PARAGLACIAL MODIFICATION OF DRIFT-MANTLED HILLSLOPES

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PARAGLACIAL MODIFICATION OF DRIFT-MANTLED HILLSLOPES

by

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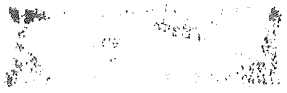

For from Him

and through Him

and to Him

are all things.

Abstract

The aim of the research reported in this thesis is to establish the characteristics of paraglacial modification of drift-mantled hillslopes in glaciated upland valleys in Norway and Scotland. Debris flow represents the principal agent of paraglacial sediment reworking, though snow avalanches and slopewash are locally important. Paraglacial hillslope modification is most widespread in areas of thick drift where initial slopes exceed *c.* 30°, void ratio exceeds *c.* 0.35, and water input is focused both spatially and temporally. Paraglacially-reworked sediments preserve most of the characteristics of the parent tills, but differ in terms of preferred clast orientation and structural and lithofacies characteristics. Stratigraphic relations between tills and reworked sediments imply cyclic alternation of glacial and paraglacial sediment transport. Paraglacial slope adjustment follows a sequence involving (1) rapid gully incision; (2) widening of gullies, and accumulation of debris cones at the slope foot; (3) reduction and destruction of inter-gully divides, and formation of a slope-foot apron of coalescing cones; (4) extensive exposure of bedrock at the crest of the slope, resulting in sediment exhaustion and progressive stabilisation. Slope profiles tend to converge on a maximum gradient of *c.* 28° and a concavity index of *c.* 0.22. At the most active sites, 2-4 m of gully lowering has occurred within decades of deglaciation, implying minimum erosion rates averaging *c.* 90 mm yr^{-1}. In Scotland delayed or renewed reworking of drift-mantled slopes has occurred several millennia after deglaciation. Radiocarbon dating of buried palaeosols indicates intermittent drift reworking by debris flows throughout the past 6.5 ka, with some evidence for accelerated activity at *c.* 2.7-1.7 cal ka BP and after *c.* 0.7 cal ka BP. Three-dimensional conceptual models are developed to describe the sequence of both immediate and delayed or renewed paraglacial hillslope modification, and the landforms and sediment associations characteristic (and diagnostic) of paraglacial landsystems in passive continental margins.

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Chapter 1

Introduction

1.1 Definition and context.

The term *paraglacial* was first coined by Ryder (1971a), and was defined by Church and Ryder (1972, p. 3059) as referring to "nonglacial processes that are directly conditioned by glaciation", and encompassing both proglacial processes and those occurring around and within the margins of a former glacier. Specifically, the term was used by Ryder (1971a, 1971b) and Church and Ryder (1972) to describe the way in which Late Pleistocene deglaciation introduced an abrupt and radical change in terrestrial erosional and depositional fluvial environments, by locally exposing vast quantities of unconsolidated glacigenic detritus which were vulnerable to erosion, reworking and redeposition by fluvial and hillslope processes. Subsequently, use of the term has broadened from that implied in the original definition. In particular, the notion of paraglacial landscape change has now been extended to incorporate nonglacial processes, landforms, landsystems and deposits conditioned by glaciation within hillslope and coastal, as well as fluvial, settings.

The paraglacial concept is particularly useful in that it describes an explicit form of geomorphological response, and attempts to explain landscape evolution and modification over a wide range of temporal and spatial scales (Harvey, 1990). The relevance of the concept is particularly evident in the context of recent retreat of mountain glaciers, which has resulted in marked changes in many mountain geomorphic systems. Floods, debris flows and landslides have modified the hydrology, sedimentology and morphology of numerous recently-deglaciated alpine environments. Further, in the last 150 years, events related to the retreat of

glacier ice are estimated to have claimed more than 30,000 lives, and to have caused damage to economic infrastructure costing in excess of one billion dollars (Evans and Clague, 1994). Whilst this thesis is specifically devoted to geomorphological and sedimentological aspects of paraglacial hillslope modification, the recognition that deglaciation can radically disrupt natural alpine processes and the lives and livelihoods of those living in alpine environments has important implications for hazard assessment and future development in such areas. In particular, the research reported in this thesis deals specifically with the paraglacial modification of drift-mantled hillslopes. The term 'drift' is used here to refer to sediment eroded, transported and deposited by glacier ice or meltwater. *In situ* drift is regarded as drift which has remained intact following emplacement, though it may have experienced short-distance remobilisation during initial deposition as a result of falling, rolling and sliding (e.g. Humlum, 1978; Eyles, 1979; Evans, 1989a, 1989b), or melt-out (Boulton, 1971; Lawson, 1979a, 1979b). In contrast, paraglacial reworking of drift constitutes a subsequent, discrete period of sediment remobilisation.

1.2 Research aims.

Although the concept of paraglacial hillslope modification has been increasingly employed within geomorphology in recent years (e.g. Owen, 1991; Ballantyne and Benn, 1994; Ballantyne, 1995a; Fitzsimons, 1996; Harrison and Winchester, 1997), several issues relating to the characteristics, causes and consequences of paraglacial modification and resedimentation of glacigenic drift remain largely unresolved. Amongst the most important are: (i) the spatial distribution of, and constraints on paraglacial activity; (ii) the timing and duration of paraglacial adjustment; and (iii) the sedimentological and morphological consequences of paraglacial activity.

In seeking to address these unresolved issues, the aims of the research reported in this thesis are essentially threefold. The first is to identify and establish the nature and extent of paraglacial modification of deglaciated drift slopes, and its sedimentological implications. Although some workers (e.g. Owen, 1991, 1994; Ballantyne and Benn, 1994) have suggested some general criteria for distinguishing paraglacially-reworked drift cover from unmodified glacigenic deposits, further analyses are required to establish the detailed sedimentological effects of paraglacial reworking of drift. Such findings may permit more accurate assessment of the spatial extent of paraglacially-reworked drift deposits, particularly in areas where the morphological effects of sediment reworking are equivocal or indistinct. Secondly, the factors that control the initiation and termination of paraglacial hillslope modification are presently poorly understood. A further aim of this research is therefore to identify the factors that condition or trigger paraglacial reworking of drift-mantled hillslopes. This is important if the causes of paraglacial slope activity are to be understood, and the intensity or lack of activity at particular sites explained. The final aim of the research reported here is to model the temporal pattern of paraglacial modification of drift-mantled hillslopes. Some authors (e.g. Ballantyne and Benn, 1996) have suggested that paraglacial modification of the form and behaviour of drift slopes may follow a common developmental sequence. The aim of this part of the research is to establish the developmental characteristics of paraglacial hillslope adjustment, and to determine whether these follow a consistent pattern.

1.3 Thesis structure.

This thesis is divided into three broad sections. The first comprises chapters 2 and 3 and outlines the context of the research. Chapter 2 summarises current understanding of paraglacial slope adjustment, highlighting areas of contention and identifying key areas requiring further research. The relevant

characteristics of the field sites investigated in the course of this project are described in chapter 3. The second main section consists of chapters 4 and 5, where attention is focused on the nature, extent, constraints and timing of paraglacial resedimentation of glacigenic drift. In chapter 4 these themes are considered within the context of both ancient and recent or current paraglacial activity, whilst in chapter 5 the concept of renewed or delayed paraglacial reworking of drift slopes is introduced. The third section of the thesis comprises chapters 6 to 8, which relate to the consequences of paraglacial reworking of drift slopes. Chapter 6 outlines the sedimentological characteristics of recent and ancient paraglacial deposits, chapter 7 considers modification of slope form, and a model of paraglacial slope adjustment presented in chapter 8. The final chapter (9) draws together the principal findings of the thesis and highlights the most promising direction for future research.

Chapter 2

Paraglacial slope adjustment: a review

2.1 Introduction.

Church and Ryder (1972) defined *paraglacial* as a term describing the nonglacial processes operating in response to the exposure of vast quantities of glacigenic drift associated with retreat of glacier ice. They intended their definition of the term to be quite general, recognising that its application was not restricted to fluvial resedimentation of drift, nor the closing phases of any particular glaciation (Church and Ryder, 1989). Indeed, the scope of the term has subsequently broadened to include hillslope and coastal, as well as fluvial, systems, and its use has covered a range of timescales.

Following widespread deglaciation, hillslope systems may be characterised by a state of imbalance, as glacially-steepened slopes mantled with glacigenic deposits become exposed to processes of erosion, transport and redeposition, particularly by debris flows. Moreover, rock slopes may progressively weaken and eventually fail in response to glacier thinning and consequent debuttressing, or as a result of seismicity associated with glacio-isostatic crustal rebound. The legacy of deglaciation may be represented in the fluvial system by a disrupted regional drainage pattern characterised by large quantities of sediment: "The occurrence of many alluvial cones and fans, in particular, indicates the lack of complete integration of the fluvial sediment transporting system." (Church and Ryder, 1972, p. 3059).

Many mid- to high-latitude coasts developed in glacigenic sediments have been described as being predominantly paraglacial, in that they have developed on

or adjacent to formerly ice-covered terrain, where glacial landforms or glacigenic sediments have had a strong influence on the nature and evolution of the coast (e.g. Carter *et al.*, 1992; Jensen and Stecher, 1992; Shaw and Forbes, 1992; Forbes *et al.*, 1995; Forbes and Syvitski, 1994; van Heteren *et al.*, 1998). The term 'paraglacial' has also been applied to lacustrine and marine systems that receive glacigenic sediment but are not in direct contact with glacier ice or icebergs (e.g. Gilbert, 1983; Powell and Molnia, 1989; Powell and Domack, 1995; Syvitski and Lee, 1997). In addition, the concept has recently been applied to the operation of some periglacial processes (e.g. Matthews, 1992a; Matthews *et al.*, 1998). These uses of the paraglacial concept reveal its growing scope over the last two decades.

Furthermore, the term 'paraglacial' has been used not only as an adjective to describe nonglacial processes and geomorphic activity conditioned by glaciation (or deglaciation), but also to describe: (1) the manner in which paraglacial processes impinge on the environment (for example, 'paraglacial reworking' or 'paraglacial resedimentation' of hillslope drift, or the 'paraglacial sediment budget' in the fluvial catchment); (2) the landsystems and landforms created by such processes (for example 'paraglacial coasts', 'paraglacial fans' or 'paraglacial terraces'); and (3) the sediments deposited or modified by paraglacial processes following both recent (post 'Little Ice Age') and ancient (Late Pleistocene) glacier retreat. The word has also been used as an adjective of time, with the 'paraglacial period' defining the time during which paraglacial processes occur (Church and Ryder, 1972). Although not defined in the literature, the term 'paraglaciation' is proposed in this thesis to refer to the cumulative and collective effects of paraglacial processes on the postglacial landscape, much as 'glaciation' describes the direct effects of erosion and deposition by glacier ice.

Whilst some have employed the term 'paraglacial' in a loose manner, (e.g. Bates and Jackson, 1987; Harrison, 1991), others have used it in a narrow sense or

rejected its use altogether. Eyles and Kocsis (1988, 1989), for example, suggested the alternative term 'glacially-influenced' to refer to alluvial fans constructed by reworking of unconsolidated glacigenic debris, regardless of age (*cf.* Eyles *et al.*, 1985, 1998). Whilst Ryder (1971a) and Church and Ryder (1972) related the term to the time period of relatively rapid sedimentation during or immediately after deglaciation when an environment is in the process of transition from predominantly glacial to dominantly fluvial conditions, they also acknowledged that paraglacial sediment movement will continue as long as drift remains easily accessible to fluvial activity, which may be long after the initial 'paraglacial period'. Thus the original definition left the term somewhat open to interpretation. If 'paraglacial' processes are considered to be restricted to a 'paraglacial period' of rapid reworking and redeposition of sediment immediately after deglaciation, when many steep drift slopes are unvegetated and intrinsically unstable, then further drift remobilisation centuries or millennia after deglaciation may not be considered a strictly 'paraglacial' response. An alternative view is offered by Ballantyne and Benn (1996), whose recognition of delayed or renewed paraglacial slope reworking implies that they consider the critical aspect of 'direct conditioning' to be the continued supply of sediment of glacigenic origin, irrespective of the environment under which such sediment is reworked or the length of the delay between initial emplacement and subsequent reworking of such sediment. A similar viewpoint is implied in the work of Church and Slaymaker (1989) regarding fluvial reworking of glacial drift, and by the use of the term 'paraglacial' to refer to littoral environments where glacial landforms or sediments have strongly influenced coastal evolution (e.g. Carter *et al.,* 1992; Shaw and Forbes, 1992; Forbes *et al.,* 1995). The difference in the scope of 'paraglacial' implied by these two conflicting viewpoints centres around the phrase 'directly conditioned by glaciation'. The former view implies that 'glacial conditioning' reflects all aspects of the post-deglaciation environment, such as abundant, potentially unstable drift, abundant meltwater and lack of vegetation

cover. The latter view implies that 'paraglacial' *sensu lato* refers to the long-term inheritance of glacial landforms and sediments in the landscape, and their influence on subsequent geomorphic activity, irrespective of the length of time elapsed since deglaciation. For the purposes of the research reported in this thesis, the latter view is adopted, *i.e.* that the presence of glacigenic sediment is considered the defining criterion of paraglacial activity, irrespective of the time of its reworking.

Despite differences of opinion over its use, the paraglacial concept has been increasingly employed within geomorphology in recent years. This review chapter aims (1) to assess the validity of applying the paraglacial concept to hillslope adjustment; (2) to summarise current understanding of paraglacial slope adjustment; and (3) to identify the key areas requiring further research. In seeking to achieve these aims, the remainder of the chapter is structured into four parts. The first is a brief review of rock slope failure conditioned by glaciation/deglaciation. The second comprises a discussion of paraglacial modification of drift-mantled slopes. This includes (i) the development of the concept in the literature, (ii) delayed or renewed activity, (iii) the use of present and recent analogues as keys to understanding Late Pleistocene - Early Holocene paraglaciation, (iv) processes and rates of paraglacial activity, (v) paraglacial modification of hillslope form, and (vi) the identification and nature of paraglacial sediments in the field. The third part presents conceptual models for the nature of paraglacial hillslope modification, and the fourth and final part consists of an outline of the principal research questions that stem from previous research on paraglacial slope modification.

2.2 Rock slope failure conditioned by glaciation/deglaciation.

Glacier thinning and the consequent debuttressing of rock slopes appears to be an important factor favouring post-glacial rock slope deformation and failure (e.g. Radbruch-Hall, 1978; Gardner, 1980; Ballantyne and Eckford, 1984; Gellatly and Parkinson, 1994; Shakesby and Matthews, 1996; Wieczorek and Jäger, 1996; Abele, 1997), and probably played a key role in initiating large-scale rock failure following deglaciation at the end of the Pleistocene (Ballantyne, 1986a, 1991b, 1997; Cruden and Hu, 1993; Rapp and Åkerman, 1993; Marion *et al.*, 1995). Rapp (1960) estimated as much as 2 m yr^{-1} rockwall retreat at one site in Sweden from rock slope failures during deglaciation.

Radbruch-Hall (1978) discovered that large-scale gravitational creep of rock slopes is caused, at least in part, by the removal of lateral support of glacier ice from glacially-steepened valley sides during deglaciation, and Bovis (1990) attributed near-surface flexural toppling in British Columbia to the same mechanism during the retreat of glaciers from their 'Little Ice Age' maxima. Indeed, it seems that many of the large rockslides that have occurred in mid-latitude environments during historical time may have been conditioned by Late Pleistocene deglaciation (Cruden and Hu, 1993), and additional debuttressing of glacially-steepened slopes could result if glaciers continue to retreat in response to anticipated warming trends (Slaymaker, 1990; Evans and Clague, 1994).

In the Scottish Highlands, for example, the majority of non-rotational rock slope failures probably took place within 5,000 years after deglaciation, a pattern interpreted by Holmes (1984) as a 'delayed' response to rock slope weakening during glaciation/deglaciation. Holmes explained this in terms of progressive failure involving fissuring and shearing through intact rock bridges and asperities, possibly as a result of further changes in cleft-water pressure, which led to gradual

reduction in slope stability, and in some cases, failure. If this interpretation is valid, it appears that many active and recent non-rotational landslides may represent the long-term legacy of rock weakening that began during Late Devensian deglaciation. It seems that rock slope weakening left many steep slopes in the Highlands in a state of critical conditional stability such that seismic activity due to differential isostatic uplift after deglaciation may have triggered failure (Sissons and Cornish, 1982a, 1982b; Ringrose and Davenport, 1986; Ringrose, 1989a, 1989b ; Ringrose *et al.*, 1991; Ballantyne, 1997; Ballantyne *et al.*, 1998).

Indeed, Ballantyne (1986a, 1995b) identified nearly 600 rock slope failures in the Scottish Highlands, many of which lie close to the limits reached by glaciers during the Loch Lomond Stadial. This distribution implies that the triggering of such failures (mostly of Early Holocene age) represents in some way a response to the growth and decay of glaciers at that time. Three factors were identified which may have operated to reduce rock mass strength prior to failure: (1) slope steepening by glacial erosion; (2) high cleft-water pressures during deglaciation; and (3) progressive joint development. Ballantyne cited high-magnitude seismic activity (due to differential glacio-isostatic uplift of crustal blocks following deglaciation at *c.* 10 ka BP), as the most likely trigger mechanism. This would explain the prevalence of failures in the Early Holocene, soon after deglaciation, and the proximity of many failures in steep-sided glacial troughs to former glacier limits.

If this is true, rock slope failure may be seen as a paraglacial response (Ballantyne, 1991b, 1995b), being conditioned by differential loading under ice, and that the timing and causes of failure events are largely independent of Holocene climatic fluctuations. The concept of a slowly-declining threshold of stability, subject to intermittent perturbations (one of which may eventually

trigger slope failure) is a plausible explanation for paraglacial rock slope failure, and underlies the notion of progressive failure of debuttressed steep rock slopes during and after widespread deglaciation. However, given the high rates of talus debris accumulation and rockwall retreat under periglacial conditions during the Loch Lomond Stadial in Scotland, for example, (e.g. Ballantyne and Kirkbride, 1987), it is often difficult to disentangle paraglacial effects from periglacial effects in explaining the history of rock slope failure in formerly glaciated areas (André, 1997).

2.3 Paraglacial modification of drift-mantled slopes.

2.3.1 Introduction

As stated above, one of the principal aims of this chapter is to summarise current understanding of paraglacial hillslope adjustment, particularly in the context of drift-mantled slopes. This section represents the work most pertinent to the thesis, covering a number of strands necessary for an understanding of paraglacial drift slope modification. Since the development of the paraglacial concept in the early 1970s much of the work carried out in this field has been related to Late Pleistocene/Early Holocene paraglaciation, and it is only within the last few years that the concept has encompassed recent, delayed or renewed hillslope adjustment.

2.3.2 Origins of the paraglacial concept in the North American Cordillera

The notion of paraglacial adjustment of hillslope systems largely emerged from work carried out in the North American Cordillera, with much of the current understanding concerning paraglacial processes dating back to the influence of Ryder (1971a, 1971b; Church and Ryder, 1972). She defined the concept and

suggested an explanation for the timing, characteristics and processes responsible for paraglacial fan construction in formerly glaciated parts of British Columbia. Ryder demonstrated that alluvial fan construction was largely dependent upon temporary conditions resulting from deglaciation, when glacial drift was reworked by streams and debris flows to form fans significantly different from alluvial fans in arid environments.

Such activity seemed to have been determined by both the abundant supply of drift on unvegetated, glacially-steepened slopes, and by long duration, high discharge flow of meltwater. Her view that paraglacial reworking of sediment probably began very shortly after widespread deglaciation was supported by the discovery of many fan deposits directly overlying till with no evidence for intervening soil development. Ryder also demonstrated that limited accumulation had taken place on these fans since the deposition of a tephra layer dated to *c.* 6.6 ka BP, and inferred that such diminution of activity reflected both a decrease in sediment supply due to the development of vegetation cover, and a reduction in runoff. The timing of shifts from sediment deposition to incision of fans appeared to have been influenced significantly by regional uplift, and multi-generation activity evident elsewhere supported the view that uplift may prolong the total period of paraglacial effects. Ryder concluded that most of the fans studied represent relict landforms on which incision has replaced accumulation as the dominant mode of geomorphic activity. Thus in the context of the post-glacial environment, Ryder regarded paraglacial fans as temporary features which pass though a series of developmental stages, acting as intermediate stores for sediment passing through the basin catchment.

Further work in Canada has been concentrated in the Bow Valley, Alberta, where it appears that an intense episode of slope adjustment dominated by debris flow activity and fan development took place within *c.* 4 ka following Late

Wisconsin deglaciation at *c.* 12 ka BP (e.g. Roed and Wasylyk, 1973; Jackson *et al.*, 1982; Eyles *et al.*, 1988). Wilson and Churcher (1984) argued that such paraglacial sediment fills were laid down in an Early Holocene warm interval, given the rich large mammalian palaeofauna they discovered in the sediments. This general pattern of an intense but relatively short-lived phase of paraglacial activity following Late Pleistocene deglaciation has been widely recognised elsewhere in the western Cordillera (e.g. Smith, 1975; Luckman, 1981; Gardner *et al.,* 1983; Eyles, 1987; Eyles *et al.*, 1987; Church and Slaymaker, 1989; Butler and Malanson, 1990; Ashmore, 1993; Sauchyn, 1993; Beaudoin and King, 1994; Johnson, 1995; Levson and Rutter, 1995; Lian and Hickin, 1996), and also on slopes exposed by retreat of glaciers from their Neoglacial limits (e.g. Church *et al.,* 1979; Gardner *et al.,* 1983; Clague, 1986; Jordan and Slaymaker, 1991).

If it could be demonstrated that paraglacial sediments in formerly glaciated areas developed in a sequential manner, directly related to alternating glacial and nonglacial conditions, then they could be recognised as time-dependent phenomena, providing tangible links for deciphering Quaternary history. The presence of thick fan alluvium resting conformably on unweathered glaciofluvial deposits in the Nenana Valley, Alaska, supports such a developmental history, with fans forming in tributary valleys downstream from Pleistocene glacier margins immediately after outwash deposition ended (Ritter and Ten-Brink, 1986).

