slump and scar" and defined (chapter 7) as a "concave scar and associated tongue of detritus produced by localized slope failure" is not common on high ground and is represented by only five examples in the mapped areas. The second, comprising debris chutes (the routeways of debris flows) and associated debris cones, is widespread. In this section the characteristics, distribution and activity of these features are described; they are then discussed in the light of current theory regarding the nature of failure and movement.

Neither slump features nor debris flows are restricted to periglacial areas, although the latter in particular appear to be common in cold environments and have been described in Spitzbergen (Rapp, 1960b; Jahn, 1976), Alaska (Price, 1969), Yukon Territory (Sharp, 1942b)and northern Scandinavia (Rapp, 1960a, 1962). Possibly the earliest account of debris flow activity in upland Britain is that of Barrow and Cunningham Craig (1912, p. 119), who described rapid mass-movement during heavy thunderstorms on Glas Maol (S.E. Grampians). The widespread slope failures that accompanied an exceptional storm in the Ben Nevis area were documented by Common (1954) and the appearance of debris flows on many Cairngorm slopes following intense rain in August 1956 was strikingly described by Baird and Lewis (1957) under the unfortunate term "summer solifluction". More detailed descriptions of debris flows, together with data on their frequency and relationship to rainfall intensity have been provided by research on features associated with the Antrim basalts (Prior et al., 1970, 1971; Prior and Douglas, 1971). Statham (1976b) described debris flow features on the Black Mountain, Carmarthenshire, and considered their present role in reshaping vegetated talus slopes. Shallow

slides producing slump features similar to those in the study areas have been described in a number of studies (a recent example is Bevan et al., 1978), though none of these relates to features at high altitudes.

9.5.2 Characteristics of forms produced by rapid mass-movement

9.5.2.1 Slump features

Three of the slump features within the mapped areas are located at the head of Glas Tholl, An Teallach (076848, map 2) on the lower reaches of a partly-vegetated debris slope at an altitude of around 650 m. The remaining example investigated is located on a vegetated debris slope on Ben Wyvis (447673, map 5) at a similar altitude.

The An Teallach examples have scars that are respectively 100 m, 25 m and 40 m long (downslope) and 30 m, 25 m and 20 m wide. The scarp at the upslope edge of each scar is roughly one metre high, representing the thickness of weathered regolith above the bedrock floor that forms the plane of failure, inclined at 32^{o}-34^{o}. Downslope of the scars are transit slopes over which the debris from the scars has travelled, eroding or burying vegetation but disturbing only the surface of the regolith. At the basal break of slope the products of failure have accumulated as disordered mounds of boulders and sand. As these mounds are largely vegetated and as oblique terraces have developed across the transit slope of the central feature it is evident that none of the failures are recent.

The Ben Wyvis slump feature is similar in dimensions to the smallest An Teallach example, but lacks a transit slope between scar and deposit. Like those on An Teallach, the landslide deposits are vegetated although the scar (which appears to be floored by frost-weathered bedrock) is not. The inclination of the slide plane is 32^{o} and the deposits rest on a slightly gentler slope (27^{o}-28^{o}). All these features fall into the category of shallow slab slides as defined by Skempton and Hutchinson (1969).

9.5.2.2 Debris flow features

Debris chutes occur in most of the mapped areas (maps 2, 3, 5, 7 and 8) but they are most abundant on An Teallach. Here they occupy both active and vegetated talus (e.g. 072844) but are densest on steep unvegetated debris slopes with shallow regolith cover. On the northern slopes of Glas Mheall Mór, for example, the author counted 45 chutes over a kilometre of slope, although many of these have been partly obliterated by frost creep. On other mountains, chutes are generally found only in association with talus, particularly vegetated talus (e.g. 166715, map 3; 313938, map 7; 337999, map 8). They are nowhere found on solifluction-prone slopes, and only rarely on blockslopes (e.g. 078838, map 2) which within the study area are rarely steep enough for flow initiation. On Beinn Eighe (Torridon), however, very large chutes sweep down from steep quartzite blockslopes on to the underlying Torridon Sandstone.