2.3.3 Application of the paraglacial concept in other environments

In addition to the North American Cordillera, evidence for paraglacial landscape change appears to be widespread in steep deglaciated valleys elsewhere. Radiocarbon dating of soils in the Peruvian Andes indicated that glacial tills were reworked and redeposited as fans after deglaciation at *c.* 12 ka BP, before

stabilising by 8 ka BP and being overlain by loess soils (Miller *et al.,* 1993). Similar evidence for paraglacial modification of hillslope drift in high mountain environments has been found in the Tian Shan (Solomina *et al.*, 1994), Japan (Suwa and Okuda, 1980, 1983; Okuda *et al.*, 1980), New Zealand (Pierson, 1980; McArthur, 1987; Blair, 1994), Norway (Blikra and Nesje, 1991; Blikra and Nemec, 1993), and in the dry, high-latitude landscape of Antarctica (Fitzsimons, 1990, 1996).

Investigation of mass-movement deposits in the Himalayan and Karakoram Mountains has also highlighted the importance of paraglacial processes in the evolution of these landscapes (e.g. Brunsden and Jones, 1984; Derbyshire, 1984; Derbyshire *et al.,* 1984; Fort *et al.*, 1989; Holmes, 1989; Owen and Derbyshire, 1989; Derbyshire and Owen, 1990; Owen *et al.,* 1995, 1996; Sharma and Owen, 1996): The Karakoram ranges are among the most tectonically active in the world, support numerous large glaciers, and contain some of the steepest and highest hillslopes on earth. Relative relief often exceeds 4000 m, and even the tributary valleys contain elevational differences of 2000 m in horizontal distances of only 1-2 km (Goudie *et al.,* 1984). These combined factors contribute to a glacial and fluvial environment dominated by processes of mass movement, most notably rockfall and debris flow, so that: "...the magnitude of relief, the overall steepness of the slopes and the scale of the debris accumulations provide an overwhelming sense of instability, mass movement and catastrophic events." (Goudie *et al.,* 1984, p. 359). However, the widespread signs of slope failure in this area generate two complicating factors when researching slope processes in the Karakoram. First, emphasis on the unusually high frequency and large magnitude of mass movement activity tends to mask the overall slow rates of operation of other geomorphic processes; and secondly, the high overall levels of slope instability make it difficult to distinguish accelerated paraglacial processes *sensu stricto* against background geomorphic activity

conditioned, for example, by rapid uplift, rapid fluvial incision and tectonic stresses.

Nevertheless, sedimentological analysis of slope and valley fills (e.g. Li Jijun *et al.,* 1984; Derbyshire and Owen, 1990; Owen, 1991) has revealed an alternating pattern of nonglacial and glacial sedimentation. Since the last glaciation (which is believed to have terminated at *c.* 45-60 ka BP), there appears to have been widespread replacement of glacial depositional surfaces by those of alluvial and debris-flow origin. For example, in a 12 km^2 area of the formerly till-covered Hunza valley, 5.23 km^2 is currently surfaced by fan debris, whereas just 1.69 km^2 is mantled by *in situ* glacigenic sediments (Li Jijun *et al.,* 1984).

As in Canada, most of the fans studied in the Karakoram are essentially ancient features, with little active aggradation. Indeed, Derbyshire and Owen (1990) found marked fan-head entrenchment and fan-toe truncation in the Hunza and Gilgit valleys, and on sedimentological grounds suggested that most fans are derived essentially from a few, large (perhaps catastrophic) mass transport events followed by modification by a greater number of smaller-scale events.

Not only do the observations made in the Karakoram accord with other work regarding the timing of the onset and termination of paraglaciation, but they also provide further evidence that large-scale paraglacial deposition may be rapid in response to shifts in geomorphic thresholds, especially rapid deglaciation. The youthfulness of some slope and valley fills on recently-deglaciated terrain certainly implies that post-depositional modification must have been extremely rapid, particularly during deglaciation, with slope movements supplying sediment more quickly than rivers could remove it (*cf.* Goudie *et al.*, 1984).

The legacy of paraglacial slope adjustment is also evident in Great Britain. Many of the landforms of the middle Findhorn Valley in Scotland, such as alluvial fans and alluvial terraces, have been viewed as being of paraglacial origin (Auton, 1990). Benn (1990, 1991, 1992b) found evidence that ice retreat in the Red Hills, Isle of Skye, was accompanied by slope instability on the newly-deglaciated valley sides, and suggested that reworking of glacigenic deposits represented paraglacial sedimentation at the margins of a retreating glacier. Slope activity was particularly high during the retreat of the Loch Lomond Readvance glaciers, as the result of the abundance of freshly-exposed unconsolidated sediment, and slope adjustment to subaerial conditions.

In Glen Roy, Lochaber, Peacock (1986) interpreted several large debris fans as subaerial paraglacial features. These fans are mantled by lacustrine silt, probably deposited when ice-dammed lakes flooded the glen during the Loch Lomond Stadial (*c.* 11-10 ka BP). The transition from sedimentation to fluvial incision after lake drainage suggests that these paraglacial fans formed within the 2-3 ka interval between Late Devensian deglaciation (*c.* 14-13 ka BP) and the Loch Lomond Stadial, with accumulation terminating as sediment became exhausted. Similarly, reconstruction of the evolutionary history of a fluvially-reworked debris cone in Glen Etive in the western Highlands revealed that after deglaciation (at *c.* 10 ka BP), paraglacial debris flow aggradation, stabilisation and pedogenesis were complete by *c.* 4.5 ka BP, again reflecting the importance of upslope sediment exhaustion in terminating paraglaciation (Brazier *et al.,* 1988). Whether the slope system permanently crosses system thresholds (Schumm, 1973, 1977) into a high sediment input regime, or whether it will recover to its former state would appear to depend in part on the rate of stabilisation and revegetation of the zones of sediment availability.

2.3.4 Present and recent analogues

Recently, a growing body of research has focused on rapid resedimentation of drift in glacier forelands deglaciated during the past few centuries (e.g. Jackson *et al.,* 1989; Mattson and Gardner, 1991; Zimmermann and Haeberli, 1992; Rickenmann and Zimmermann, 1993; Ballantyne and Benn, 1994, 1996; Winchester and Harrison, 1994; Ballantyne, 1995a; Harrison and Winchester, 1997). While documenting extensive erosion of recently-deglaciated drift slopes by debris flows in Norway, Ballantyne and Benn (1996) commented that: "An additional feature of previous studies is that dating of the most rapid period of paraglacial reworking of drift is often of poor resolution ... It is also notable that most studies of paraglacial resedimentation of glacigenic drift have concentrated on the nature and sedimentological characteristics of paraglacial landforms and deposits, with limited mention of the effects of paraglacial activity in modifying slope form." (Ballantyne and Benn, 1996, p. 1174)

In response to these limitations in earlier research, they aimed to establish the nature and timescale of paraglacial resedimentation and hillslope modification after recent glacier retreat at two sites, Fåbergstølsdalen and Bergsetdalen, in the Jostedal area of western Norway. They achieved this by establishing the ages of the onset and termination of debris movement, estimating the volume and rate of sediment accumulation, and by assessing the processes responsible for reworking and redeposition of drift (Ballantyne and Benn, 1994, 1996; Ballantyne, 1995a). Both sites were completely deglaciated at *c.* 9 ka BP, but reoccupied by glaciers that reached their maximum extent in the 18th century AD before retreating. Debris flow, triggered in part by snowmelt, was recognised as the principal agent of paraglacial reworking of glacigenic sediments, and analysis of flow deposit stratigraphy revealed a sequence of slope stabilisation and dormancy punctuated by phases of glacial/paraglacial sediment transfer.

Within 50 years of exposure by ice retreat, steep drift-mantled slopes in Fåbergstølsdalen have developed into gullied badlands, with extensive exposure of bedrock on upper slopes and redeposition of sediments as coalescing debris cones (Ballantyne and Benn, 1994). In contrast to the widespread gully activity at Fåbergstølsdalen, paraglacial debris cones at nearby Bergsetdalen have only formed downslope of rock gullies, and completely stabilised (with some sediment sources having been exhausted) within 100-200 years of deglaciation (Ballantyne, 1995a).

Ballantyne and Benn (1996) hypothesised that the different responses of the two sites to paraglacial resedimentation may reflect differences in initial slope gradient: the steep (*c.* 35°) drift slopes of Fåbergstølsdalen have experienced failure and gully incision along most of the valley side, whereas the slightly gentler slopes of Bergsetdalen have been eroded only where the crests are fed by water from gullies upslope. Moreover, the two Norwegian sites could be seen as representing different stages within the sequence of paraglacial slope adjustment, with unstable drift slopes at Fåbergstølsdalen continuing to experience adjustment, unlike the sediment-exhausted, inactive sites in Bergsetdalen. This conceptualisation of paraglacial slope modification is considered later.

Ballantyne and Benn (1994, 1996) conjectured that the paraglacial behaviour of slopes exposed by recent glacier retreat in Norway might offer a valid analogue for Late Pleistocene paraglaciation in Scotland and elsewhere, and that analogous relict landforms in the Scottish Highlands may have developed over similar timescales of centuries or even decades following Late Pleistocene deglaciation. Comparison of inactive gullied drift slopes and debris cones in the Scottish Highlands with areas of active or recent paraglacial modification in Norway suggests that rapid paraglacial reworking of valley-side drifts was of major importance in the post-glacial evolution of such areas, though not all such

slopes were modified in this way. Moreover, though some forms of paraglacial resedimentation (particularly fluvial reworking of glacial drift and progressive failure of rock slopes in large basins) may operate on a timescale longer than the duration of the Holocene (Church and Slaymaker, 1989), the suggestion made above is consistent with the view of those researchers who have proposed that unstable valley-side drifts may have been eroded, redeposited and stabilised within a timescale of centuries rather than millennia (e.g. Jackson *et al.*, 1982; Eyles *et al.,* 1988; Harrison, 1991, 1993; Wright, 1991).

2.3.5 The concept of delayed or renewed paraglacial slope adjustment

As stated earlier (section 2.1), the definition of paraglacial hillslope adjustment that is adopted influences perception of the concept and its ability to represent accurately environmental processes. Church and Ryder (1972, 1989) implied that deglaciation leaves inherently metastable slopes that are susceptible to rapid failure and/or reworking within the 'paraglacial period' during or immediately after deglaciation. Defining paraglacial hillslope activity in terms of glacigenic sediment supply implies the existence of slopes that may experience failure and/or reworking at any time after deglaciation, whether in the geomorphic short-term or as a 'delayed or renewed' paraglacial response in the medium-term (Ballantyne and Benn, 1996).

In the Scottish Highlands, the picture of an abrupt paraglacial pulse following Late Pleistocene deglaciation is complicated by evidence for delayed or renewed reworking of valley-side drift deposits within the past few centuries. From a total of 740 landslides or areas of sliding identified in Scotland by Ballantyne (1986a), 82 were classified as 'recent', with 69 involving reworking of drift or regolith, primarily by recent debris flow activity. In his extensive lichenometric study of Scottish debris flow deposits, for example, Innes (1983b)

found that over 70 debris flows in the Lairig Ghru (Cairngorm Mountains) occurred between 1970 and 1980 alone. Innes also maintained that there was no lichenometric evidence of hillslope debris flow activity prior to AD 1390, and that the majority of hillslope debris flows in Scotland have occurred within the last 500 years, and particularly within the last 250 years. His findings may be challenged on the grounds that later flows often obscure earlier ones, thus introducing a bias to his lichenometric dating, but it seems indisputable that a drastic increase in debris flow activity has occurred within the past few centuries; had the rates identified by Innes been sustained throughout the Holocene, much greater thicknesses of debris flow deposits would have accumulated than are now evident (Ballantyne, 1993). However, many of the debris flows investigated by Innes involved reworking of rockfall talus (*cf.* Luckman, 1992) rather than glacigenic drift, and hence cannot be regarded as manifestations of 'paraglacial' activity. The causes of delayed or renewed reworking of hillslope drift in Scotland remain elusive (Ballantyne, 1991a, 1991b).

Elsewhere there is evidence that some episodes of enhanced drift slope modification during the Holocene coincided with periods of general climatic deterioration. In a plethora of studies within Northwest Europe, (e.g. Kotarba and Strömquist, 1984; Starkel, 1984; Innes, 1985a; Rapp and Nyberg, 1988; Jonasson, 1991, 1993; Blikra and Nemec, 1993; Blikra, 1994; Nesje *et al.*, 1994, Alexandrowicz, 1997; Kotarba and Baumgart-Kotarba, 1997) the onset of renewed erosion of uplands during the Holocene is attributed to established episodes of climatic deterioration, particularly those that occurred around *c.* 3.0-2.5 ka BP and during the 'Little Ice Age' of the 16th - 19th centuries AD. On the basis of geomorphological and lichenometric surveys of debris cones and rockfall talus in Spitsbergen, André (1985, 1986) also inferred that debris fans and rockfall talus cones were built mainly during the 'Little Ice Age'.

However, the view that climatic deterioration is responsible for triggering episodes of enhanced debris flow activity rests partly on apparent coincidences in timing, and partly on the untested assumption that 'climatic deterioration' is widespread, synchronous and automatically associated with renewed or enhanced landscape instability in upland areas. The timing argument is difficult to sustain. Although some aspects of upland erosion appear to have been triggered or enhanced within the past few centuries, the available dating is insufficiently precise to relate enhanced mass movement activity to general 'Little Ice Age' climatic deterioration, or to dismiss other explanations.

It is possible, though, that enhanced paraglacial debris flow activity within the past few centuries was triggered in Northwest Europe (and possibly elsewhere) by a general increase in the frequency and ferocity of storms during the 'Little Ice Age' (Lamb, 1979, 1982, 1985). Exceptionally high magnitude rainstorms at this time may have initiated a general intensification of debris flow activity by triggering slope failure and lowering the threshold for subsequent events through the removal of vegetation cover (Brazier and Ballantyne, 1989). Although this idea is at best circumstantial, it suggests that the significance of the 'Little Ice Age' and earlier periods of climatic deterioration may have been indirect and local (Strunk, 1991), increasing the probability of destructive storms during such periods. This view is supported by physically-based modelling of the role of climate, vegetation and pedogenesis in affecting slope stability. Such modelling has demonstrated the susceptibility of particular soils to failure during intense rainstorms (Brooks, 1997; Brooks *et al.,* 1993b, 1995; Brooks and Richards, 1993, 1994). This suggests that recent failure and reworking of ancient drifts may partly reflect a gradual decline in slope stability due to progressive pedogenesis, and partly the influence of extreme rainfall events during the Holocene, representing the crossing of slope stability thresholds and leading to delayed reworking of drift.

Wells and Harvey (1987) found this to be true for drift slope fans in northern England.

In contrast to the view that the delayed or renewed reworking of drift slopes by debris flows during the Holocene represents a response to climate, others have emphasised the role of anthropogenic degradation or destruction of vegetation as a major cause of drift slope failure (e.g. Fairburn, 1967; Harvey *et al.*, 1981; Harvey and Renwick, 1987; Harvey, 1992; Strunk, 1997). In particular, Innes (1982, 1983b, 1983d) discounted progressive weathering and climatic change as causes of intensified Late Holocene debris flow activity on Scottish mountains, citing instead land-use changes (particularly burning and overgrazing) as possible causal factors, but his arguments are unsupported and seem unlikely on mountains where most debris flows originate in rock gullies. Stratigraphic investigations, pollen analysis and radiocarbon dating of buried soils within a debris cone in Glen Etive have, nevertheless, indicated a relationship between human disturbance and fluvial reworking of early Holocene debris flow deposits at *c.* 550 yr BP (Brazier *et al.,* 1988). Further, the observed increase in debris flow activity following contemporary forest clearance may provide a viable analogue to Holocene land-use change and associated slope activity (e.g. Pierson, 1980).

Uncertainty in elucidating the causes of delayed or renewed paraglacial erosion stems partly from 'casual causation' - the observed tendency to link erosion with particular causes only through assumed coincidence in timing and assumed knowledge of all possible causes and the mechanisms by which they operate. Indeed, in some cases it is impossible to refute the possibility that the timing of erosional events or episodes may be linked to events of random occurrence, and bears little relation to either the climatic or anthropogenic hypotheses. Matthews *et al.* (1986) cite uncertain links between climate and

process, imperfect dating, differential spatial sensitivity and the intermittent operation of different processes at different times as just some of the difficulties in ascribing a particular geomorphological response to a particular cause. Further, the relationship between revegetation and slope inactivity is particularly difficult to ascertain, representing something of a 'chicken and egg' problem, and it would appear that the combined influence of multiple factors in triggering debris slope failures will hamper any quantitative assessment of the relative role of any single factor.

2.3.6 Processes and rates of paraglacial modification of drift slopes

One of the most conspicuous legacies of drift slope modification in many mountain areas takes the form of debris cones or fans located along the flanks of formerly glaciated valleys. These features often reflect paraglacial reworking of potentially-unstable drift deposits, and whilst rock-slope landsliding may incise or bury drift mantles downslope, debris flows, fluvial activity and snow avalanches appear to be most important in contributing to redistribution of drift. The nature and importance of these processes in a paraglacial context is assessed below.

Debris flows

Debris flow or 'sediment gravity flow' has been defined as: "...a process by means of which granular solids, sometimes mixed with relatively minor amounts of entrained water and air, move readily on low slopes." (Johnson, 1970, p. 433). The term is employed here to refer to the rapid downslope flowage of poorly-sorted rock debris and soil, mixed with water and air, but it is also used to refer to the landforms produced by individual flows. Debris flows are neither restricted to talus or drift slopes nor to alpine or periglacial environments, but are regarded as being responsible for much of the reworking of glacigenic deposits in recently-

deglaciated terrain, and the consequent resedimentation of material as coalescing debris cones and fans (e.g. Ryder, 1971a, 1981; Church and Ryder, 1972; Lawson, 1979a; Jackson *et al.*, 1989; Derbyshire and Owen 1990; Ballantyne and Benn, 1994).

Debris flows have been variously classified according to a host of sedimentologic and behavioural parameters (e.g. Carter, 1975, Lawson, 1981b; Hansen, 1984; Shultz, 1984; Pierson and Costa, 1987; Coussot and Meunier, 1996), whilst Brunsden's (1979) classification concerns the scale and nature of the source area: catastrophic flows, hillslope flows and valley-confined flows. Catastrophic flows originate in a 'large-scale event' (Brunsden, 1979), and tend not to affect paraglacial hillslope modification. In contrast, hillslope and valley-confined flows are much smaller and less dramatic, but are associated with paraglacial reworking of glacigenic deposits. Hillslope flows are those that flow down an open hillslope, and are not topographically constrained. Valley-confined flows are confined for much of their length to a pre-existing gully or valley.

In general, debris flows display three distinguishable geomorphological zones (van Steijn *et al.*, 1995). The first comprises a source area intersected by steep chutes or ravines. Coarse material accumulates on the floors of these chutes, which are continued downslope by gullies cutting into talus or drift slopes. In general, erosion is the dominant process in this part of the system. Farther down the slope, levées are found on both sides of the track, the lower part of which may be sinuous, especially when slope angle is low. Transport and deposition coexist in this zone, but in its steepest parts erosion may still exceed accumulation. Gullies excavated in drift on the upper slope often decline in width and depth downslope, and levées usually tend to diminish in height towards the slope foot (Nieuwenhuijzen and van Steijn, 1990; Ballantyne, 1995b). Perhaps surprisingly, the debris flow track between the levées in the runout zone is often little affected

by the passage of flow, and may support flattened but otherwise undisturbed vegetation. The size of flow tracks varies widely, being related to the volume of the flow, the abundance and coarseness of debris and the length of the runout slope. In the Alps, for example, van Steijn *et al.* (1988) recorded flow lengths of 240-570 m and widths of 3-30 m, with mean gradients of 19-28°. Finally, there is a debris flow terminus where the levées form a frontal lobe, which may be complex in form due to the presence of several lobes or levées arranged beside or on top of each other.

Failure and flow occur when a rise in pore-water pressure causes a reduction in shearing resistance. Shearing resistance (SR) in unconsolidated debris can be expressed as:

$$SR = C' + (W \cos \alpha - u) \tan \phi'$$

where C' is the effective cohesion of the debris, W is the weight of debris at a potential failure plane, α is the slope angle, u is pore-water pressure and ϕ' is the effective friction angle of the debris. If pore-water pressure increases, the frictional strength of the debris [$(W \cos \alpha - u) \tan \phi'$] is reduced. If the overall shearing resistance falls below the shearing force generated by the downslope component of the weight of debris ($W \sin \alpha$), then failure occurs.

In cold, humid mountain environments, this happens in a number of ways. Many flows begin as shallow planar slides on steep slopes (Rapp and Nyberg, 1981; Innes, 1983a; Zimmermann and Haeberli, 1992) which rapidly disintegrate through dilatancy or by liquefaction to become a flow (Costa, 1984). Alternatively, Gardner (1983) observed that in steep gullies in the Rocky Mountains, flood torrents may be transformed into valley-confined debris flows by the addition of debris from the channel bed and sides, and van Steijn *et al.*

(1988) invoked failure of debris dams in rock gullies in the French Alps as the main cause of flow initiation.

Many debris flows are triggered by intense rainstorms that cause a rapid rise in pore-water pressures (e.g. Baird and Lewis, 1957; Jahn, 1976; Rapp and Strömquist, 1976; Larsson, 1982; Harvey, 1986; Addison, 1987; Rapp, 1987; Reid *et al.*, 1988; Mathewson *et al.*, 1990). Caine (1980) attempted to define boundary conditions of rainfall intensity and duration that control the onset of slope failure of the debris flow type, and plotted this as a limiting curve which has the form:

$$I = 14.82\ D^{-0.39}$$

where I = rainfall intensity (mm hr^{-1}), and D = duration of rainfall (hr). This threshold can be regarded as only an approximate indicator of slope failure conditions, however, as other variables including antecedent soil moisture conditions, hydraulic gradient, slope angle, and soil structure, depth and texture all influence propensity for failure. The importance of antecedent moisture conditions was illustrated by Zimmermann and Haeberli (1992), who recorded more than 600 debris flows in the Swiss Alps during two separate events in 1987. In the first, rainfall totals reached 100-180 mm at the time of debris flow initiation, but intensities seldom exceeded 12 mm hr^{-1}. The second period of debris flow activity followed a total of 120-130 mm of rain, during which intensities reached 40 mm hr^{-1}.

Saturation of debris by glacial outburst floods, rapid snowmelt or melting ground ice may also initiate debris flows (e.g. Sharp and Nobles, 1953; Jackson, 1979; Clague *et al.*, 1985; Lundqvist, 1988; Harris and Gustafson, 1988, 1993; Desloges and Church, 1992; Catto, 1993; De Graff, 1994). Lawson (1979a, 1981a, 1981b, 1982) attributed the disaggregation and flow of ice-cored drift at

the snout of Alaska's Matanuska Glacier to excess pore-water pressures and seepage pressures associated with ice ablation, and Ballantyne and Benn (1994) recorded the occurrence of debris flows triggered by the rapid melt of snow at the heads of drift gullies in western Norway. Johnson's (1971, 1984b, 1995) findings that debris flow sediments in the Canadian Yukon were frequently deposited against or over snow or ice masses are significant with respect to deposit stability, because under changing climatic conditions such deposits may become conditionally unstable. In particular, degradation of permafrost in alpine environments as a result of climatic warming may make slopes more susceptible to debris flow activity (Zimmermann and Haeberli, 1992; Schlyter *et al.*, 1993; Haeberli, 1996).

However, before generalisations about the processes responsible for debris flow activity can be made, individual site factors must be taken into account. These may mask any general trends by enhancing susceptibility of some areas to debris flow activity (Brazier, 1987; Blikra and Nesje, 1997). Innes' (1982, 1983c) work on debris flows in Scotland illustrated the influence of local lithology on debris flow activity, which tends to be more prominent on rocks that yield cohesionless sand-rich regolith, such as sandstone and granite, than in schist areas of silt-rich regolith cover. The susceptibility of areas of sandy drift or regolith to flow may reflect the associated high infiltration rates, which permit a rapid rise in the water table during intense precipitation (Ballantyne, 1986a), or the cohesionless nature of sandy debris and absence of a complete vegetation cover at high altitude. A further important local control on the distribution of debris flow activity is sediment availability (*cf.* Statham, 1976a; Strachan, 1976), with glacially-scoured areas of Scotland such as Knoydart, Morar and Morvern offering only the thinnest of drift cover and hence limited debris flow activity (Ballantyne and Harris, 1994).

Nevertheless, it appears that ancient paraglacial debris flow activity may have resulted from the progressive reduction of shearing resistance of debris under a variety of circumstances following deglaciation. Large, isolated flow events may reflect the crossing of an internal threshold of the slope system, and need not have any obvious long-term climatic significance (Nyberg and Lindh, 1990). After deglaciation, intense and prolonged rainfall, snowmelt and high antecedent moisture conditions are all likely to have contributed to elevated pore-water pressures on unvegetated, steep slopes that were possibly rendered less stable as a result of seismic shocks and vibrations associated with fault movements due to glacio-isostatic unloading (Kotarba, 1992; Ballantyne, 1995b).

The way in which debris flows move is also contentious. Johnson (1970) and Johnson and Rodine (1984) have advocated a visco-plastic rheological model in which a raft or 'plug' of debris is carried downslope by laminar flow along the sides and base of a channel. Van Steijn *et al.* (1995) believe that this model is suitable for highly concentrated (cohesive) flows with a relatively high clay content, particularly where slope angle is low. Others, however, have suggested a flow model in which dispersive pressures caused by inter-particle collisions maintain the mobility of the material (Takahashi, 1978, 1980; Carling, 1987), especially where the debris is more dilute and low in clay content (Innes, 1983b). The transport of large boulders within debris flows has been variously attributed to buoyancy, the viscosity of the fluid phase, excess pore-water pressures and dispersive pressures within the flow (e.g. Hampton, 1975, 1979; Rodine and Johnson, 1976; Pierson, 1981; Johnson and Rodine, 1984). The concentration of fines near the base of flows results from 'kinematic sieving' (van Steijn *et al.*, 1995), whilst coarse debris tends to move upwards within a mobile flow, and is then carried to the front by greater surface velocities (Takahashi, 1981). Concentrations of boulders may temporarily dam the flow, so that debris flows tend to move in a series of short surges (Campbell, 1974; Okuda *et al.*, 1980),

with velocities of up to 16 m s^{-1} (Curry, 1966). The deposition of levées at the margins of flow tracks probably reflects lower velocities at the edge of flows, though it has been attributed to bulldozing by the advancing flow (Sharp, 1942) or dispersive sorting (Pierson, 1980).

Surface runoff and snow avalanching

Drift-mantled slopes are modified not only by debris flow, but also by concentrated surface runoff and snow avalanching. Strömquist (1983, 1985) considered the importance of surface wash erosion of drift, which can lead to the development of steep-sided gullies which redistribute sediment downslope. Such water movement may occur within the tracks of pre-existing debris flows, and fluvial facies have been deposited within debris cones at the foot of slopes (e.g. Derbyshire and Owen, 1990). Wash deposits are often of finer sediment than those of debris flows, and sorting may be present. In sections through paraglacial 'alluvial' fans in the Canadian Rockies, Eyles and Kocsis (1988) and Eyles *et al.* (1988) observed layers of massive and crudely-bedded sands and gravels, which they interpreted as fluvial facies intercalated between debris flow units. In some cases, these facies were recognised as the upper surfaces of massive debris flows, reworked by fluvial processes. The intercalation of slopewash horizons and massive debris flow deposits has also been widely reported (e.g. Brazier and Ballantyne, 1989; Ballantyne and Benn, 1994).

In many mountain areas, snow avalanches play a key role in modifying the form of drift slopes and in transporting debris downslope, particularly in areas of gentle slopes, where other erosional processes of similar magnitude may be absent because they require steeper gradients to be initiated (Nyberg, 1984). The importance of snow avalanching as a geomorphological agent was first illustrated by Rapp (1960), although avalanche transport is probably of less importance than

debris flows in reworking sediments on recently-deglaciated terrain in most mountain environments. The most widespread geomorphological effect of wet snow avalanches is the erosion of debris from the upper parts of debris slopes, and its redeposition further downslope (Nyberg, 1989). Snow avalanche paths frequently show evidence of debris flowage (Kostaschuk *et al.*, 1986; Sauchyn, 1986), and Luckman (1992) considered that the role of snow avalanching might be enhanced by a symbiotic relationship between debris flow and snow avalanche activity, whereby debris flows supply avalanche tracks with loose debris for later removal.

Snow avalanches are a very specific type of mass flow, in which the supporting matrix, the snow, disappears from the deposit soon after its emplacement. Consequently, snow avalanche deposits are recognisable by several characteristic features, which Blikra and Nemec (1994) listed as: predominantly angular, clast-supported and openwork gravel texture, with large interstices filled with a secondary, water-lain sandy matrix; large clasts relative to bed thicknesses; apparently disorganised clast fabric; uneven and often discontinuous bed geometries; and composite sheet-like beds comprising debris deposited by successive avalanches. Other characteristics of snow-avalanche deposits include linear ridges of debris, debris tails on the lee sides of surficial blocks, lichen patches, boulder holes, scratch marks and melt-out clasts perched in unstable positions atop other boulders (Ward, 1985; Ballantyne, 1995b).

Rates of operation

Discussion of the rates of geomorphological processes requires some caution, in that the literature on the topic is inevitably biased towards sites of pronounced activity and high-magnitude events. Concepts of landscape sensitivity, persistence and resilience must be held in balance with emphasis on

process mechanisms, and a reliable estimation of the spatial and temporal extent of paraglacial activity sought. Nevertheless, it is undeniable that debris flow represents one of the most effective agents of sediment transfer associated with steep slopes in mid- to high-latitude mountain environments. In his classic investigation of geomorphic processes in the Kärkevagge area of northern Sweden, Rapp (1960) made a continuous study of the types, location, frequency and quantitative importance of several slope processes over an eight-year period. He found that debris flows accounted for about 46% of the total volume of debris moved by rapid mass movement processes, moving on average 20 t km^{-2} yr^{-1}, and that 'dirty avalanches' moved on average 14 t km^{-2} yr^{-1}. In comparison, rockfall activity moved just 1-6 t km^{-2} yr^{-1}. Similar monitoring work by Gardner (1979, 1983) in the Canadian Rocky Mountains demonstrated that the pattern of activity differs spatially, depending on both the magnitude of debris flows in terms of sediment load transported and their frequency. Similar conclusions were reached by Innes (1985a) on the basis of research in Norway and the Scottish Highlands.

It appears that the frequency of major debris flow events tends to be greater in mid-latitude mountains than in arctic environments. For example, André (1985, 1986) tentatively suggested that rainfall-triggered debris flows in Spitsbergen recur every 100-300 years, and Rapp and Nyberg (1981) estimated that 50-400 years is typical for Swedish Lappland. In contrast, a recurrence interval of 10-40 years was recorded for major events in the Swiss Alps (Nieuwenhuijzen and van Steijn, 1990), and Selby (1976) calculated a 30 year interval for flows in New Zealand. Although for recurrence interval comparisons to be of value, basin conditions such as sediment supply and the hydrometeorological regime must remain constant (Jackson *et al.,* 1989), these contrasts probably reflect the greater frequency of extreme rainfall events in many mid-latitude mountains.

In their work on recently-deglaciated slopes in western Norway, Ballantyne and Benn (1994) found that ground surface lowering by paraglacial debris flows at Fåbergstølsdalen has averaged 2.5-4.7 m in less than 50 years over much of the gullied study area, equivalent to a minimum erosion rate of 50-100 mm yr^{-1}. This compares with results from cone surveys at Bergsetdalen which implied average rates of sediment accumulation of 8-44 mm yr^{-1}, with an inferred average rate of drift removal of 37-94 mm yr^{-1} at one site (Ballantyne, 1995a). These overlapping ranges of rates of paraglacial gully erosion at the two locations suggest that they may be regarded as being reasonably representative for steep drift-covered slopes in western Norway during the decades immediately after recent ('Little Ice Age') deglaciation. It should be stressed, however, that the above rates represent averages. It is possible that sedimentation rates were markedly higher immediately after deglaciation, and declined through time as sediment sources became progressively exhausted. Nevertheless, these values highlight the potentially high rates of paraglacial erosion and surface lowering, and are worth contrasting with Selby's (1982) estimated average regional rate of ground lowering of 0.1 mm yr^{-1} for temperate, alpine environments.

2.3.7 Paraglacial modification of slope form

Documented changes in the form of valley-side slopes that have resulted from specifically paraglacial activity are limited to Ballantyne and Benn's (1994, 1996) work in western Norway. At Fåbergstølsdalen they calculated that over a period of 50 years two effects of paraglacial erosion and redeposition on slope form were evident: first, deep incision of the upper part of the slope, causing a lowering of the gradient of the upper rectilinear slope by *c.* 5° and exposing bedrock at the head of the slope; and second, partial infill of the basal concavity by reworked sediment. The overall result was therefore a gentler drift slope with

an average reduction of gradient from a mean angle of *c*. 35° to a mean angle of *c*. 30°, and a less pronounced overall concavity (Ballantyne and Benn, 1994).

Although debris flows have been initiated in fluvial channels of *c*. 13° under increased runoff conditions (Takahashi, 1981; Haeberli, 1996), according to Takahashi (1981), the lower threshold for hillslope debris flow initiation is *c*. 27°, so whilst debris flows at Fåbergstølsdalen are likely to continue to excavate sediments that accumulate in the gullies due to the collapse of gully walls, Ballantyne and Benn suggested that further paraglacial modification of gully floor gradient may in fact be self-limiting, culminating in the evolution of 'final form' valley slopes and the termination of paraglacial slope modification. Future modification of drift slopes may therefore take the form of gully widening and coalescence, with progressive erosion of interfluve 'arêtes' and associated accumulation of debris on the lower slopes. Finally, Ballantyne and Benn hypothesised that the exposure of bedrock at the head of the oldest gully system at Fåbergstølsdalen may suggest that the final form of paraglacially-modified valley sides may consist of three parts. These are: (1) an upper bedrock slope, (2) a mid-slope area of reworked sediment that takes the form of coalescing debris cones with gradients below 25-30°, and (3) a slope-foot zone of debris flow deposits overlying till or bedrock.

This idea regarding the termination of slope form modification accords with earlier work by Harvey *et al.* (1981). They suggested that the cessation of late Holocene drift modification may have been caused by a reduction of slope profile: as incision and headward retreat of gullies takes place, and gully floors cut down to bedrock, profiles would reduce in gradient, resulting in the backfilling of debris up into the lower parts of gullies, subsequent reduction of gully erosion, and eventual revegetation and stabilisation.

2.3.8 The nature of paraglacial sediments

Paraglacial sediments are notoriously difficult to distinguish from *in situ* drift sediments, largely because they tend to retain many of the sedimentological characteristics of the unreworked parent sediments. Furthermore, both *in situ* glacigenic sediments and reworked tills often display common matrix and clast characteristics, reflecting similar particle orientation mechanisms and sediment source characteristics (Pe and Piper, 1975; Boulton and Paul, 1976; Lawson, 1988; Holmes, 1989; Ballantyne and Benn, 1994; Benn, 1994a, 1994b; Owen, 1994). Therefore, it seems likely that some superficial deposits previously mapped as tills (especially flow tills) may in fact be paraglacial debris flow facies (e.g. Wright, 1983, 1991; Eyles *et al.*, 1988; Lawson, 1988; Ballantyne and Benn, 1994; Fernlund, 1994; Owen, 1994), and that the extent of paraglacial resedimentation of till facies may have been greatly underestimated (Dardis *et al.*, 1994).

The characteristics of paraglacially-reworked glacigenic sediments have been described by several workers (e.g. Theakstone, 1982; Derbyshire *et al.*, 1984; Wilson and Churcher, 1984; Eyles *et al.*, 1988; Lawson, 1988; Campbell and Evans, 1990; Johnson and Hansel, 1990; Merritt *et al.*, 1990; Owen, 1994; Ballantyne and Benn, 1994). Although there is some limited evidence for the usefulness of particle size analyses in distinguishing unmodified and paraglacially-reworked tills (e.g. Landim and Frakes, 1968), other researchers have suggested that remobilised sediments are distinguishable from *in situ* deposits only in terms of clast fabric and large-scale structures (e.g. Ballantyne and Benn, 1994). Particle alignment in debris flow deposits tends to show a preferred downslope or down-flow orientation (Innes, 1983a; Mills, 1984), whereas unmodified clasts in basal tills are often strongly aligned down-valley in the direction of glacier movement.

Where exposed in section, individual debris flow units often take the form of massive, ungraded diamictons, with clasts of various sizes embedded within a matrix of sand- or silt-rich fines. Stratigraphic characteristics may include lenticular structures visible in transverse sections, the alternation of openwork structures with clast- or matrix- supported elements, and localised inverse grading. Movement may cause shearing or deformation of underlying sediments, particularly in high-density, low-viscosity flows (Lawson, 1988). Individual contacts may be defined by discontinuities or by thin beds of silt or sand that reflect surface wash following flow immobilisation (Nieuwenhuijzen and van Steijn, 1990).

However, Lawson's work (e.g. 1982) illustrated how debris flow deposits vary according to flow type. His type I deposits are generally structureless, reflecting the lack of internal deformation, whereas grain support in the shear zone of his type II flows may be by liquefaction, grain interaction, or transient turbulence, and consequently such flows may show normal grading at the base. At high clast concentrations, inertial grain interaction may produce inverse grading (Shultz, 1984). For type III flows, where the shear zone extends the full thickness of the flow, inverse grading may be evident throughout. Finally, in type IV flows, grain support is primarily by grain interaction and pore fluid movement, resulting in a fine-grained normally-graded deposit. Basal traction gravels may be present in flow types II, III and IV.

Furthermore, detailed study of the sedimentary properties of recent debris flow deposits in the French Alps has revealed downslope variations within the deposits, especially between channel-side and outer parts of levées and between levées and lobes (van Steijn, 1988; van Steijn *et al.*, 1988; Nieuwenhuijzen and van Steijn, 1990; Bertran and Texier, 1994; van Steijn *et al.*, 1995). These studies concern changes in levée height, mean clast size and the general stratification

pattern, and emphasise the need for careful analysis of stratified glacigenic slope deposits, and the specification of sample-point locations (Young, 1969).

It is not only *in situ* glacigenic deposits which are easily mistaken for paraglacial deposits, for the sedimentology of periglacial solifluction deposits may differ only slightly, if at all, from paraglacial slope deposits. Whilst this discussion lies outside the scope of this thesis, it is developed in recent work (e.g. Wright, 1991; Harrison and Winchester, 1992; Bertran, 1993; Harris, 1996; Harrison, 1996; Harris, 1998).

2.4 Paraglacial hillslope modification as a conceptual model.

The underlying assumption of the paraglacial concept as formulated by Church and Ryder (1972) is that deglaciation represents a fundamental change in the terrestrial erosional environment. Sediment that may have attained stability within the glacial depositional environment may be potentially unstable with respect to the subaerial environment that succeeds it. Thus proglacial and post-glacial rivers excavate glacial sediment at rates far in excess of the 'normal' sediment supply expected in nonglacial environments. For periods of several thousands of years or more, sediment yield of deglaciated basins may be unrelated to the primary production of debris by nonglacial processes, and consequently fail to conform to either dynamic equilibrium models of landscape evolution (Johnson, 1984a; Slaymaker, 1984; Church and Slaymaker, 1989) or conventional concepts of drainage basin sediment delivery (Church *et al.*, 1979; Leonard, 1986a; Harbor and Warburton, 1993; Ashmore, 1993; Leemann and Niessen, 1994). This situation was encapsulated by Church and Slaymaker (1989) when they noted that: "...the natural landscape of British Columbia is imprisoned in its history." (Church and Slaymaker, 1989, p. 453). In this sense, the paraglacial

concept of landscape change is a specialised example of relaxation time (Luckman and Fiske, 1997).

Although caution is required when commenting on concepts of geologic 'equilibrium' following short-term observation (especially when thresholds are exceeded on an unquantified duration-frequency series), Church and Ryder (1972) suggested a tentative model (Figure 2.1) to represent the paraglacial sediment cycle associated with fluvial sediment transport. This contrasts with the hypothetical geological 'norm' of denudational processes where erosion, transport and deposition are supposedly in 'equilibrium'. Although tectonic and climatic instability render their paraglacial concept hypothetical for large areas, Church and Ryder's model is useful in indicating that under paraglacial sedimentation there will still be a considerably higher sediment yield during deglaciation than long-term 'normal' considerations might indicate.

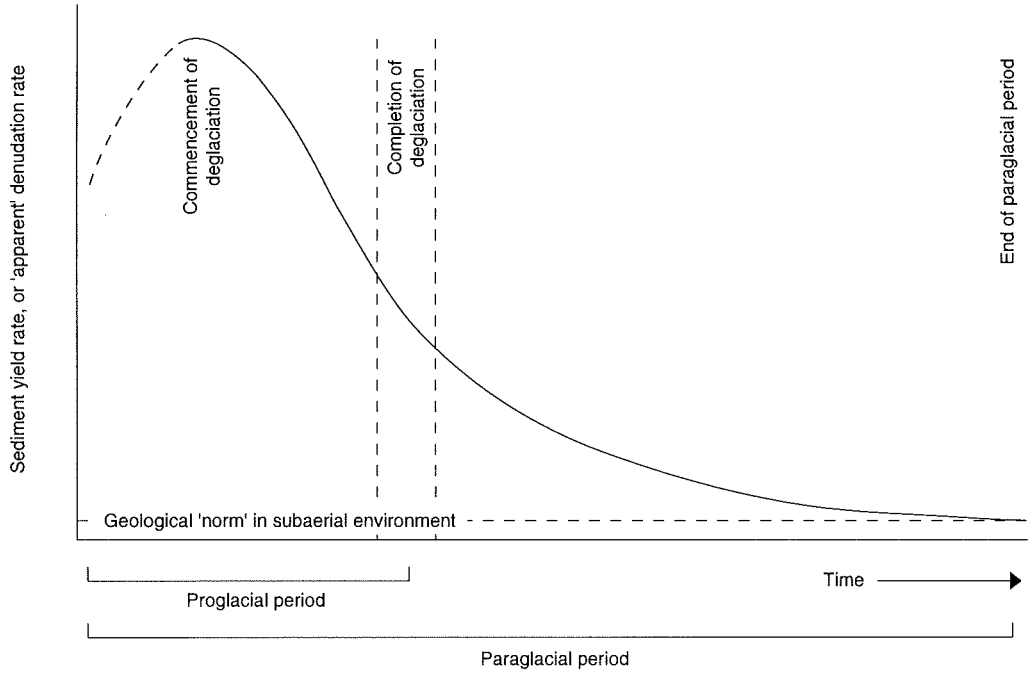


Figure 2.1. The paraglacial sedimentation cycle (after Church and Ryder, 1972).

The spasmodic character of sedimentation and erosion was well-illustrated by Clague (1986, Figure 10), who recognised that the terrestrial Quaternary stratigraphic record of British Columbia is a product of brief sedimentation events separated by long intervals of nondeposition and erosion. Most sediments in valleys were deposited during periods of ice sheet growth and deglaciation. In contrast, during nonglacial periods, sedimentation was more restricted and occurred at lower rates, preceded by a short episode of valley incision as streams adjusted to nonglacial conditions.

More recently, a study of Holocene sediment yields of rivers in British Columbia by Church and Slaymaker (1989) concluded that the length of the 'paraglacial period' is dependent on spatial scale, increasing in duration with basin size. In upland catchments close to sediment sources, sediment yields can be expected to peak as deglaciation commences, but in more distal reaches, sediment throughputs should peak later as the sediment travels through the system (Figure 2.2). Thus in downstream reaches of large catchments, sediment loads may not begin to increase until after peak yields in proximal reaches. Furthermore, upland paraglacial cycles may be largely complete within a few centuries of deglaciation, whereas larger basins may still be responding to deglaciation for several millennia. Benn and Evans (1998) suggest that it is possible for a landscape to fail to reach its nonglacial sediment equilibrium before the onset of the next glaciation.