The source areas for all debris chutes examined have gradients of 30° or more. Altitude exerts little control on chute distribution: they extend downslope to almost 100 m in Glen Dibidil (map 7) and chute sources range in altitude from 350 m (337999, map 8) to over 900 m (073853, map 2). Aspect also has no apparent influence on chute distribution.

All the mapped debris flow features are of the type classified by Brunsden (1979) as hillslope flows. Within this category of debris flow it is possible to recognise two subtypes, referred to hereafter as simple and multiple. Simple chutes begin as narrow, shallow gulleys on the steep upper reaches of debris or talus slopes. As the slope angle lessens, erosion is succeeded by deposition and the chute becomes delimited by parallel levées of debris (Sharp, 1942b). Simple chutes terminate in a solitary debris lobe, indicating formation in the course of a single event. Good examples furrow the northern slopes of Glas Mheall Mór (figure 9.54).

Multiple debris flow features are fed by much larger gulleys cut in either slope deposits or rock. The former type



Figure 9.54: The tracks of "simple" debris flows on the northern flank of Glas Mheall Mór, An Teallach. The flows cut through an area of deflation terraces and terracettes and terminate in the rock gorge occupied by the Allt Coire a'Mhuilinn.

of source is common on the debris slope at the head of Coire Mór, An Teallach (066845; figure 9.55), the latter 500 m farther east at the head of Glas Tholl. The lower part of multiple chutes is also delimited by a pair of parallel levées. Near the foot of the slope, however, levées produced by the most recent debris flow event cross others, often moss- or lichen-covered, that relate to earlier events and demonstrate repeated use of the same routeway. At the terminus of large multiple chutes repeated deposition has led to the accumulation of well-marked debris cones similar to those described by Johnson and Rahn (1970), Campbell (1974), Jahn (1976), Statham (1976b) and others. Repeated use of such chutes is attributable to the wider gulley "catchments" at their heads.

The transition from erosion to deposition on debris chutes is difficult to define, as levées sometimes extend upslope well into the gulleyed zone, especially on simple chutes and those developed on coarse detritus (particularly talus). On surveyed chutes (figure 9.56) the transition usually occurs where the slope drops below 24°-28°, although for some multiple chutes in Coire Mór the change occurs on gradients as low as 20°, still steeper than the 16° transition measured by Statham (1976b) on multiple chutes in Wales.

For nine simple chutes surveyed on Glas Mheall Mór, gradient at the gulley head (the presumed site of initial failure) proved remarkably uniform, ranging from 32.0° to 34.5° (table 9.13). Regolith thickness at the site of failure ranged from 0.4 to 0.7 m. Many chutes have reaches steeper than the failure plane where they cross rock outcrops (figure 9.55). Some originate at or near the base of cliffs, which suggests that they may have been triggered by rockfall. Others (e.g. at 080859, map 2) run downslope to the crest of cliffs, indicating that such flows have supplied debris to the talus slopes below.

The angle at which flows have come to rest is much less uniform than the angle of initiation, ranging from 11.0° to 23.5° for a sample of seven debris chutes. Even the mean gradients of the lowest five slope segments are very variable (15.5°-24.0°). In two chutes the most recent debris flows have not even reached the foot of the slope but form boulder dams similar to that described



Figure 9.55: "Multiple" debris flows at the head of Coire Mor, An Teallach. Many of these features terminate downslope on debris cones that are evidence of repeated use of the same routeways.

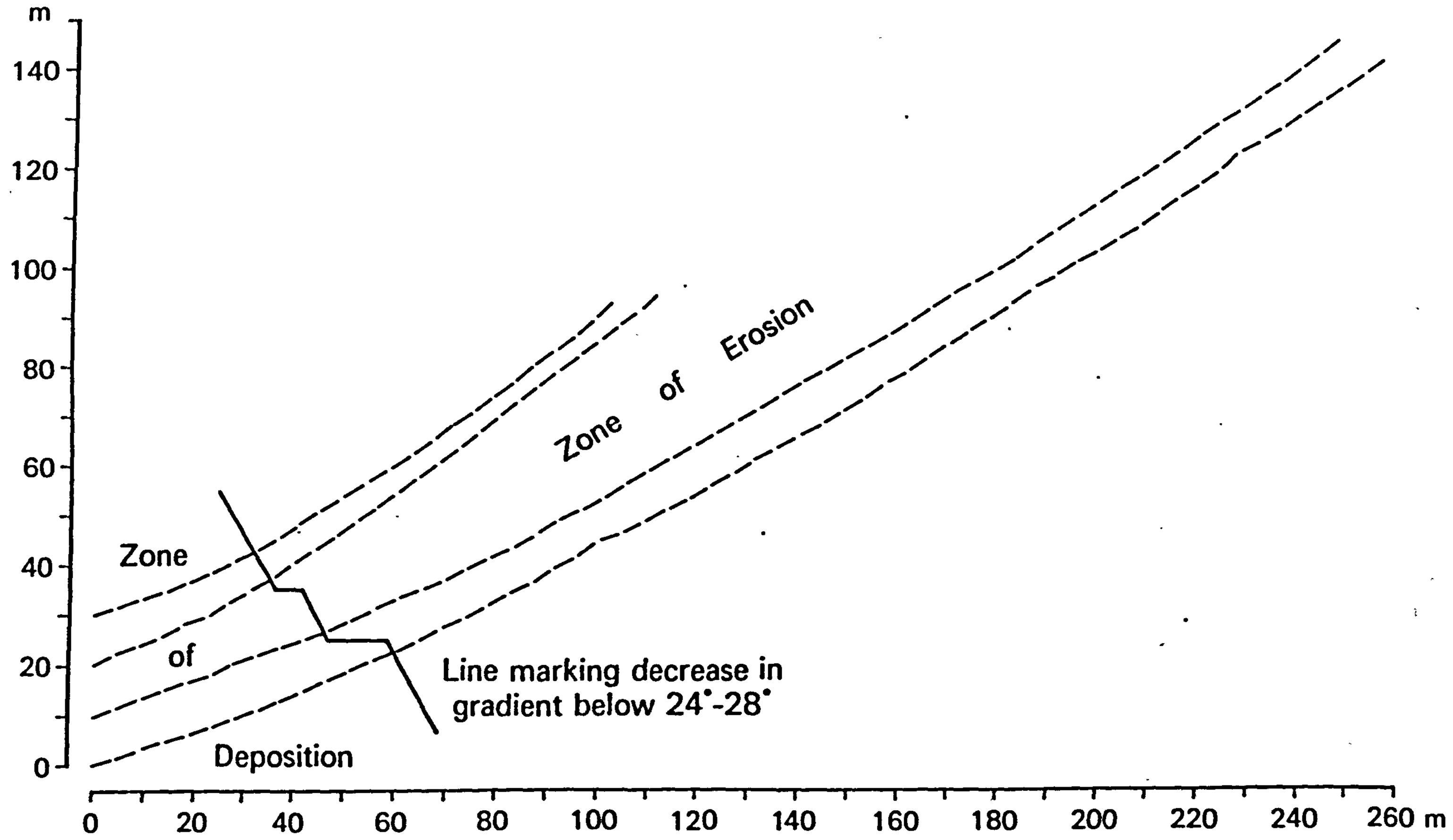


Figure 9.56: Surveyed profiles of four debris chutes on Glas Mheall Mór, An Teallach. Each dash represents a 5 m surveyed segment.

Table 9.13

Summary of survey data for nine simple debris flows, Glas Mheall Mór

Flow no.	Failure angle	Maximum gradient	Minimum gradient	Angle of rest of debris lobe	Mean of 5 lowest readings
1	32.0	32.0	N.M.	N.M.	N.M.
2	34.0	40.0	N.M.	N.M.	N.M.
3	34.0	34.0	10.0	13.0	16.0
4	33.0	38.0	11.0	11.0	15.5
5	33.0	36.0	17.0	17.0	19.5
6	33.5	44.0	13.5	16.5	19.0
7	34.5	39.0	15.0	15.0	19.5
8	34.0	38.5	22.5	23.5	24.0
9	33.0	46.0	12.5	16.5	17.5

All gradients are given in degrees. These were measured at 5 m intervals using an abney level. Flows 1 and 2 terminate in a rock gorge, so measurements cannot be given for the last three figures.

by Statham (1976b) that block the chutes at gradients of 25° and 27°. The range of slopes on which flows come to rest is therefore great, and it appears that some may become immobilized on gradients as little as 5° less than that of the failure plane.