Harbor and Warburton (1993) also explored this idea graphically, plotting hypothetical sediment yield variations over time for three different basin scales (Figure 2.3). Two important points emerged from this work: first, basins with an identical sediment yield may vary greatly in size, because of their positions on their respective curves (for example S in Figure 2.3); second, for sediments measured at different times (T_1, T_2 and T_3 in Figure 2.3), the relative magnitude

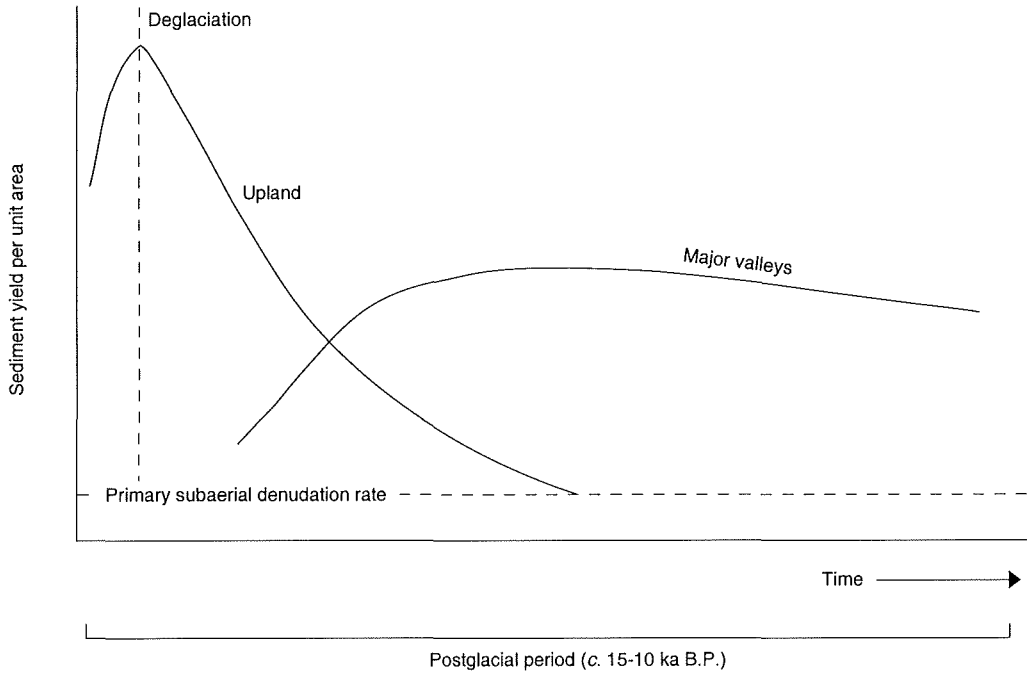


Figure 2.2. The paraglacial sedimentation cycle, as modified by Church and Slaymaker (1989) to account for the effect of spatial scale on the temporal pattern of sediment yield.

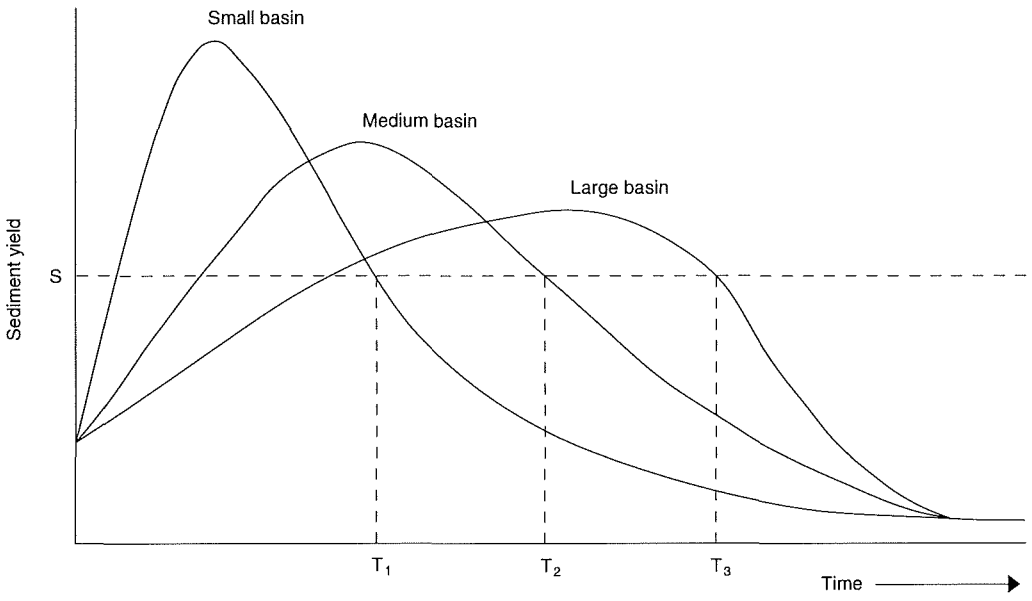


Figure 2.3. Paraglacial sedimentation cycles for drainage basins of different scales, given as hypothetical sediment yields over time (after Harbor and Warburton, 1993).

of sediment yield from basins of different sizes will vary.

If a drainage basin has been totally deglaciated, the fluvial notion of paraglacial sedimentation is relatively straightforward, with sediment rates declining and eventually reflecting 'normal' postglacial denudation rates, as illustrated in Figure 2.1. However, the persistence of glaciers in many alpine areas complicates this definition and understanding of the temporal nature of paraglacial sedimentation (Brooks, 1994). In such a setting, paraglacial sediments have been continuously introduced into the hillslope and fluvial systems over the postglacial period, so that paraglacial sedimentation has not ceased. Exposure of fresh drift following successive periods of Neoglacial glacier retreat thus causes rejuvenation of the reworked glacigenic component, often producing discernible surges in postglacial sedimentation rates, as observed by Brooks (1994) in British Columbia. Brooks distinguished extensive paraglacial sedimentation arising from Late Pleistocene deglaciation (the 'transitional' paraglacial period) from more localised paraglacial activity attributable to intermittent Neoglacial fluctuations (the 'persistent' paraglacial period). Brooks suggested that the shift from the 'transitional' to the 'persistent' paraglacial period occurs when fluvial sedimentation reflects only 'normal' postglacial denudation of the landscape, although both paraglacial periods may be superimposed upon each other. This concept is illustrated in Figure 2.4, and must be recognised when investigating paraglacial sedimentation in any drainage basin containing glacier ice.

One of the attractions of the paraglacial concept is that by considering conditioning of nonglacial processes by glaciation (or deglaciation), it allows for a long-term historical-based, conceptual explanation of landscape evolution and modification as opposed to the short-term, more empirically-based cause and effect approach. However, a major question facing current research on paraglacial slope modification relates to the nature and timing of the paraglacial response. On

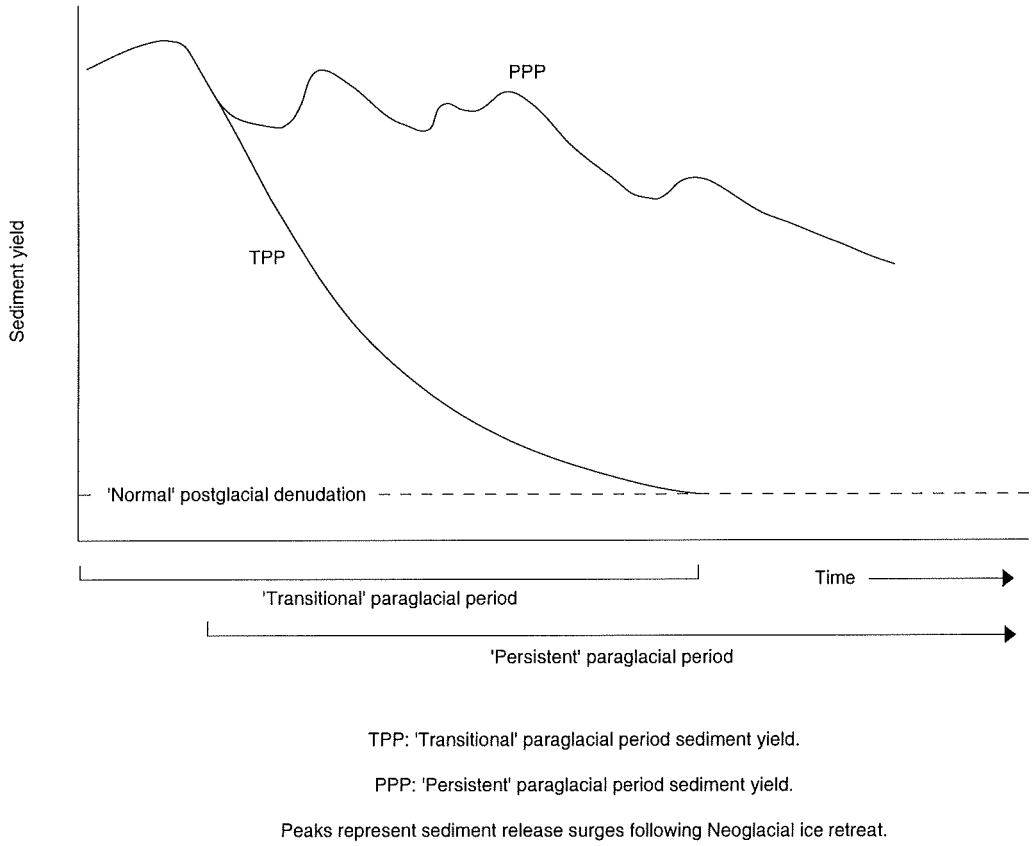
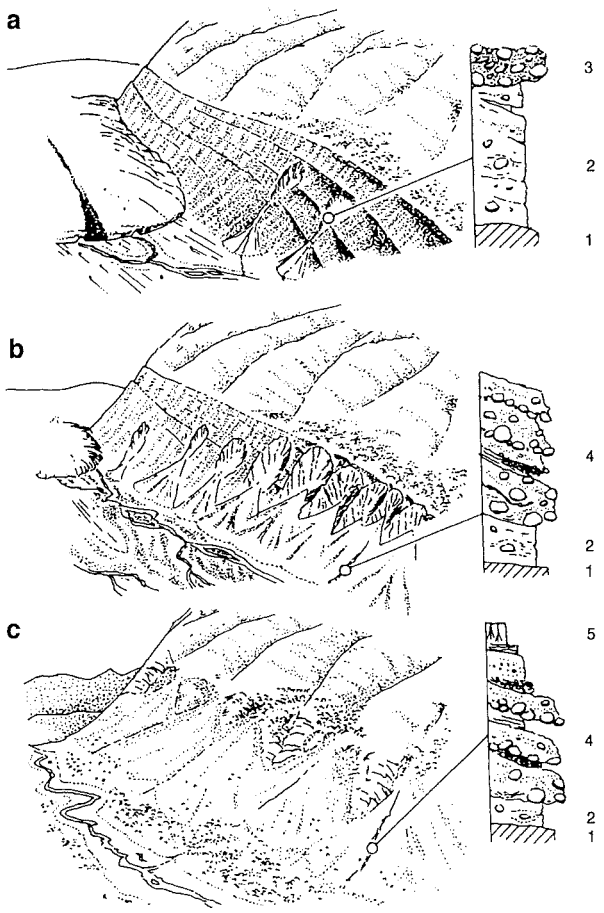


Figure 2.4. Hypothesised 'Transitional' and 'Persistent' paraglacial periods and associated paraglacial sedimentation (after Brooks, 1994).

the basis of work undertaken in Norway, Ballantyne and Benn (1996) have suggested that the behaviour of drift slopes experiencing paraglacial adjustment may follow an evolutionary pattern (Figure 2.5). In their model, gully incision (Figure 2.5a) is followed by gully widening and progressive debris cone accumulation (Figure 2.5b), and the subsequent erosion of the interfluves between gullies (Figure 2.5c). It predicts that the steepest valley-side slopes will be those most recently exposed. Ballantyne and Benn perceived their two Norwegian sites as representing different stages in the process of paraglacial slope adjustment, with unstable drift slopes at Fåbergstølsdalen continuing to experience activity, compared to the sediment-exhausted inactive slopes at Bergsetdalen. This



(a) Initial slopes exposed by glacier retreat, showing lateral moraines and the onset of gully incision.
(b) Advanced gully development and the deposition of coalescing debris cones or fans downslope.
(c) Exposed bedrock and stabilised, vegetated gully systems at the head of largely relict debris fans. By this final stage, paraglacial sediment reworking and slope adjustment have effectively ceased owing to the diminution of sediment supply.

Facies key: 1. Bedrock; 2. Subaerial sediments relating to an earlier cycle of paraglacial sediment reworking; 3. Ice-marginal deposits; 4. Paraglacially-reworked sediments: debris flow deposits and intercalated slopewash deposits; 5. Soil horizons.

Figure 2.5. Schematic representation of landforms and sedimentary facies associated with three stages in the paraglacial reworking of steep drift slopes in glaciated valleys (Ballantyne and Benn, 1996).

sequence represents the response of the geomorphic system to a stimulus, and its readjustment by feedback mechanisms to a new stable form (*cf.* Schumm, 1977).

By plotting the rate of sediment transport against time for these sites it is possible to describe the possible sequence of paraglacial landscape evolution (Figure 2.6). In Bergsetdalen, hillslope adjustment was limited both spatially and temporally, with activity being terminated by the exhaustion of sediment supply

(Figure 2.6a). At Fåbergstølsdalen, slopes did not respond immediately after deglaciation, but were suddenly and rapidly degraded by paraglacial processes, which are still active though possibly declining in rate. It may also be possible to build on this sequence by adding the hypothetical responses of longer term delayed (1), renewed (2) or even cyclical (3) drift slope adjustment (Figure 2.6b).

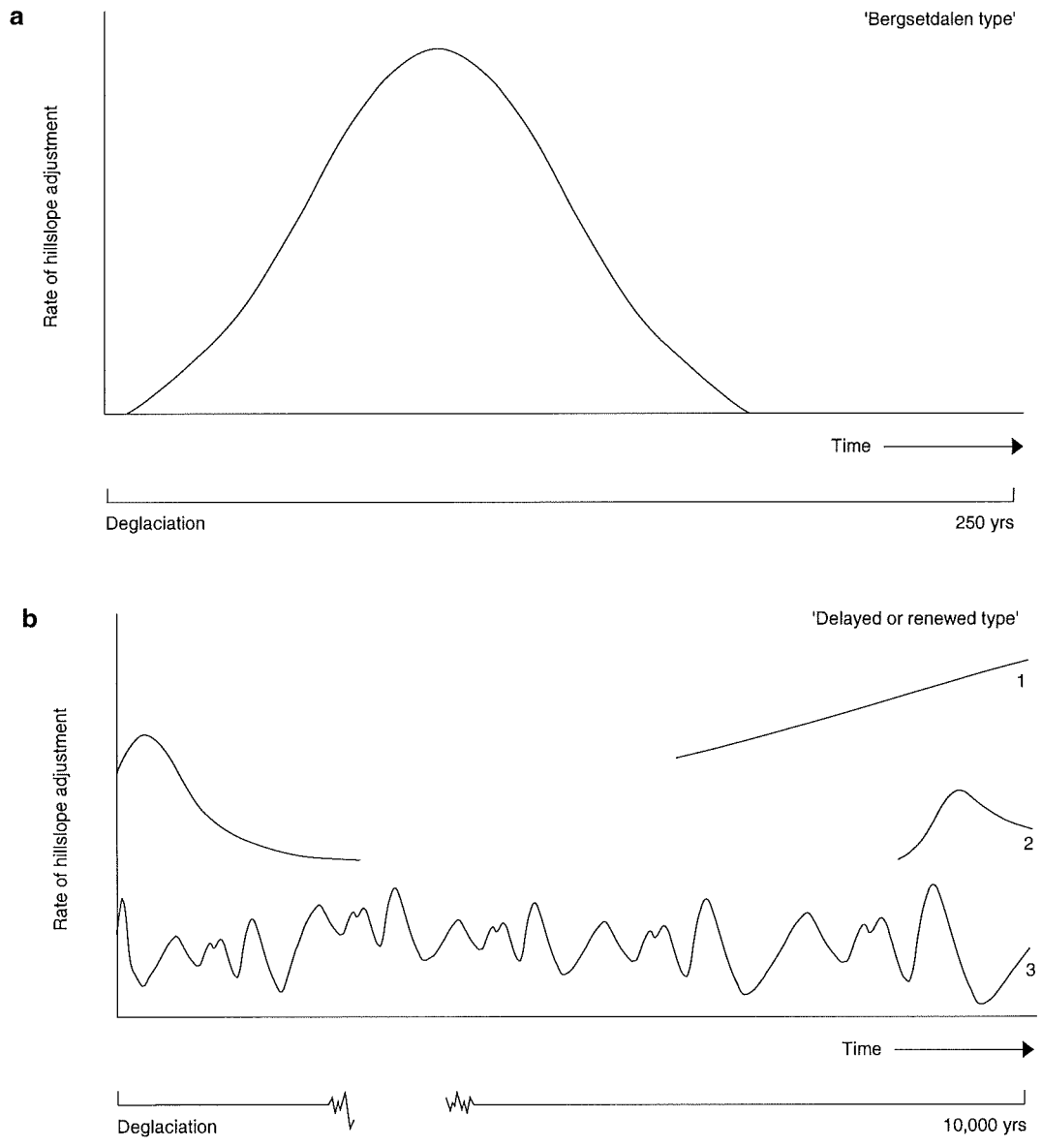


Figure 2.6. Two possible forms of paraglacial hillslope adjustment through time.

This developmental sequence may also be considered conceptually in

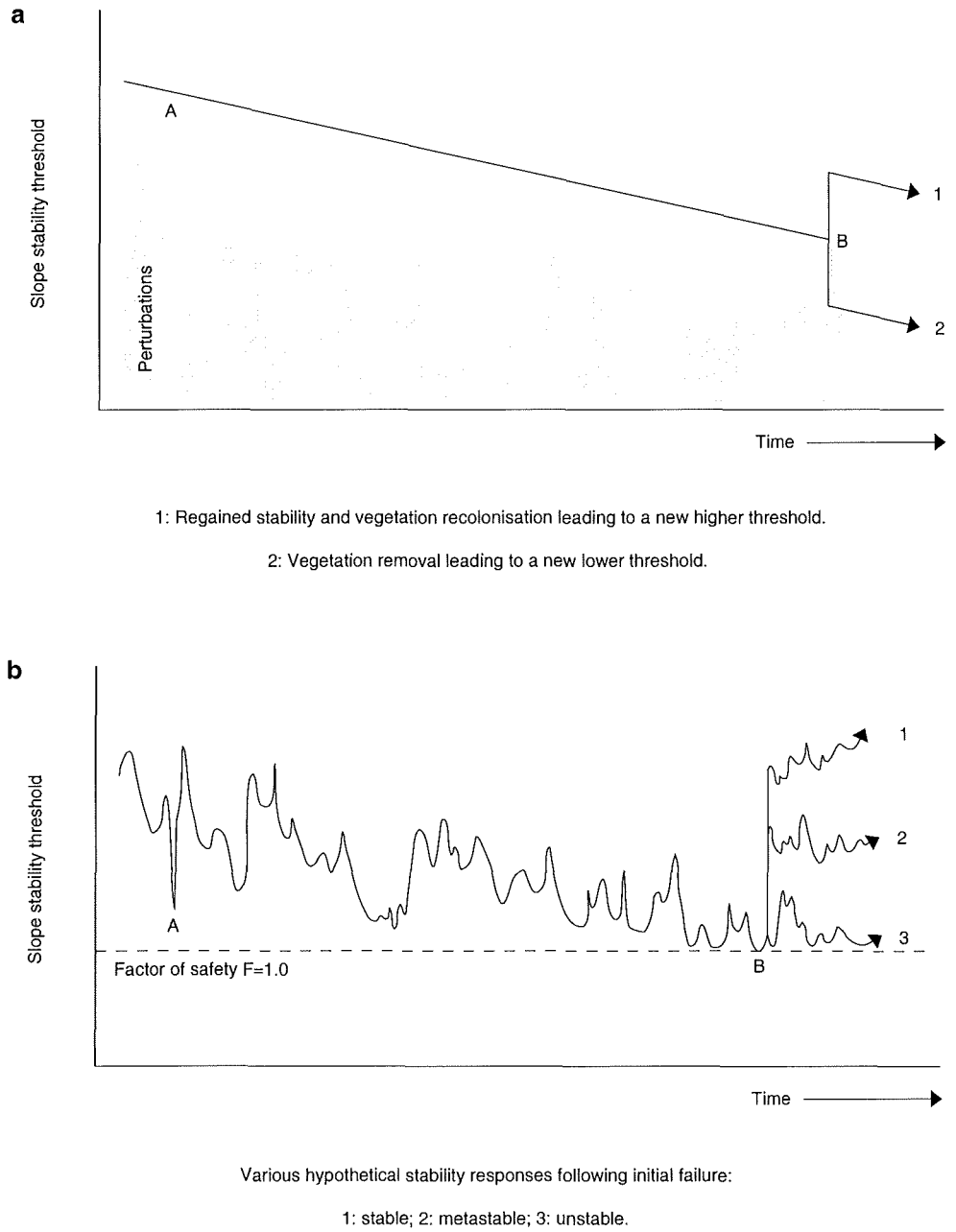


Figure 2.7. Declining slope stability threshold possibilities as the cause of delayed or renewed paraglacial modification of drift slopes.

terms of declining slope stability - an approach which may shed further light on the nature and cause of delayed or renewed slope activity (Figure 2.7). Both diagrams illustrate that the extrinsic perturbation which finally triggers failure (B) is not necessarily the largest (A) (*cf.* Jonasson and Strömquist, 1987; Nyberg,

1987). In the first case (Figure 2.7a), random perturbations are superimposed on a generally declining stability threshold. Following failure, the 'new' stability threshold could follow either a higher (1) or a lower (2) level of stability. In the second case (Figure 2.7b), as slope stability progressively diminishes, failure occurs at the point when the slope factor of safety (defined by the ratio of shearing resistance to shearing force, SR/SF) is reached. Following this, the landscape could either adopt (1) stable, (2) metastable or (3) unstable behaviour.