The zone of deposition exhibits several interesting characteristics. Paired levées closely follow local changes in the direction of maximum slope, curving round the margins of talus cones, debris cones and even moraines (078860, map 2). Where the lower parts of chutes have crossed turf the levées have been deposited directly on to the turf cover and vegetation between the levées has remained undisturbed (Rapp, 1962), indicating very low shear stresses at the base of the mobile flow in the depositional zone. From this it may be inferred that the mode of transport in this zone was "true" flow, with the moving mass behaving as a viscous fluid. Yet the competence of flow in some cases must have been extraordinarily high, as boulders with lengths exceeding one metre (the largest measured 180 cm x 120 cm x 85 cm) are found in the levées of the largest chute in Glas Tholl, a feature over 450 m long averaging 6-7 m in width with levées up to 2 m high. Most chutes are much smaller features, 1-3 m in width with levees 0.3-0.5 m high.

Reports of recent debris flow activity in upland Britain (e.g. Baird and Lewis, 1957; Statham, 1976b) and even on low ground (Prior et al., 1970; Prior and Douglas, 1971) suggest that rapid movement in this form occurs not infrequently. This is borne out by the author's own observations in the period 1976-9. On Glas Mheall Mór a single chute was formed between July 17 and July 30, 1976, and four new chutes appeared over the winter of 1978-9, possibly related to the rainstorm of 4th October 1978. On the east side of Drumochter Pass two flows were active during rainstorms on July 22 or 23, 1977, and a further flow on the west side of the Pass was active in late September or early October 1978. When Coire an Lochain in the Cairngorms was visited on August 4, 1977, a flow with its source on rock slabs at the foot of the headwall had recently descended on to low angle slopes on the corrie floor. The abundance of loose sand in the area of the terminus suggested that the flow had occurred only shortly

beforehand. Finally, the ubiquity of debris flow as a form of rapid mass-movement in Scotland was demonstrated by the occurrence of a small flow from a vegetation-free gulley (the Gutted Haddie) on Arthur's Seat (251 m) near the centre of Edinburgh during intense rain in October 1977. All the flows documented above resulted in the deposition of considerable quantities of sand in the area of the lobe terminus, most of which was rapidly carried away by wash and deflation.

Only one example of a shallow slide was witnessed during the same period. This took place at an altitude of about 350 m on the lower slopes of Beinn Odhar near Tyndrum during heavy rain in mid-September 1978, when a large slab of till over 2 m thick failed along a slip plane of estimated gradient 30^{o}-35^{o}. Failure may have been triggered by vibrations from trains approaching the summit of the Tyndrum-Bridge of Orchy col causing liquifaction, rupture and movement in the form of a flowslide similar to those described by Rapp (1960b) and Prior et al., (1970).

9.5.3 Discussion

The measurements reported above of gradient at the point of initiation of debris flows indicate that such flows generally begin on inclines of 32^{o} or greater. Debris- or talus-mantled slopes with gradients in excess of 32^{o} are found in all the mapped areas, yet on many such slopes (e.g. the north-west flank of Ben Wyvis, the north-west and south-west slopes of Scùrr na Clach Geala, in the eastern Fannichs) debris chutes are absent, and on most others they are rare. On only two debris slopes on An Teallach (those shown on figures 9.54 and 9.55) is their distribution sufficiently dense that they form the dominant features produced by mass-movement. Yet on other debris slopes on the same mountain (e.g. south of Glas Mheall Mór and Scùrr Creag an Eich) they are also rare or absent.

There are several possible explanations for the high density of chutes on the west flank of Bidean a' Ghlas Thuill and the north-west slope of Glas Mheall Mór. First, the aspect of these slopes means that they are likely to experience the brunt

of cyclonic rainstorms. Secondly, the regolith on the upper parts of these slopes is thin (0.4-0.7 m), and, thirdly, sandy and therefore cohesionless. Finally, the upper parts of both slopes are largely unvegetated.

Convincing evidence has been supplied by Johnson and Rahn (1970) to indicate that the initial slope failure that generates debris flows takes the form of shallow planar slides. The factor of safety (F) for an infinite planar slide in cohesionless soil on a slope with angle α may be expressed as

$$F = \frac{(\cos^2\alpha - r_u) \tan \phi'}{\cos \alpha \sin \alpha} \qquad (9.14)$$

where ϕ' is effective angle of shearing resistance and r_u is pore water pressure ratio (Statham, 1976b, after Skempton and DeLory, 1957). The latter term is given by

$$r_u = \frac{u}{\gamma z} \qquad (9.15)$$

where u is pore water pressure, γ is the unit weight of soil and z is the thickness of the slide. Failure occurs when F drops below unity. Given that, for any site, α, ϕ, γ, and z are effectively constant, failure will occur when u (pore water pressure) is high relative to the product of γ (unit soil weight) and z (effectively depth of regolith). Variations in γ are unlikely to be great (probably of the order of 1.7-2.1 tonnes m^{-3}) and hence the critical factors determining the susceptibility of steep cohesionless debris slopes to shallow sliding are pore water pressure and soil depth. The importance of the former is manifest in the association of debris flow activity and intense rainstorms (when rate of water input to the soil exceeds rate of throughflow drainage, producing high pore pressures). However, it is the second factor, soil depth, that probably explains the susceptibility of the two west- to north-west-facing debris slopes on An Teallach to failure and consequent debris flow, as the regolith on both is generally shallower (0.4-0.7 m) than on steep south-facing debris slopes on the same mountain (c. 1.0 m or thicker), a difference attributed in section 8.4.5 to greater Lateglacial frost weathering on south-facing slopes.

The regolith mantle on the Fannichs and Ben Wyvis also tends to be over one metre thick, and in such areas stability is increased by cohesion resulting from the presence of a higher clay-silt fraction in the soil and possibly by a binding cover of vegetation. The former factor explains the absence of debris flows on solifluction-prone slopes, in that although a relatively high clay-silt fraction favours ice segregation and therefore solifluction, it increases cohesion and therefore soil strength.

The metamorphosis from slide to debris flow is incompletely understood. It occurs when

> " ... initial movement (sliding failure) of slabs of soil and wedges of gulley fill causes remoulding of the saturated moving mass into viscous debris-laden mud ..." (Campbell, 1974, p. 339).

Such "remoulding" allows debris to travel in the form of a viscous fluid over much gentler slopes than those on which failure occurs. Johnson and Rahn (1970) suggested that the change from sliding to flow reflects three processes. The first is the reduction in strength of the mobile mass from peak to residual values, but for shallow loosely-compacted frost-weathered regolith such a change is likely to be small. Secondly, the addition of further water to the mobile mass may reduce grain-to-grain contacts, resulting in progressive fluidization. Finally, they note that fluidization may be hastened by the "piling up" of mobile sediment. Support for the final mechanism is provided by the work of Hutchinson and co-workers (Hutchinson and Bhandari, 1971; Hutchinson et al., 1974) on the role of undrained loading in mass-movement: if a water-soaked mass is compressed under further sediment so that pore water cannot escape it tends to become buoyant and flow. Support for the operation of this mechanism is provided by observations of debris flows in motion (Sharp, 1942b; Campbell, 1974) which describe flows advancing by a series of surges as boulder dams trap sediment then are breached as more sediment accumulates on top. Acceptance of this explanation enables interpretation of the observed zones of erosion and deposition as follows: on slopes over 24^{o}-28^{o} (the transition angle for surveyed chutes on Glas Mheall Mór) undrained loading under the advancing flow results in failure and flow of the underlying soil;

on gentler slopes, however, the soil is too stable for failure, so that flow takes place over the soil surface. As boulder dams are breached in the depositional zone boulders are pushed to the margins of the flow, forming levées. Under this model it is undrained loading that results in erosion on steep slopes, rather than tractive forces operating at the base of the flow. The variability in the gradients at which flows come to rest probably reflects sediment supply and therefore the size of the flow. In the depositional zone flow will only continue as long as there is sufficient sediment to allow undrained loading; as sediment is lost to the levées flow will slow and eventually cease.