Similar conceptual models (Figure 2.8) can be constructed for particular aspects of paraglacial slope adjustments, such as slope form and gully sidewall gradient. A number of hypothetical responses are illustrated, as for example in the gully depth/width diagram, which reveal the possibly complex paraglacial adjustment of slope form through incision, slope decline, slope retreat and aggradation. The search for an understanding of why different valley-side slopes have responded in different ways and at different times to paraglacial reworking

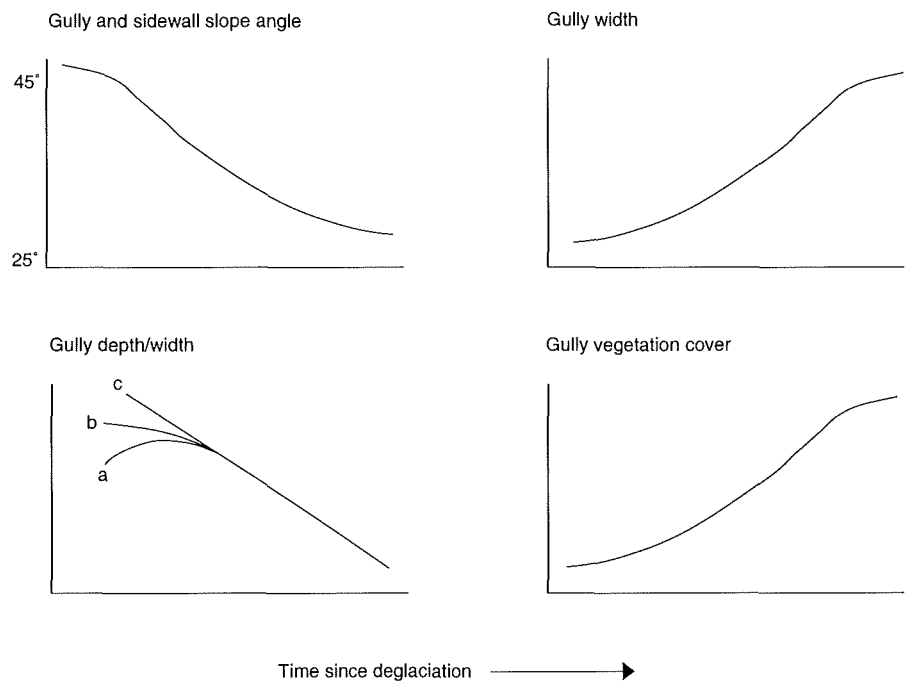


Figure 2.8. Hypothetical paraglacial modification of slopes through time.

offers a fruitful area for further research, and if a greater understanding is to be gained, such ideas require testing.

2.5 Research questions.

The concept of paraglacial reworking of glacigenic drift on deglaciated hillslopes has attracted considerable interest in recent years, but several uncertainties, untested theories and unanswered questions remain. Five such questions are summarised below.

1 How widespread is paraglacial drift slope modification?

A feature of the existing literature is the lack of a clear understanding of the spatial distribution of paraglacially-modified drift cover - whether paraglacial slope modification should be regarded as the norm, or as an over-emphasised, marginal concept. Such uncertainty stems partly from the incomplete nature of the stratigraphic record, and partly from the concentration by investigators on sites of pronounced paraglacial activity.

2 What criteria can be used to distinguish reworked drift cover from unmodified drift cover?

It is generally acknowledged that paraglacially-reworked drift is difficult to distinguish from *in situ* glacigenic deposits, at least on the grounds of clast angularity, clast shape, matrix granulometry, clast imbrication or dip, or fabric type. Further sedimentological analyses are required to establish whether other criteria may be more successful. These may also provide process and stratigraphic implications relating to paraglacial reworking.

3 What are the constraints on paraglacial reworking of drift slopes?

There has been a shortage of work regarding the limiting factors which are required for paraglacial slope modification to occur. This is important if the intrinsic causes of paraglacial slope activity are to be understood, and the intensity or lack of activity at particular sites explained.

4 What are the factors responsible for the initiation and termination of delayed or renewed paraglacial drift slope adjustment?

It is unclear as to why drift slope modification was delayed or renewed in some areas after deglaciation and not in others. Further, why did some drift slopes apparently remain initially stable, despite the action of destabilising factors, only to fail later? This question seeks not only to address the hitherto unexplained initiation of such delayed or renewed activity, but also the termination of activity and restabilisation of slopes.

5 Can we establish and validate models of paraglacial slope adjustment through time?

It is notable that most studies of paraglacial resedimentation of glacigenic drift have paid little attention to the effects of paraglacial activity in modifying slope form. The above question assesses the claims made recently by some researchers (e.g. Ballantyne and Benn, 1996) that paraglacial modification of the form and behaviour of drift slopes can be represented by a series of time-dependent, sequential models (Figure 2.5). If paraglacial hillslope adjustment does follow a common course, and general patterns are deemed to be valid, then the current level of understanding of the temporal nature of paraglacial slope adjustment will be significantly enhanced.

Chapter 3

The field sites

3.1 Rationale.

Aspects of recent and ancient paraglacial hillslope modification and resedimentation were investigated in detail at ten field sites in northwest Europe - six in southern Norway (Figure 3.1) and four in northern Scotland (Figure 3.2). The underlying rationale for conducting research in Norway and Scotland lay in the assumption that the paraglacial behaviour of slopes exposed by recent glacier retreat in Norway might offer an analogue for Late Pleistocene paraglacial resedimentation of drift in Scotland and other formerly-glaciated mountain environments (Ballantyne and Benn, 1994, 1996).

Ice covers and their related paraglacial landsystems have formed in a wide range of plate tectonic and structural settings, but the bulk of the earth's glacial record can be shown to have been deposited and preserved in basins within extensional settings (Eyles, 1993). In such basins, source area uplift and basin subsidence fulfil the tectonic preconditions for the initiation of glaciation and the accommodation and preservation of glacigenic sediments. Tectonic setting, particularly subsidence rates, also dictate the type of glacigenic facies and facies successions that are deposited. In the context of paraglaciation, it is notable that paraglacial reworking of glacial landforms and sediments is most effective in extreme terrain environments undergoing rapid uplift, owing to the abundance of steep slopes, the action of high-energy processes such as debris flows and ablation-triggered floods, and the widespread availability of unconsolidated drift and unstable bedrock (e.g. Eisbacher and Clague, 1984; Owen *et al.*, 1995). In contrast, paraglacial activity is less effective in tectonically-stable settings, or

where ice has advanced out of high-relief mountain regions to the foothills. In such settings the preservation potential of ice-marginal landforms is greater (Benn and Evans, 1998).

Within this context, it is acknowledged that the research reported from the Norwegian and Scottish field sites described in this chapter is confined to paraglacial activity on intermediate terrain in passive continental margins, where shelf deposition is controlled primarily by extrinsic processes such as climate (Mitchell and Reading, 1986). Consequently, it may be possible that the findings reported from these northwest European field sites are not fully representative of paraglaciation in other tectonic settings, such as high relief alpine fold mountain belts in New Zealand or the Karakoram-Himalaya.

With that caveat in mind, the choice of particular sites in Norway and Scotland was determined by the following parameters: (i) the presence of steep, drift-mantled hillslopes with clear morphological evidence for sediment reworking; (ii) a well-documented history of ice advance and subsequent retreat from which the pattern and chronology of paraglacial activity may be reconstructed; and (iii) the presence of deep gullies incised into the drift slope, permitting close examination of the internal architecture of both source and reworked sediments.

Relevant characteristics of the selected field sites are considered in turn in the following four sections. To simplify site description, the Norwegian sites are divided between the Jostedalsbre region and the Jotunheim massif, and the Scottish field sites are split between the Northwest Highlands and the Grampian Highlands.

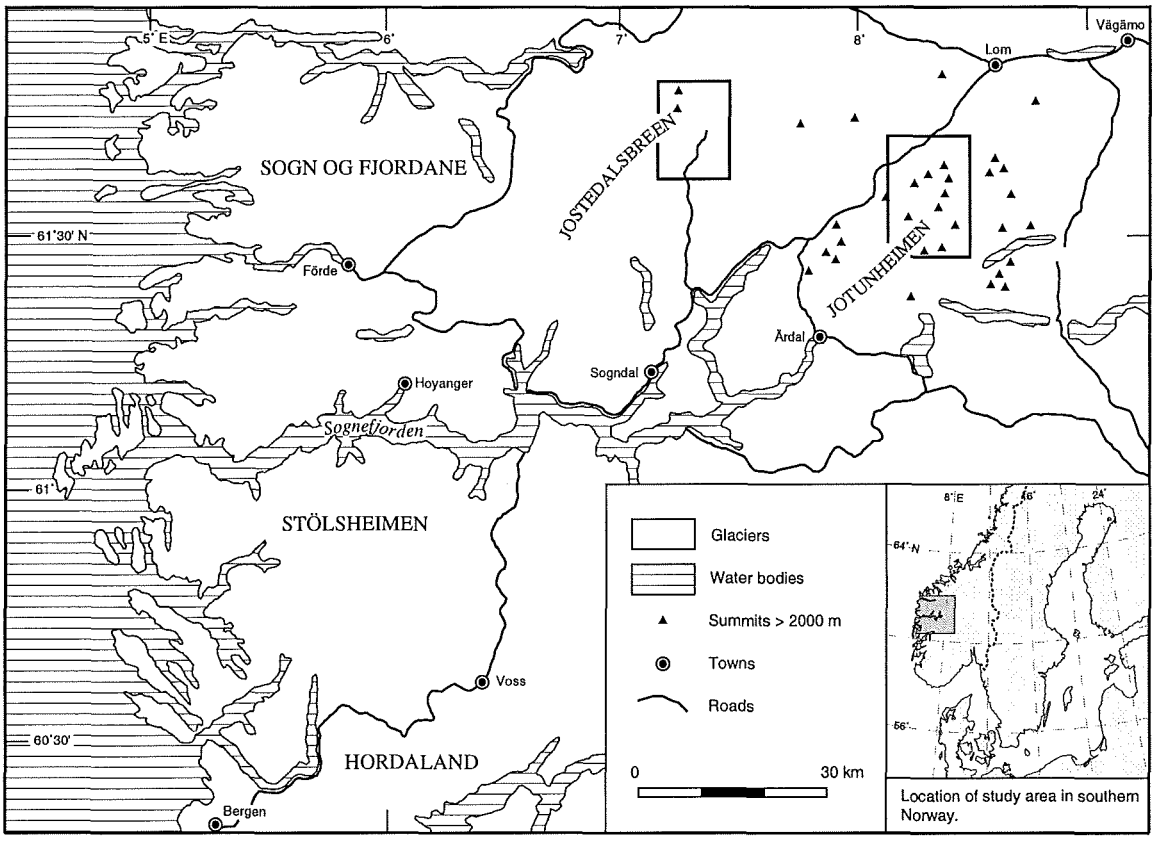


Figure 3.1. Location of the Jostedalsbre and Jotunheim field site areas in southern Norway.

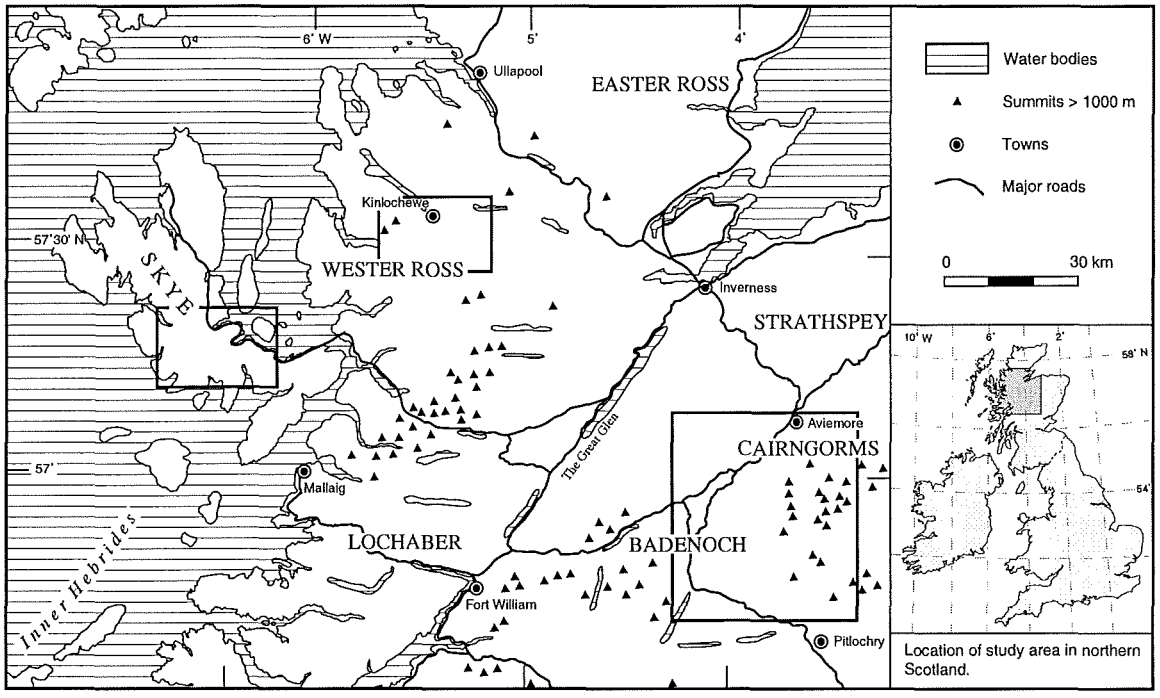


Figure 3.2. Location of the Northwest Highlands and Grampian Highlands field site areas in Scotland.

3.2 Jostedalsbreen.

Paraglacial activity was investigated in the foreland areas of four neighbouring outlet glaciers which drain the Jostedalsbre ice cap: Bergsetbreen, Fåbergstølsbreen, Lodalsbreen and Nigardsbreen.

General setting

Jostedalsbreen (61°40' N, 7°05' E) is located in the Sogn og Fjordane county of west-central Norway, approximately 170 km north-east of Bergen (Figure 3.1). The area is underlain by Precambrian basement gneisses of granitic and quartz-dioritic composition, predominantly of Caledonian age (Holtedahl, 1960; Holtedahl and Dons, 1960). In the Jostedal trunk valley to the immediate east of the ice cap, a strong structural trend runs WSW to ENE.

With an overall length of *c.* 100 km, and covering an area of 486 km^2 (Erikstad and Sollid, 1986), Jostedalsbreen is the largest ice cap in continental Europe. Much of the ice surface has an altitude in excess of 1600 m, rising in the north-east to the nunatak summit of Lodalskåpa (2083 m). The ice cap is characterised by a maritime climate and is nourished by a high annual precipitation that may exceed 2500 mm (Østrem *et al.*, 1988). Numerous outlet glaciers descend into steep-sided, deeply-incised surrounding valleys, where the glacier forelands receive an estimated 1000-2000 mm annual precipitation (Erikstad and Sollid, 1986) and experience mean annual temperatures ranging from +2°C to +4°C (Matthews, 1987). The four forelands investigated are located in tributary valleys that drain east into northern Jostedalen (Figure 3.3).

Although these four tributary valleys vary in foreland altitude (from 340 m at Nigardsbreen's present glacier terminus to 860 m at Lodalsbreen), their slopes

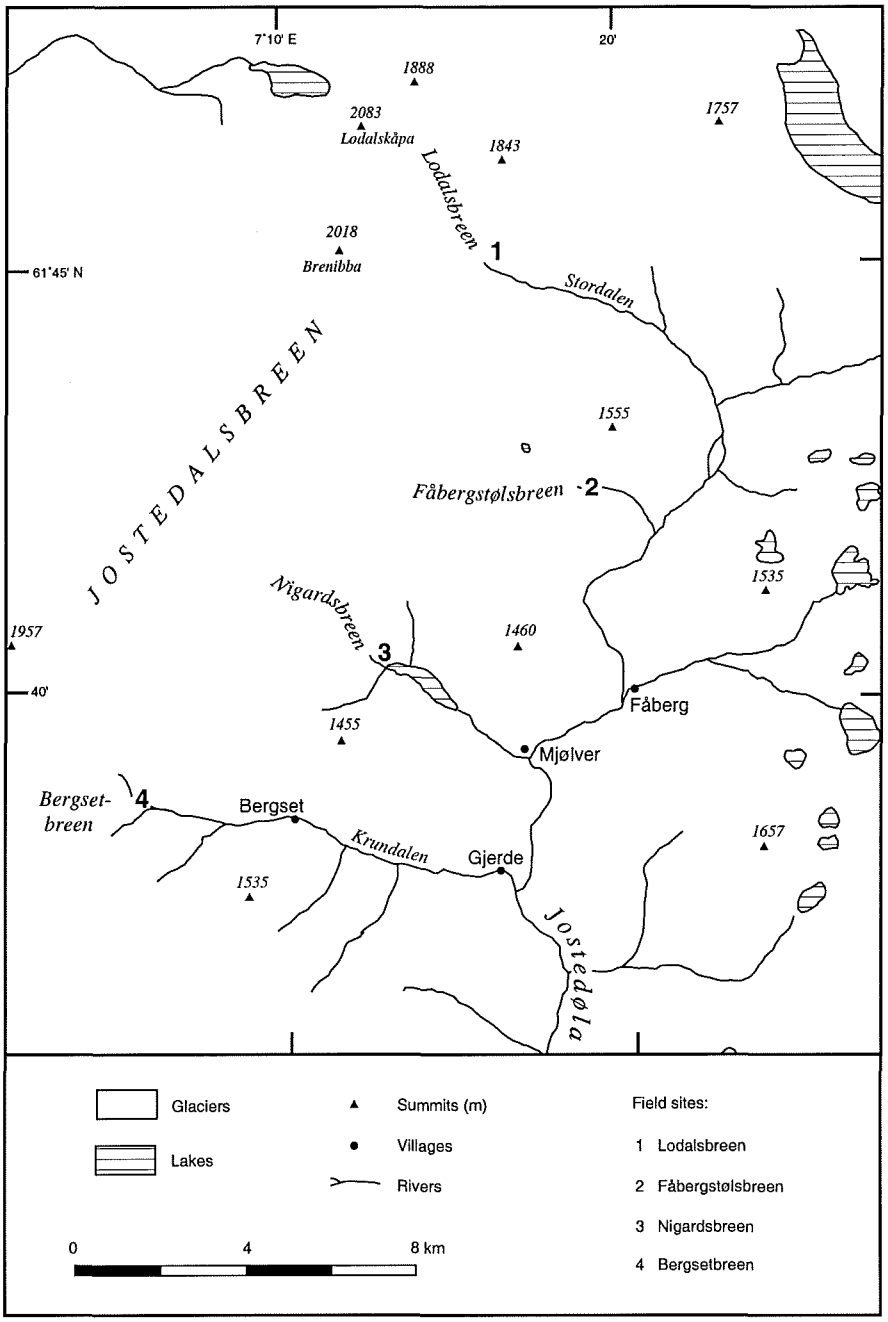


Figure 3.3. Location of the field sites in upper Jostedalen, Norway.

have a fairly similar general form characterised by three zones: (1) a very steep (50-70°) gullied rockwall, (2) a steep (<30-40°) cover of locally-gullied glacigenic drift of variable thickness, and (3) a slope-foot zone of rockfall, debris flow or snow avalanche debris overlying basal till or bedrock (Figure 3.4). Exceptions to this general form are found on the foreland of Bergsetbreen, where a wide

Figure 3.4. Fåbergstølsdalen, a recently-deglaciated valley in the vicinity of Jostedalsbreen. The thick cover of steep, valley-side glacigenic drift was exposed by retreat of Fåbergstølsbreen from its Late Neoglacial ('Little Ice Age') maximum down-valley (to the right). Widespread gullying of the drift is clearly visible.

hanging valley (Vetledalen) joins Bergsetdalen from the south, and in Fåbergstølsdalen, where the underlying structural dip imparts a pronounced valley asymmetry, so that the southern flank of the valley consists of a steep (60°) bare rock slope, whilst the northern valley side parallels the underlying structure and is underlain by rock slopes with an average gradient of 37°.

Holocene glacial history of the Jostedalsbre region

Following the Late Weichselian glacial maximum, most of the Jostedalsbre area was deglaciated during the middle of the Preboreal chronozone at *c.* 9.5 ka BP (Vorren, 1973; Andersen, 1979; Aa, 1982; Rye *et al.*, 1997). In the late Preboreal chronozone and early Boreal chronozone, however, several outlet glaciers of Jostedalsbreen advanced to positions beyond their subsequent

Late Neoglacial ('Little Ice Age') maxima (Nesje *et al.*, 1991; Nesje and Kvamme, 1991), supporting the view that deglaciation occurred shortly after *c.* 9 ka BP (Karlén and Matthews, 1992).

Temperature reconstructions for the Holocene climatic optimum suggest that ice completely disappeared from the Jostedalsbre plateau between *c.* 8 and 6 ka BP (Nesje and Kvamme, 1991; Nesje *et al.*, 1991), implying that the four field sites were also ice-free at this time. The first significant Neoglacial glacier advance in the area is thought to have occurred between 3.7 and 3.1 ka BP (Mottershead *et al.*, 1974; Mottershead and Collin, 1976; Nesje *et al.*, 1991; Nesje and Dahl, 1991). However, it appears that no previous Neoglacial advance of outlet glacier ice in the Jostedalsbre area exceeded in extent that which occurred in response to climatic cooling during the 'Little Ice Age' (Matthews, 1991; Nesje and Kvamme, 1991; Nesje *et al.*, 1991; Karlén and Matthews, 1992; Matthews and Karlén, 1992; Nesje and Dahl, 1993; Matthews *et al.*, 1996), which was especially severe during the seventeenth century AD (Lamb, 1979, 1985; Grove, 1985, 1988). 'Little Ice Age' glacier expansion in the region culminated during the mid-eighteenth century AD (Erikstad and Sollid, 1986; Bogen *et al.*, 1989; Bickerton and Matthews, 1993), approximately coincident with an increase in incidence of landslides, avalanches and floods in this area (Grove, 1972, 1988; Grove and Battagel, 1983; Innes, 1985; McCarroll, 1993, 1995; Blikra, 1994; Matthews and McCarroll, 1994; Nesje *et al.*, 1994).