Slump features also originate through shallow slide failures, but these tend to be located near the foot of steep slopes rather than near the top and to take place in deeper regolith. This probably accounts for their infrequency, in that only during exceptional rainstorms will pore-water pressures rise sufficiently to initiate failure. All the observed examples travelled relatively short distances, suggesting that liquifaction of the mobile mass was incomplete. This suggestion is supported by the survival of large intact "blocks" of soil amongst "quick" mud during the recent slide on Beinn Odhar.

It remains to be explained why debris flows follow such narrow tracks. In the case of multiple chutes fed by gulleys the answer lies in the "funnelling" effect of the source gulleys. In the case of the simple chutes failure is apparently very localized, which suggests that the upper parts of such chutes coincide with percolines down which moisture transport is concentrated.

Finally, it is noteworthy that the belief that periglacial slopes are particularly susceptible to debris flow activity on account of the occurrence of frozen ground at shallow depth does not hold for upland Britain, where many flows occur in summer. That most chutes are found on high rather than low ground is attributable to more intense rainfall, steeper slopes, lack of vegetation and above all the shallowness of the often cohesionless mantle of frost-weathered debris that cloaks high mountain slopes.

9.6 Conclusion

In the introduction to this chapter the question was posed: "why have similar slopes responded in different ways to periglacial conditions?" As the subsequent discussion developed, one factor emerged as the primary determinant of the nature of mass-movement, namely the character of the debris mantle, itself strongly related to the nature of periglacial weathering (chapter 8) and thus, ultimately, to lithology (chapter 6). To explain the distribution of different types of mass-movement features it is useful to distinguish three broad classes of regolith : blockslope deposits, debris slope deposits containing an appreciable clay-silt fraction and debris slope deposits containing a negligible clay-silt fraction.

The mobilization of blockslope deposits during the Lateglacial cold periods produced massive boulder sheets and associated lobes. This movement probably occurred under permafrost conditions, with the block deposit moving en masse as a result of annual heaving and resettling of the active layer and ceased with the onset of warmer (Flandrian) conditions. Since then such deposits have remained immobile save possibly for localized failure and consequent debris flow.

During the Lateglacial cold periods debris slopes with an appreciable clay-silt component responded in a slightly different manner, producing shallower boulder sheets and lobes through a combination of frost creep (again associated with an annual freeze-thaw cycle in the former active layer) and solifluction (or, more correctly, gelifluction) of underlying fines during thaw of segregated ice lenses. The latter process has continued to the present day, producing unsorted solifluction sheets and lobes and mobilizing ploughing boulders that move more rapidly than the surrounding regolith. It is possible that the shallow boulder lobes formed in Lateglacial times on such slopes are reactivated following exceptionally deep freezing of the ground. Where vegetation cover is incomplete surficial frost creep is the dominant form of mass-movement, producing flights of turf-banked terraces. Debris chutes are rare on such slopes as the clay-silt fraction that renders the soil susceptible to ice lens formation and

therefore solifluction also increases shear strength through cohesion and thus minimises shallow slide failures.

Frost creep of the essentially cohesionless regolith of debris slopes containing a negligible clay-silt component produced a variety of Lateglacial mass-movement features, from massive boulder sheets and lobes where the regolith was thick through shallower debris sheets to debris terraces. On lee slopes postglacial vegetation colonization of the risers of the last-mentioned and surficial frost creep on the unvegetated treads produced "normal" turf-banked terraces. Oblique and deflation terraces developed ab initio during the Flandrian in response to a combination of wind action and surficial frost creep. Solifluction features, including ploughing boulders, are absent as the regolith is too coarse to favour ice lens formation. Conversely, debris chutes are abundant as the cohesionless soil has relatively low shear strength and is thus susceptible to failure when pore pressures rise during rainstorms.

Thus regolith alone plays a fundamental role in determining the nature of slope response to periglacial mass-movement. Its importance in regulating the distribution of other types of landform is examined in the following three chapters.