Pattern and chronology of recent ice retreat at the Jostedalsbre field sites

Holocene glacier variations of the Jostedalsbre outlet glaciers have been intensively investigated, and those attributable to the most recent period of glacier retreat are of particular relevance to the study of current paraglacial activity on recently-deglaciated terrain. The pattern and chronology of the 'Little Ice Age'

advances of Bergsetbreen, Fåbergstølsbreen, Lodalsbreen and Nigardsbreen have been reconstructed by Bickerton and Matthews (1993) on the basis of lichenometric dating of recessional moraines supplemented by historical evidence and, in the case of Nigardsbreen, radiocarbon dating (Matthews *et al.*, 1986). This research indicates that these glaciers attained their 'Little Ice Age' maximum positions within the period AD 1690 to AD 1775. From these maximum limits, ice retreat from the late eighteenth century onwards was gradual, and interrupted by several readvances or still-stands, as evidenced by the numerous recessional moraine ridges deposited inside the outermost end moraine (Andersen and Sollid, 1971; Bickerton and Matthews, 1992, 1993). Intermittent retreat throughout the nineteenth century was arrested in the early twentieth century by two minor regional readvances (1903-1911 and *c.* 1921-1930; Winkler, 1996), which were in turn succeeded by several decades of uninterrupted glacier retreat (Nesje, 1989). Bergsetbreen experienced an early rapid response to twentieth century warming, retreating 408 m between 1931 and 1944, whereas at the larger glacier tongues of Fåbergstølsbreen and Nigardsbreen, the most rapid period of retreat occurred during the 1960s, when retreat rates averaged > 100 m yr^{-1} (Østrem, 1988; Winkler, 1996).

Winkler's (1996) record of glacier-front variations demonstrates that since the late 1960s the smaller outlet glaciers of Jostedalsbreen have stopped retreating, and have subsequently readvanced. Bergsetbreen, for example, has advanced strongly, covering *c.* 250 m between 1966 and 1995. In contrast, the larger glaciers were still experiencing retreat until the late 1970s, and have only started to advance in the last decade. Nigardsbreen advanced 124 m between 1988 and 1995, and Fåbergstølsbreen advanced 88 m during the period 1992-95.

3.3 Jotunheimen.

Ancient and recent paraglacial activity was also investigated in Jotunheimen, in and around the upper parts of Leirdalen and Visdalen.

General setting

The Jotunheim massif (61°30 N, 8°20' E) is located in south-central Norway, approximately 40 km inland of Jostedalsbreen (Figure 3.1). The area was upthrust during the Caledonian orogeny, and the sites under investigation are dominated by a core of ultramafic rocks and layered pyroxene-feldspar gneisses belonging to the granulite facies of metamorphism (Holtedahl and Dons, 1960; Battey and McRitchie, 1973, 1975). Marked foliation (Battey, 1965) and south-dipping lag faults (Battey and McRitchie, 1973) further characterise these localities.

Jotunheimen is the highest mountain range in northern Europe, containing numerous peaks that rise from valley floor elevations of *c.* 1000 m to summits over 2000 m, the highest being Galdhøpiggen (2469 m). The massif includes some 300 glaciers, which range from small ice caps to valley and corrie glaciers (Matthews, 1987), and permafrost has been detected above *c.* 1600 m (Østrem, 1964, 1965; King, 1986; Ødegard *et al.*, 1992; King and Åkerman, 1993). Mean annual temperature in upper Leirdalen and Visdalen, estimated from data for neighbouring meteorological stations, is *c.* -1° C (Green and Harding, 1980; Matthews, 1987; Ødegard *et al.*, 1992; Førland, 1993; Matthews *et al.*, 1997). Mean annual precipitation lies between 1000 and 1500 mm (Erikstad and Sollid, 1986; Førland, 1993).

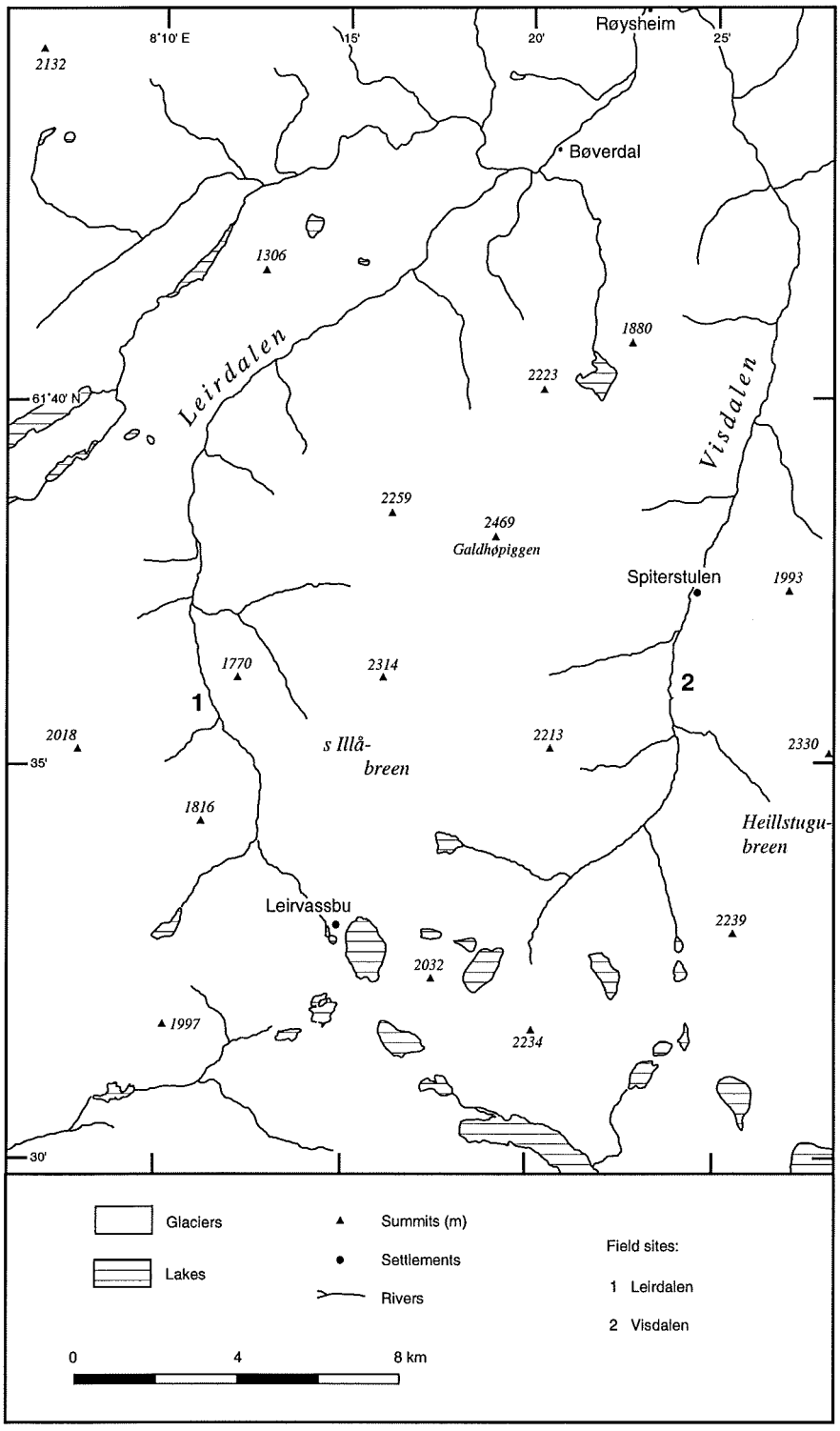


Figure 3.5. Location of the field sites in central Jotunheimen, Norway.

Leirdalen and Visdalen (Figure 3.5) are trunk valleys which dissect the central Jotunheim massif and possess upper valley slopes characterised by: (1) an

upper steep (50-70°) gullied bedrock slope, (2) a thin mantle of primarily rockfall debris overlying bedrock, and (3) a lower slope zone of locally-incised vegetated glacigenic drift of variable depth. Both upper Leirdalen and upper Visdalen are joined by hanging valleys whose uppermost reaches are occupied by glacier ice, with glacier forelands at *c.* 1550 m elevation.

Holocene glacial history of Jotunheimen

Following ice sheet glaciation of the Jotunheim area during the Late Weichselian, widespread deglaciation took place at *c.* 9 ka BP in the Preboreal chronozone (Vorren, 1973; Andersen, 1980; Nesje and Rye, 1990; Shakesby *et al.*, 1990; Nesje and Dahl, 1993). As in the vicinity of Jostedalsbreen, the results of detailed mapping in west Jotunheimen suggest that a readvance occurred during the final phase of Late Preboreal ice sheet wastage (Shakesby *et al.*, 1990).

An apparent absence of glacier ice on the mountain plateau west of Jotunheimen during the Atlantic chronozone (Nesje *et al.*, 1991) implies that many, if not all of the Jotunheim glaciers disappeared at this time. Climatic conditions deteriorated after the late Atlantic chronozone, but there is no evidence for a Neoglacial expansion episode in Jotunheimen prior to those which occurred at a small number of glaciers at *c.* 2.7 and *c.* 1.3 ka BP (Griffey and Matthews, 1978; Matthews and Shakesby, 1984). Although a small number of pre-'Little Ice Age' Neoglacial moraines have been identified outside the limits of the subsequent 'Little Ice Age' advance at small, high-altitude glaciers in eastern Jotunheimen (Østrem, 1965; McCarroll, 1991), the overall pattern of Neoglacial ice expansion in Jotunheimen is one of maximum ice extent during the 'Little Ice Age' (Griffey and Matthews, 1978; Matthews and Shakesby, 1984; Matthews, 1987; Nesje and Rye, 1990; Shakesby *et al.*, 1990; Matthews, 1991; Karlén and Matthews, 1992) particularly during the 18th Century AD (Liestøl, 1967;

Matthews, 1974, 1975, 1977a, 1977b; Grove, 1988; McCarroll, 1989). This period of climatic deterioration was also associated with enhanced rapid mass-movement on hillslopes in Leirdalen (Innes, 1985).

Pattern and chronology of recent ice retreat in Jotunheimen

Many of the glaciers in Jotunheimen probably retreated from their 'Little Ice Age' limits shortly after AD 1750 (Erikstad and Sollid, 1986). Since many glacier forelands in the Jotunheim are well-endowed with recessional moraines, it is inferred that gradual retreat was interrupted by several readvances or still-stands. Detailed lichenometric dating studies at Storbreen provide evidence in support of this contention, recording alternating phases of glacier retreat and readvance or stillstand throughout the nineteenth century (Matthews, 1977b).

Reflecting Jotunheimen's more continental climate, the Jotunheim and Jostedalsbre glacier-fronts have fluctuated asynchronously during the twentieth century (Winkler, 1996). Throughout this period, overall retreat in the Jotunheim was interrupted by only one regionally-significant readvance, dated at *c.* AD 1920-1930 (Winkler, 1996; McCarroll, personal communication, 1997). Moreover, there is no correlation of glacier-front oscillations between Jotunheimen and Jostedalsbreen during the first three decades of the twentieth century (Winkler, 1996). The period of general glacier retreat that occurred in the Jostedalsbre region during the middle of this century was, however, also recorded in Jotunheimen, though the annual retreat rates were generally slower than at Jostedalsbreen (Matthews *et al.*, 1995). Mass balance studies at a number of glaciers (Winkler, 1996) have demonstrated an overall ice mass decrease over recent decades, with positive balances recorded only since the early 1990s. With the exception of some glacier fronts in the west of the region that have exhibited a

more or less stable state during the last few years, there is no current glacier advance in Jotunheimen (Winkler, 1996).

3.4 The Northwest Highlands of Scotland.

Resedimentation of steep, valley-side drift was studied at two sites in the Northwest Highlands of Scotland: the Western Red Hills, on the Isle of Skye, and Glen Docherty, in Wester Ross.

General setting

The Northwest Highlands are here considered to encompass the mainland of Scotland north of the Great Glen, and the Hebrides (Figure 3.2). The Western Red Hills in central Skye (57°17' N, 6°8' W) form a compact area of *c.* 30 km^2, flanked in the west by Glen Sligachan and penetrated to the north and east by fjords (Lochs Sligachan and Ainort respectively; Figure 3.6). Their rounded summits reach heights exceeding 700 m and represent the eroded remnants of a Palaeogene plutonic centre, part of the British Tertiary Igneous Province (Harker, 1904; Richey, 1932; Emeleus, 1983; Bell and Harris, 1986). The field site is dominated by annular acid intrusions of granite, granophyre and felsite, though basalt lavas are also preserved on the western summit of Glamaig (Bell and Harris, 1986).

Glen Docherty, on the northwest mainland (57°35' N, 5°15' W) crosses the Moine Thrust on its 7 km descent from the watershed at 245 m to Loch Maree near Kinlochewe (Figure 3.6). Adjacent summits nowhere exceed 539 m (Carn a' Ghlinne). Although Glen Docherty parallels the Loch Maree Fault, the strike of underlying strata have not been affected (Dixon, 1886). The principal lithologies at this site are Moine metasedimentary strata, predominantly mica-schists and

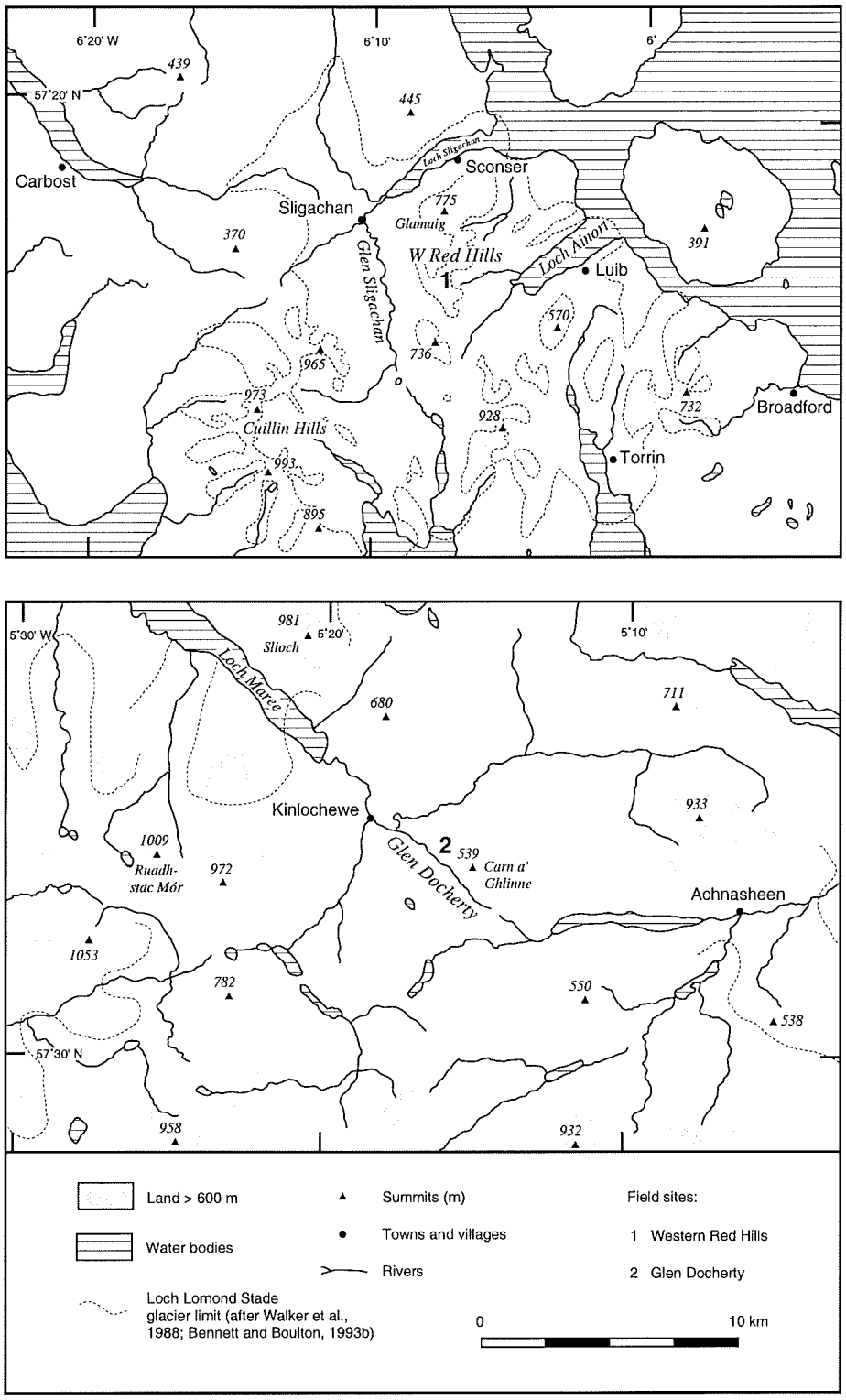


Figure 3.6. Location of the field sites in the Northwest Highlands, Scotland.

gneisses, plus less extensive shales (Harris and Johnson, 1991). Lewisian and Torridonian rocks are found only in the lowermost part of Glen Docherty by Loch Maree (Johnstone and Mykura, 1989).

Both field sites experience mean annual temperatures of *c.* 8.5°C and mean annual precipitation of 2000 - 2500 mm (Birks, 1973; Meteorological Office, 1977). The slopes of the Western Red Hills exhibit an upper slope form characterised by a veneer of frost-shattered diamictic regolith interrupted only locally by bedrock. Lower slopes are gullied, vegetated and predominantly drift-mantled, though a shallow cover of 'granitic scree' (Ballantyne, 1991c) frequently extends downslope onto the glacigenic sediment. At Glen Docherty both valley-sides support locally-gullied vegetated glacigenic drift, though bedrock appears extensively at the slope crest of the south-facing valley side. Both sites are largely treeless and dominated by dwarf-shrub communities. Peat formation is widespread on poorly-drained ground (Birks, 1972, 1973).

Glacial history of the Northwest Highlands field sites

Though the Northwest Highlands of Scotland were repeatedly glaciated during the Pleistocene, the only glacial events for which there is considerable evidence are movement of the last ice sheet across the area during the Dimlington Stade of *c.* 26-13 ka BP, and the brief regeneration of local valley and corrie glaciers during the Loch Lomond Stade of *c.* 11-10 ka BP (Gray and Coxon, 1991) prior to rapid climatic amelioration (Atkinson *et al.*, 1987; Dansgaard *et al.*, 1989). Both field sites were submerged beneath the Devensian ice sheet, whose upper limit is marked in the region by high-level periglacial trimlines (Ballantyne *et al.*, 1987; Dahl *et al.*, 1996; Ballantyne *et al.*, 1997), and they also lie within the limits of the Loch Lomond Stade readvance (Ballantyne, 1989; Bennett and Boulton, 1993a; Bennett, 1994).

Attempts to define the limits of local glacier advances in central Skye have produced widely differing views regarding their number, timing and extent (Charlesworth, 1956; Anderson and Dunham, 1966; Birks, 1973; Sissons, 1977a; Walther, 1984). Ballantyne (1989) proposed that the Western Red Hills were last glaciated during the Loch Lomond Stade by outlet glaciers draining a major ice field centred over the Cuillin Hills. Such local glacier ice deposited thick drift deposits on slopes below 300 m - 400 m within the field site (Ballantyne and Benn, 1991b). Benn *et al.* (1992) concluded that final deglaciation of Skye proceeded in two stages: the first reflecting a decline in precipitation and marked by numerous glacier stillstands and readvances; the second characterised by uninterrupted retreat and local glacier stagnation in response to a rapid increase in temperatures at the end of the stade (Benn *et al.*, 1992).

Glacial landforms and deposits in Wester Ross were first described by Peach *et al.* (1912, 1913a, 1913b). During the last glacial maximum the area was covered by an ice sheet 700 - 900 m thick (Ballantyne *et al.*, 1997). Ice sheet retreat was briefly interrupted by the Wester Ross Readvance (Robinson and Ballantyne, 1979; Sissons and Dawson, 1981), which is thought to have occurred at *c.* 13.5 ka BP (Ballantyne *et al.*, 1987). Sissons's (1977b, 1982) reconstruction of Loch Lomond Stade glaciers in this area was revised by Ballantyne (1986b) and Bennet and Boulton (1996) to show a more extensive former ice field, and several outlet glaciers. This included an ice divide at the head of Glen Docherty, from where ice flowed west to Loch Maree, depositing thick valley-side drift in Glen Docherty, and east to Achnasheen (Benn, 1989, 1992, 1996). The Loch Lomond Stade glacier limits in lower Glen Docherty are marked by recessional moraines ridges (Bennett and Boulton, 1993b), which suggests that ice initially underwent active retreat, accompanied by brief stillstands or readvances. An absence of recessional moraines further up Glen Docherty, however, suggests that subsequent ice retreat may have been uninterrupted.

Postglacial vegetation history of the Northwest Highlands field sites

The Holocene vegetation record for the Red Hills area on Skye has been reconstructed from lake sediment cores and peat profiles obtained at ten local sites (Figure 3.7): Loch Ashik (Walther, 1984; Walker *et al.*, 1988; Walker and Lowe, 1990), Loch Cill Chriosd (Birks, 1973; Birks and Williams, 1983), Loch Meodal (Birks, 1973; Birks and Williams, 1983), Elgol (Walker and Lowe, 1990), Glen Varragill (Walther, 1984; Walker *et al.*, 1988; Benn *et al.*, 1992), Druim Loch (Walker and Lowe, 1990), Sligachan, Marsco, Luib and Clach Oscar (Walker *et al.*, 1988; Benn *et al.*, 1992). Many of these sites share a common sequence of early to mid-Holocene vegetation changes, typified by successive maxima in Poaceae/*Rumex, Empetrum nigrum, Juniperus, Betula* undifferentiated, *Corylus avellana*-type and *Alnus glutinosa* (Lowe and Walker, 1991; Benn *et al.*, 1992).

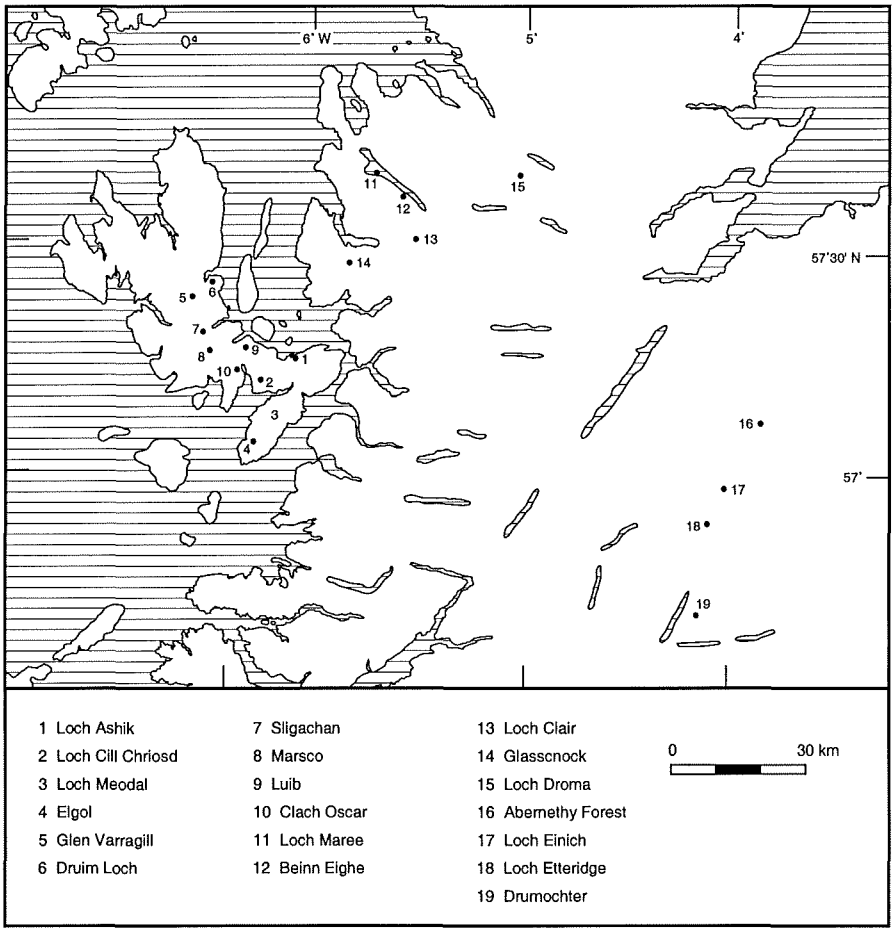


Figure 3.7. Location of the Scottish pollen sites mentioned in the text.

Woodland communities thrived until *c.* 5 ka BP when they were largely replaced by grassland and heath. Nevertheless, despite a long history of human occupation on the east coast of Skye, the southern parts of the island appear to have retained large swathes of woodland until *c.* 300 yr BP when widespread clearance accompanied the introduction of cattle grazing (Birks and Williams, 1983). A distinctive feature of the pollen stratigraphy at Loch Ashik is the lack of evidence for any significant anthropogenic influence on vegetation at *c.* 5 ka BP, and also for a short but emphatic pine phase recorded at *c.* 4 ka BP (Birks and Williams *et al.*, 1983).

The Holocene vegetation history of Glen Docherty has been reconstructed with reference to pollen records obtained from five nearby sites (Figure 3.7): Beinn Eighe (Durno and McVean, 1959), Loch Droma (Kirk and Godwin, 1963), Loch Maree (Birks, 1972), Loch Clair (Pennington, *et al.*, 1972) and Glassnock (Robinson, 1977). Early Holocene plant communities in the vicinity reflected open dwarf shrub heath with significant frequencies of Poaceae, *Rumex, Lycopodium selago, Salix* and *Empetrum.* Shortly after this early phase *Empetrum* declined and *Juniperus* expanded rapidly, though this was largely superseded by both *Betula* and *Corylus* before *c.* 9 ka BP. At Loch Maree, this *Betula* community was associated with the presence of *Quercus* and *Ulmus.* Around 8 ka BP it was replaced by *Pinus* woodland, which dominated until *c.* 4.2 ka BP (Birks, 1972). At nearby Loch Clair, however, *Pinus* increased at *c.* 6.5 ka BP, and was replaced by Poaceae and *Calluna* at *c.* 2.9 ka BP (Pennington *et al.*, 1972). This diachroneity in forest history may reflect former changes in catchment hydrology, or the influence of fire (Durno and McVean, 1959; Birks, 1972), either natural in origin and related to former dry periods (Tipping, 1996), or associated with forest clearances (Ballantyne *et al.*, 1987) prior to the subsequent rise of heather moorland (Birks, 1972; Pennington *et al.*, 1972).

3.5 The Grampian Highlands of Scotland.

Resedimentation of steep hillslope drift was examined at two sites in the Grampian Highlands of Scotland: Glen Einich, in the Cairngorm massif, and the Pass of Drumochter, in Badenoch.

General setting

The Grampian Highlands (Figure 3.2) are bounded by the Great Glen and the Highland Boundary (Read, 1935). Within this wider area, the Cairngorm Mountains comprise the largest area of high plateau in Britain, and are incised in the west by Glen Einich (57°05' N, 3°48' W), a steep-sided glacial trough, *c.* 12 km south of Aviemore (Figure 3.8). Loch Einich occupies the upper glen, situated between the dramatic crags of Braeriach (1296 m) and Sgòr Gaoith (1118 m). The site is underlain by Cairngorm Main Granite (Harrison, 1986), except at the head of the glen where Dalradian schists crop out (Read, 1935; Johnson, 1991). Structurally, the course of Glen Einich is defined by one of a series of NE-SW fractures parallel to the Caledonian trend (Harrison, 1986; Hall, 1996).

The Pass of Drumochter is a glacial breach (Linton, 1949) that dissects the Grampian Highlands adjacent to the Gaick plateau, 8 km south of Dalwhinnie (56°51' N, 4°15' W; Figure 3.8). At its summit the Pass reaches 450 m elevation, and adjacent tops attain heights of up to 803 m (Meall a Dobharchain). Solid geology comprises Dalradian psammites and semipelites (Barrow *et al.*, 1913; Read, 1935; Johnson, 1991; Stephenson and Gould, 1995) which are here disposed in a large-scale isoclinal fold termed the Drumochter Dome (Thomas, 1979, 1980).

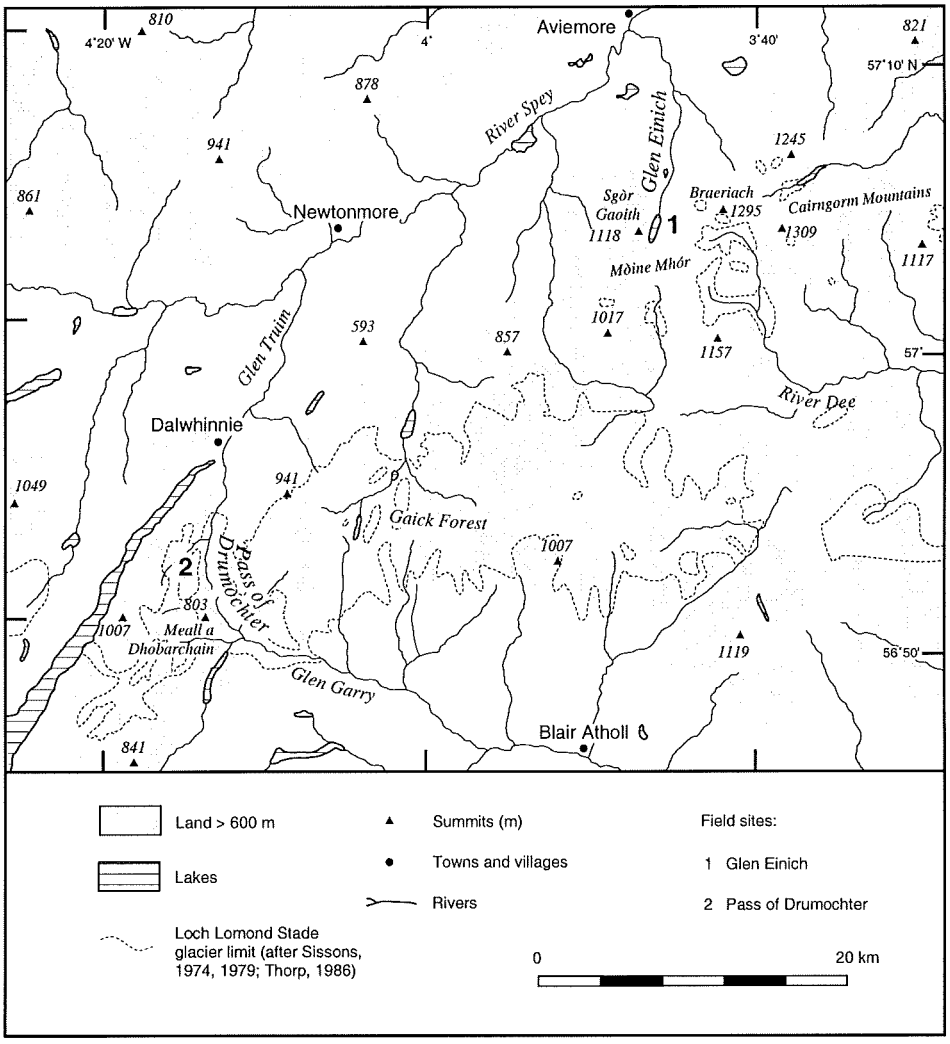


Figure 3.8. Location of the field sites in the Grampian Highlands, Scotland.

Mean annual temperature at both Grampian sites is *c.* 6° and average annual precipitation lies between 1500 and 1750 mm (Green, 1974; Meteorological Office, 1977). In upper Glen Einich the valley-wall rock slopes are flanked by a lower apron of steep vegetated glacigenic drift which is gullied and locally covered with a veneer of relict talus (Figure 3.9). The eastern flank of the glen supports a number of corries. In contrast, the lower slopes of the rounded Drumochter Hills possess a vegetated mantle of thick glacigenic drift which buries underlying bedrock and is locally-gullied. Modern vegetation in Glen Einich and Drumochter consists of open peatland and dwarf shrub communities at

Figure 3.9. Reworked deglaciated drift slopes in the upper Glen Einich basin, Grampian Highlands. The lower valley-side cover of drift was deposited by Late Devensian ice and is currently undergoing widespread modification and resedimentation.

lower level, with tundra-like taxa at higher altitudes.

Glacial history

Multiple Pleistocene glaciations have had a dramatic impact on the Grampian Highlands. The landscape character of the Cairngorms reflects selective linear glacial erosion of a rolling preglacial plateau (Sugden, 1968; Brazier *et al.*, 1996a). Such selective erosion may be attributed to the former presence of a thin cold-based ice cap over the high plateau with erosive warm-based ice streams occupying glacial troughs and breaches. In contrast to early opinion (Linton, 1949, 1955; Galloway, 1958), it is now considered that at its maximum extent the Late Devensian ice sheet was sufficiently thick to submerge the main Cairngorm massif including Glen Einich (Sugden, 1970; Glasser, 1995; Brazier *et al.*, 1996b). Models of local deglaciation range from those advocating

active retreat of glaciers upvalley (Hinxman, 1896; Jamieson, 1908; Barrow *et al.*, 1912, 1913; Hinxman and Anderson, 1915; Bremmer, 1929), to rapid, large-scale stagnation of the ice sheet and *in situ* downwastage (Sugden, 1970). More recently, on the basis of remapping of landform assemblages (including ice-marginal features in lower Glen Einich), Brazier *et al.* (1996b) have proposed five distinct periods of ice sheet decay in the northwest Cairngorms, incorporating the emergence of summits above a thinning ice sheet, active downwasting of invasive ice from Strathspey and the Dee valley, localised readvances within retreat phases, and the late development of a local ice cap with outlet glaciers centred on the Mòine Mhór. A radiocarbon date from Loch Etteridge in the nearby Spey valley indicates widespread deglaciation by *c.* 13.1 ka BP (Sissons and Walker, 1974), and it appears that the whole valley was ice free by *c.* 11.2 cal ka BP (K.D. Bennett, 1996b). Although opinion differs over the extent of Loch Lomond Readvance in the Cairngorms (Sugden, 1970; Sugden and Clapperton, 1975; Sissons, 1979; Bennett and Glasser, 1991), it is agreed that Glen Einich remained unglaciated during this readvance (M.R. Bennett, 1996) and was therefore subject to severe periglacial conditions at this time.

Prior to Sissons' (1974) contribution, the only work published on the glaciation of the Drumochter area was that by Barrow *et al.* (1913), who deemed local glacier ice to be important in the Gaick whilst surrounded by a thinning ice sheet, but specified no limits. Later, Sissons (1974) suggested renewed local glaciation of the area during the Loch Lomond Stade, following complete decay of the last ice sheet at *c.* 13 ka BP (Sissons and Walker, 1974). This view was established by reconstruction of a local ice cap over the Gaick plateau with associated outlet glaciers, one of which occupied the Pass of Drumochter. Support for this interpretation includes excellent 'hummocky' recessional moraine sequences that extend for 14 km from upper Glen Garry through the Pass of Drumochter into Glen Truim. Abrupt limits of hummocky terrain in Glen Truim

and Glen Garry were assumed to mark the termini of former outlet glaciers, particularly in Glen Garry where they are succeeded down-valley by outwash. Sissons (1974) suggested that Loch Lomond Stade ice was only briefly at its maximal extent and thickness, exhibiting a decline in the altitude of the ice surface between the summit of the Pass and termini in Glen Truim and Glen Garry from *c.* 650 m to *c.* 480 m. Active deglaciation is indicated by the presence of numerous recessional moraines.

Postglacial vegetation history of the Grampian Highlands field sites

Lateglacial and Holocene environmental changes in the Cairngorms area have been reviewed by Gordon (1993) and K.D. Bennett (1996a, 1996b). A radiocarbon-dated Lateglacial sequence from Abernethy Forest (Birks and Mathewes, 1978; Figure 3.7) shows that the earliest pioneer communities at *c.* 14-13 cal ka BP consisted of sedges and grasses growing on the recently deglaciated substrate. By *c.* 13.5 cal ka BP, the vegetation cover resembled an arctic shrub-tundra with *Betula nana* and *Empetrum* dominant. At *c.* 13.1 cal ka BP *Juniperus* and *Betula pubescens* appear to have increased, indicating colonisation by taller shrubs and trees. However, between *c.* 13 cal ka BP and *c.* 11 cal ka BP shrubs declined and *Artemisia* - type herbaceous cover flourished with some arctic-alpine species. Birks and Matthewes (1978) attributed this change to increased aridity in the eastern Grampians relative to the western Highlands during the Loch Lomond Stade.

The Holocene vegetation record from a peat sequence at the head of Loch Einich (Birks, 1975) begins with birch pollen dominant, but replaced by pine by *c.* 6.7 cal ka BP. Pine remains the dominant pollen type for the rest of the sequence, although it decreases towards the top of the sequence as pollen of peatland plants (especially *Calluna* and Cyperaceae) increase. A general decrease in tree cover

(especially pine) after *c.* 4 cal ka BP, and a corresponding increase in abundance of *Calluna* and herb taxa has also been observed in pollen records from other local sites (Birks, 1970; O'Sullivan 1974a, 1975, 1976, 1977; Birks and Mathewes, 1978; Preece *et al.*, 1984; Rapson, 1985). This shift has been attributed to ground moisture conditions unfavourable for pine regeneration (Dubois and Ferguson, 1985, 1988; Pears, 1988), forest clearances and grazing of domestic stock (Pears, 1968; O'Sullivan, 1974a, 1974b, 1975; Birks, 1975), although the causes, role and timing of forest fire in the Cairngorms, as elsewhere in Scotland, are poorly understood (K.D. Bennett, 1996a, 1996b; Tipping, 1996).

Whilst the Cairngorm pollen sites are of regional importance for reconstructing the Holocene vegetation history of Drumochter, the record obtained from Walker's (1975a, 1975b) site at the col of the Pass of Drumochter is of particular value, and broadly accords with results from nearby locations (Sissons and Walker, 1974; Walker, 1975b; Macpherson, 1980). Three pollen assemblage zones were identified in the Drumochter profile, which spans the Holocene. The lowermost zone represents birch woodland interspersed with patches of grassy hazel scrub and occasional stands of pine, oak and elm. *Juniperus* and *Empetrum* appear to have been restricted largely to upper slopes during this time. Although *Betula* remains the dominant tree pollen in the middle zone of the pollen spectrum, its dominance is subsequently terminated by the immigration of *Corylus avellana*-type. The establishment of pine-birch forest below *c.* 800 m characterises the upper zone, which also records a steady rise in *Alnus* frequencies, suggesting a gradual increase in moisture induced by higher precipitation levels. An overall decline in tree pollen concentrations in the uppermost part of this zone is accompanied by expansion of dwarf shrub heath and peat, a transition which Walker (1975a) attributed tentatively to the first stage of (Neolithic) human interference. Given its isolation and relatively severe climate, however, it is unlikely that the Drumochter area ever supported a dense

human population. Nevertheless, in the Grampians to the southeast of Drumochter, Huntley (1981) has shown that grazing pressures of the last 200 years were particularly severe in reducing the extent of woodland at mid-altitude sites.

Chapter 4

Paraglacial modification of drift slopes in Norway

4.1 Introduction.

This chapter focuses attention on the extent, nature, constraints and timing of ancient and recent or current paraglacial resedimentation of steep glacigenic drift at the six Norwegian field sites, and is organised into four main sections. Following an outline of field and analytical methods (section 4.2), the chapter assesses the extent (4.3), processes (4.4), constraints (4.5) and timing and duration (4.6) of paraglacial modification of drift. Principal results are summarised in the concluding section (4.7).

4.2 Methods.

Investigation of the characteristics of paraglacial hillslope modification at the Jostedalsbre and Jotunheim field sites involved (1) geomorphological mapping of the distribution and morphology of drift slopes, (2) instrumental survey of slope form, (3) measurement of the dimensions of gullies incised into hillslope drift, and (4) analysis of the sedimentological characteristics of *in situ* glacigenic drift deposits. For field mapping purposes 1:5,000 scale base maps were produced from Økonomisk Kartverk 1:20,000 and Statens Kartverk 1:50,000 sheets. Consultation of ground and aerial photographs of the forelands of Bergsetbreen, Fåbergstølsbreen, Lodalsbreen, Nigardsbreen and part of Leirdalen aided accurate mapping of gullies and other features. A key to all geomorphological maps is provided in Figure 4.1.

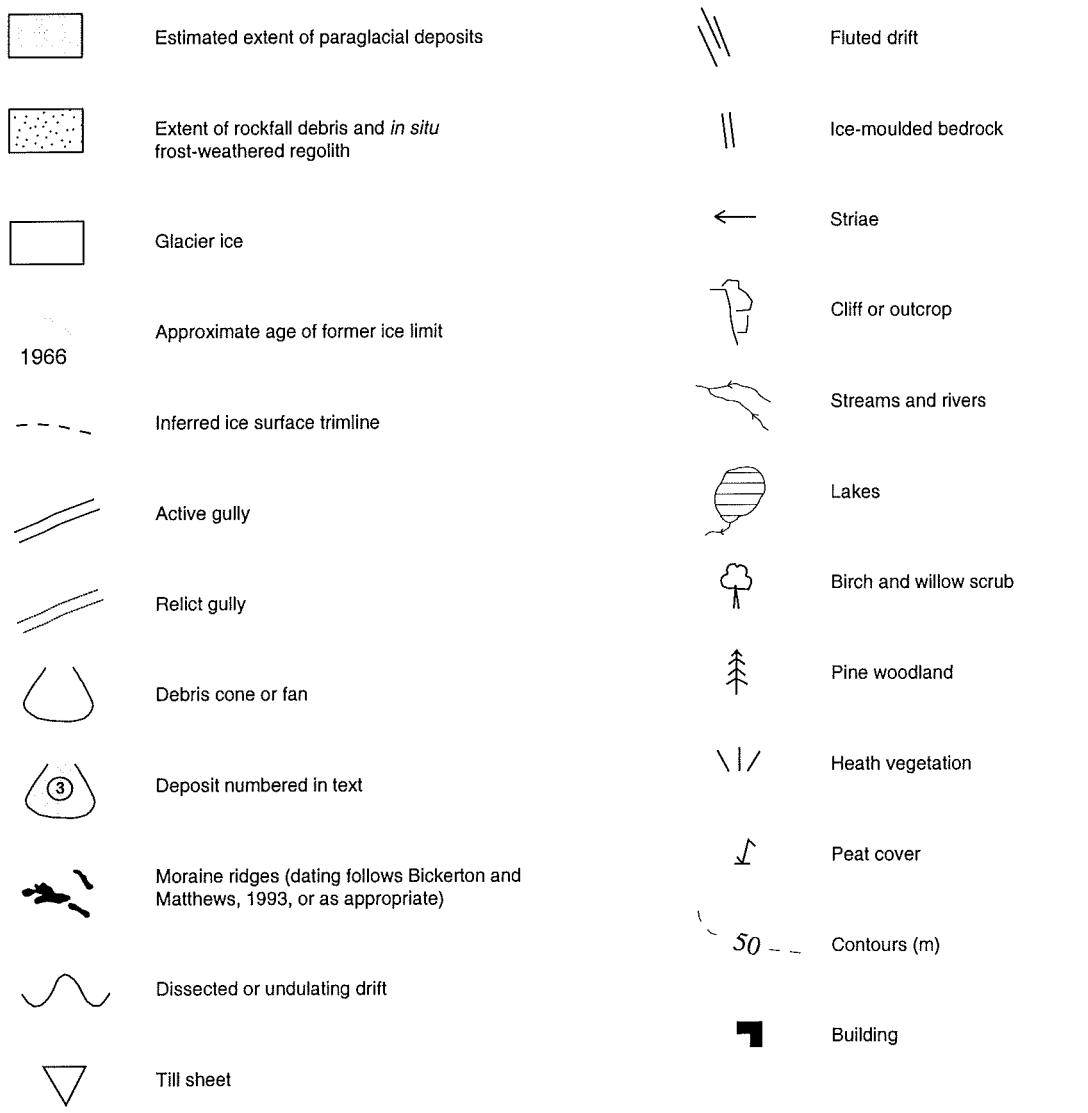


Figure 4.1. Key to geomorphological maps.

A number of terms used below require definition. 'Gully density' provides a surrogate measure of the degree of drift modification, and is defined as the total number of relict and active gullies incising recently-deglaciated ('Little Ice Age') valley-side drift within 1 km of the present glacier terminus (or the uppermost gully where gullied drift extends upvalley beyond the current ice margin). For sites not occupied by glacier ice or well outside the 'Little Ice Age' glacial limits (i.e. Leirdalen and Visdalen), gully density was calculated as a mean frequency per kilometre for the entire field site. 'Percentage reworked drift' describes the volume of drift that has been reworked and redeposited downslope as a percentage of the total volume of the valley-side drift accumulation. Drift area was calculated by multiplying drift slope width (downvalley) by slope length (the distance from the slope foot to the base of the headwall), and drift volume estimated by multiplying drift area by mean drift thickness (assuming a regular decline in the underlying ground surface). A measure of the volume of reworked drift was obtained by multiplying the number of gullies in a given area by the mean gully volume (see below). This procedure may slightly underestimate drift volume if a hollow underlies the drift surface, or (more likely) slightly overestimate the volume of reworked drift if such drift overlies buried moraines. Whilst this measure of percentage reworked drift is fairly crude, consistency of calculation allows general comparisons to be made between different field sites.

At Fåbergstølsbreen and in Leirdalen, slope profile measurements and other survey data were obtained using a Wild T1000 EDM (Electronic Digital Mapping) unit. Because of the relative inaccessibility of the Lodalsbre and Visdalen field sites, slopes at these localities were surveyed using an abney level, ranging rods and a 30 m tape. Repeatability tests suggest that the abney level readings are accurate to within ± 0.5° (Young, 1974). To reduce problems of subjectivity, both electronic and manual survey readings were taken at ten pace intervals along each survey transect, although extra measurements were made at

distinct breaks of slope. Gully dimensions were also surveyed to determine drift thickness and the volume of remobilised sediment. Gully cross-sectional area was calculated at several points along surveyed gully axes, and mean cross-sectional area was multiplied by the planimetric area of the surveyed gully to obtain an approximation of the volume of sediment removed. From these data, mean gully volume was calculated for some sites, expressed within 95% confidence limits (Tables 4.1 and 4.2). Finally, to permit assessment of the possible role of sediment characteristics on the distribution of paraglacial slope reworking, bulk (*c.* 2 kg) samples of fine-grained (< 2 mm) sediment were removed from valley-side drift deposits for laboratory granulometric analyses. These samples were reduced to *c.* 500 g for analysis and dried at 105° C for 3 days. Sediment coarser than 710 μm was dry-sieved, and the grain-size distribution of sediment finer than 710 μm analysed on a Coulter LS100 laser granulometer. Intact samples of matrix material were also removed in steel cylinders, sealed, and were later dried, weighed, saturated and re-weighed to calculate void ratio as a measure of sediment packing (Attewell and Farmer, 1976).

4.3 Extent of paraglacial modification of drift.

Although Owen *et al.* (1995) estimate that 2-31% of the total area of all landforms mapped in three Himalayan valleys comprise paraglacial fans, a dominant feature of the existing literature on paraglacial reworking of hillslope drift is the lack of a clear understanding of the extent of paraglacial redistribution of valley-side sediment. Consequently, it remains uncertain whether paraglacial modification of deglaciated drift slopes should be regarded as a 'normal' or exceptional attribute of formerly-glacierized upland environments.

Profile	Maximum depth (m)	Length (m)	Mean width (m)	Volume (10^3 m^3)
Bergsetbreen				
*	ND	ND	ND	25.6
*	ND	ND	ND	9.5
Mean	*ND*	*ND*	*ND*	*17.5±22.8*
Fåbergstølsbreen				
Fa	4.0	53	5.3	0.59
Fb	9.1	95	5.8	1.27
Fc	9.2	98	5.0	2.25
Fd	11.1	78	34.1	7.98
Fe	7.4	51	23.0	3.89
**	6.2	113	25.4	9.0
**	7.0	69	20.5	3.5
**	11.0	122	39.6	22.8
Mean	*8.1*	*85*	*19.8*	*6.41±5.5*
Lodalsbreen				
La	5.0	96	12.1	1.23
Lb	6.8	97	8.2	2.58
Lc	16.0	75	15.1	8.71
Ld	7.2	192	31.6	5.21
Mean	*8.7*	*115*	*16.8*	*4.43±3.8*

Table 4.1. Dimensions of gully systems surveyed at field sites around Jostedalsbreen. * data from Ballantyne (1995a); ** data from Ballantyne and Benn (1994). Note that volumetric values from Bergsetbreen represent the volumes of debris cones, not gullies. Mean volumes are expressed with 95% confidence limits. ND: no data.

Profile	Maximum depth (m)	Length (m)	Mean width (m)	Volume (10^3 m^3)
Leirdalen				
LEa	4.0	369	17.2	7.67
LEb	6.8	202	44.8	23.79
LEc	5.2	205	32.8	11.22
LEd	12.8	247	63.5	69.58
LEe	10.0	277	63.0	49.76
Mean	*7.8*	*260*	*44.3*	*32.4±26.5*
søre Illåbreen				
Sa	1.2	12.4	2.1	0.016
Sb	1.5	14.1	1.9	0.020
Mean	*1.35*	*13.3*	*2.0*	*0.018±0.01*
Visdalen				
Va	14.8	210	45.0	67.78
Vb	14.0	310	61.5	65.67
Vc	7.2	241	37.4	22.79
Vd	8.0	235	48.2	38.86
Ve	7.2	112	31.0	8.26
Mean	*10.2*	*222*	*44.6*	*40.67±26.1*
Heillstugubreen				
Ha	1.6	6.4	4.4	0.015
Hb	1.6	29.9	8.9	0.165
Hc	1.3	19.7	4.9	0.063
Hd	1.0	34.2	8.3	0.126
Mean	*1.4*	*22.6*	*6.6*	*0.092±0.08*

Table 4.2. Dimensions of gully systems surveyed at field sites in Jotunheimen. Mean volumes are expressed with 95% confidence limits.

4.3.1 Jostedalsbreen

The distribution of paraglacial modification of steep hillslope drift at the four Jostedalsbre field sites is shown on the geomorphological maps of these valleys (Figures 4.2 to 4.5), and quantitative indices summarising the extent of

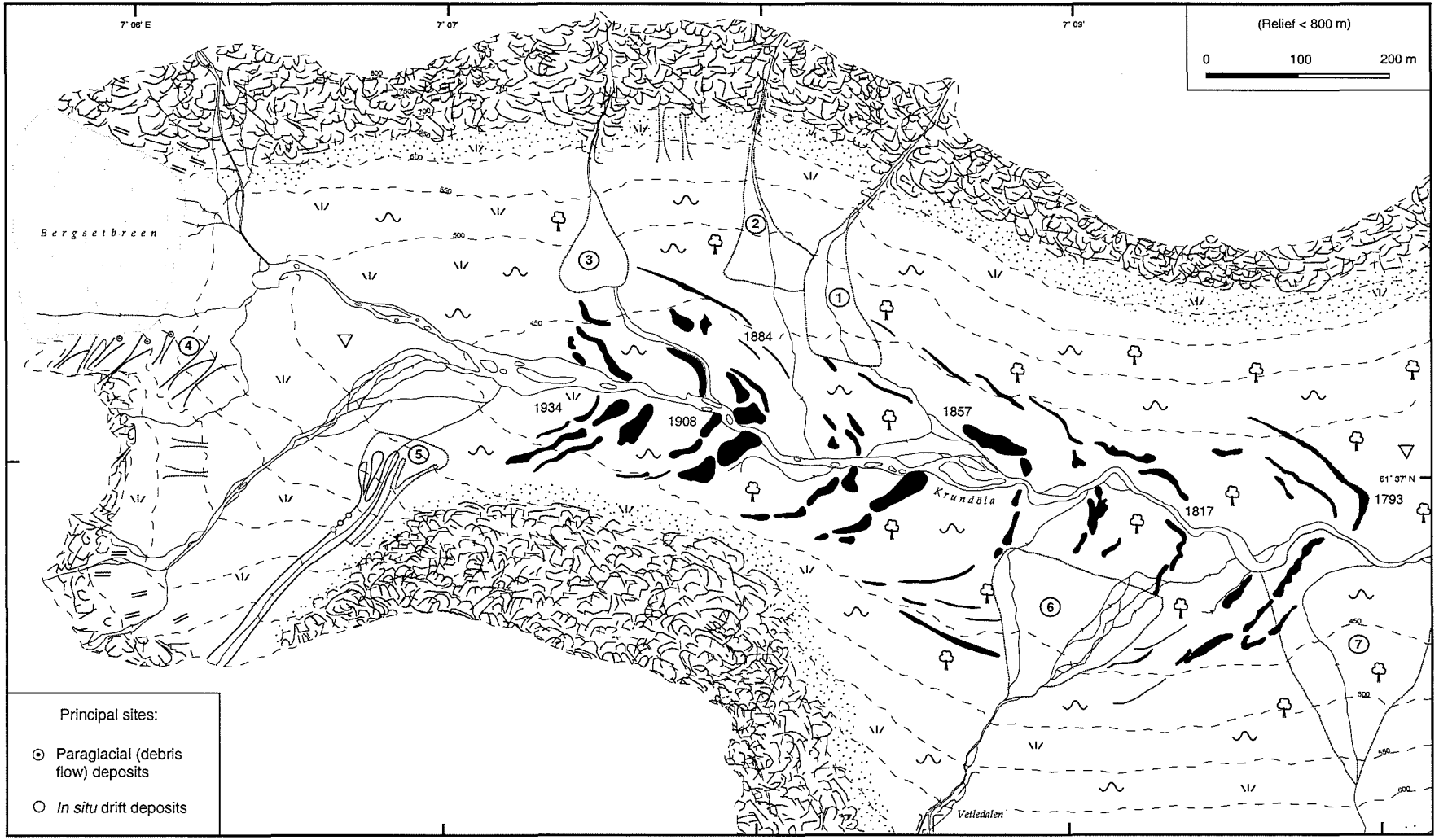


Figure 4.2. Paraglacial activity in the foreland of Bergsetbreen, Norway. Key in Figure 4.1.

paraglacial reworking at these sites are presented in Table 4.3. In Bergsetdalen (Figure 4.2) the upper rock slope rises above the valley to summit altitudes exceeding 1600 m, and is locally penetrated by deep re-entrants. At the foot of the rock slope a thin veneer of rockfall talus overlies drift. Elsewhere in the valley, glacigenic deposits mantle the entire foreland area, except where these have been removed by paraglacial activity. Resedimentation of drift within the limits of the 'Little Ice Age' glacial maximum is largely confined to a small area of gullying near the present glacier snout (site 4 in Figure 4.2) and five debris cones, four of which lie beneath rockwall gullies (sites 1-3, and 5 in Figure 4.2). Cones 2 and 3 contain respectively *c.* 25,600 m^3 and *c.* 9,500 m^3 of reworked sediment (Ballantyne, 1995a). The volume of cone 6 at the mouth of Vetledalen was calculated by Ballantyne to be in the order of 400,000 m^3, but this feature is exceptional in that it has formed through the reworking of debris derived from a broad tributary valley (Vetledalen) rather than from a valley-side gully. Although gully dimensions were not surveyed at this site, the volumes of the debris cones 2 and 3 suggest that no more than 5% of drift within Bergsetdalen appears to have been reworked since the onset of glacier retreat from the 'Little Ice Age' maximum (Table 4.3). Mean gully density in the Bergsetbre foreland is low (13 gullies per kilometre).

Field site	Gully density (n km^{-1})	% reworked drift*	Mean gully volume (10^3 m^3)*
Bergsetbreen	13	4.8±6.2	ND
Fåbergstølsbreen	45	15±13	6.4±5.5
Lodalsbreen	108	44±38	4.4±3.8
Nigardsbreen	20	ND	ND

Table 4.3. Extent of paraglacial modification at all field sites around Jostedalsbreen, Norway. Values apply to the uppermost 1 km in each foreland. * 95% confidence limits. ND: no data.

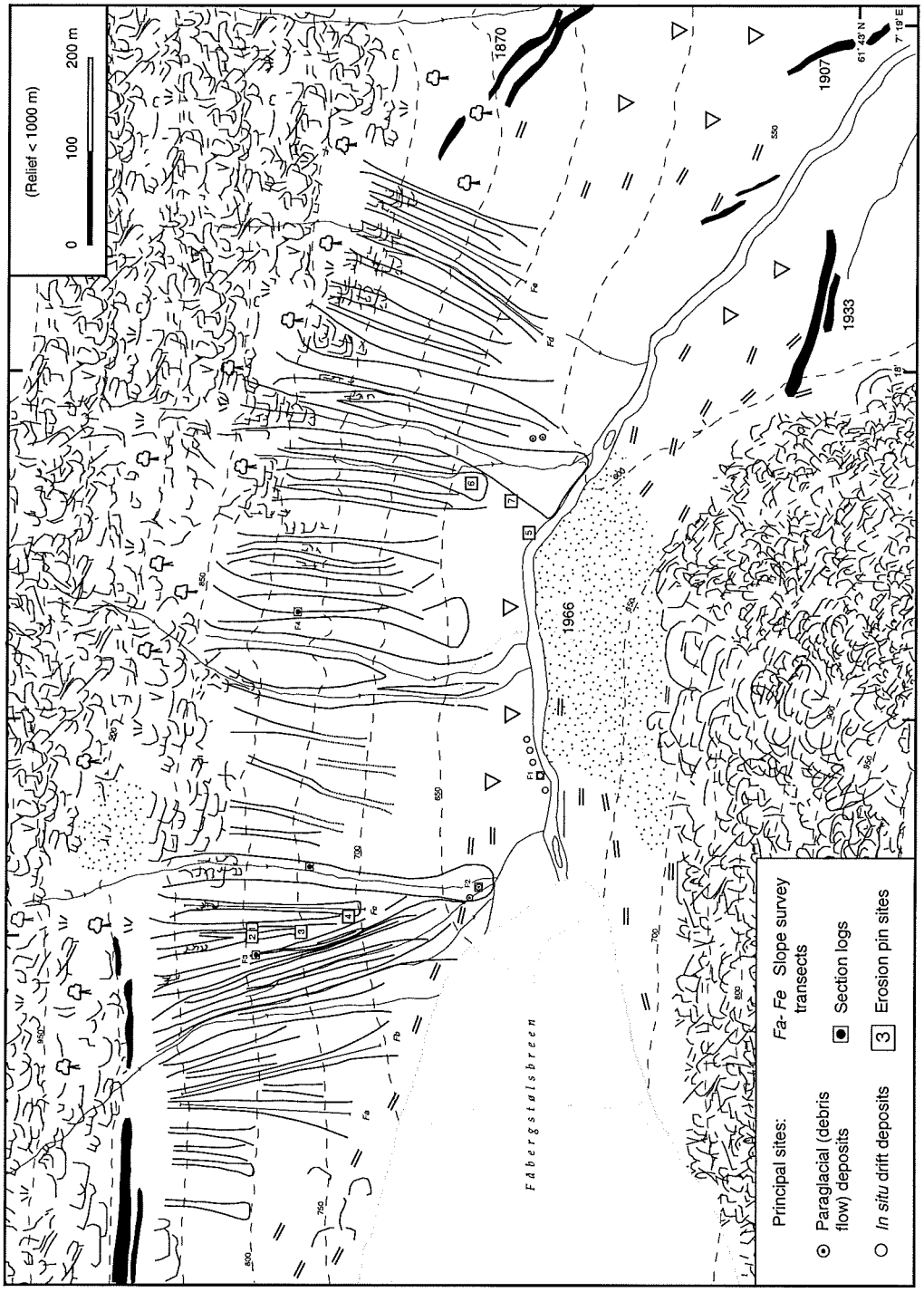


Figure 4.3. Paraglacial activity in the foreland of Fåbergstølsbreen, Norway. Key in Figure 4.1.

Investigation of the characteristics of paraglacial modification in Fåbergstølsdalen was confined to the upper valley between the glacier snout and the moraine marking the AD 1930 advance limit of Fåbergstølsbreen (Bickerton and Matthews, 1993; Figure 4.3). In this zone, the rockwall overlooking the south side of Fåbergstølsdalen is too steep to support drift deposits, and the valley floor beneath it consists chiefly of ice-moulded bedrock slabs that locally support a thin veneer of *in situ* drift, rockfall or avalanche debris. North of the river, where the underlying bedrock rests at a gentler gradient, three slope zones are represented: an upper zone of vegetated rock slabs which are deeply-cut by gullies draining the ice cap; a mid-slope zone dominated by broad gullies that incise valley-side drift deposits and locally expose bedrock in the gully heads; and a slope-foot zone comprising rock slabs, *in situ* drift and paraglacial debris cones and fans. Data from 9 slope surveys (including 3 sets of gully dimensions reported in Ballantyne and Benn, 1994; Table 4.1) indicate that gully volumes range from 590 m^3 to 22,800 m^3, with a mean value of 6,410 m^3. Gully density in Fåbergstølsdalen is 45 gullies per kilometre, and an estimated 15% (± 13% at 95% confidence) of valley-side drift has been reworked through paraglacial activity (Table 4.3).

Mapping was undertaken on the Lodalsbre foreland (Stordalen) between an end moraine marking the position of the glacier snout in *c.* AD 1826 (Bickerton and Matthews, 1993) and the gullies that lie farthest upvalley, approximately 750 m north of the snout of Lodalsbreen in 1996 (Figure 4.4). On either side of Stordalen the steep upper rockwalls are overlooked by the margins of the Jostedalsbre ice cap, which occasionally overhangs the south side of the valley. The rockwall is more extensively gullied than at the other Jostedalsbre sites, and is generally steeper, particularly above Lodalsbreen, where the underlying bedrock rests at *c.* 70°. Nevertheless, a thick cover of drift extends around the glacier foreland on all middle and lower slopes, and exhibits signs of widespread paraglacial modification. In Stordalen, drift on the south side of the

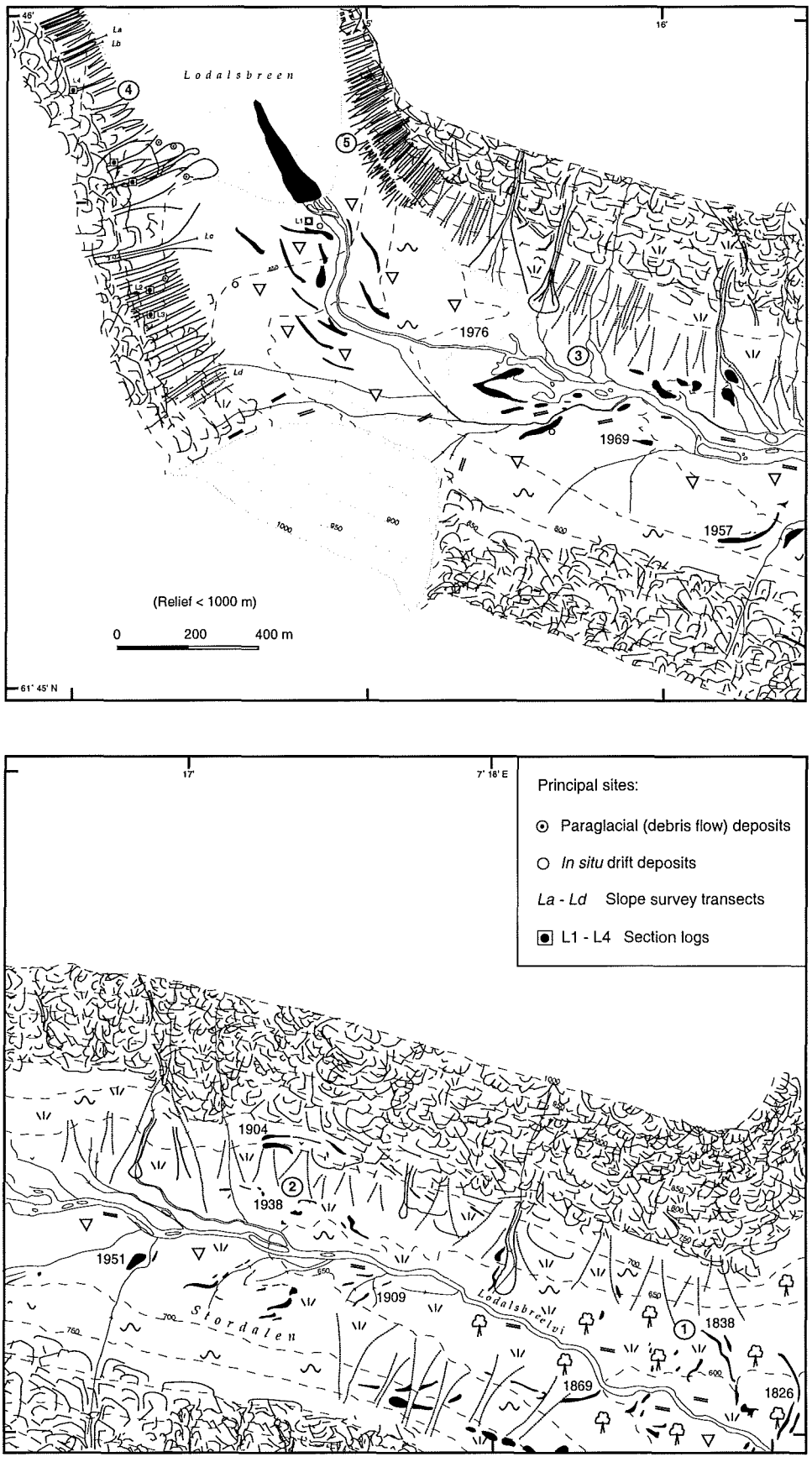


Figure 4.4. Paraglacial activity in the foreland of Lodalsbreen, Norway. Key in Figure 4.1.

river has been swept downslope and redeposited in a low-angled slope-foot debris apron, whilst on the north side drift is incised by gullies below the rockwall, and has been redeposited downslope in debris cones. In the immediate vicinity of Lodalsbreen, gullying is particularly extensive beneath massive, near-vertical cliffs on both valley sides. West of the glacier snout two sets of gullies and cones lie one above the other, each appearing to define a former glacier margin position and a corresponding drift limit. Survey of four gullies on the Lodalsbre foreland yielded estimates of sediment removed of *c.* 1,230 m^3 to *c.* 8,710 m^3, with a mean of 4,430 m^3 (Table 4.1). Gully density is extremely high (108 per kilometre), and approximately 44% (± 38% at 95% confidence) of valley-side drift has been reworked. Since this index is based on calculation of gully volume, it excludes the volume of redeposited drift stored in slope-foot aprons beneath ungullied yet reworked slopes, and hence probably underestimates the true proportion of reworked drift at Lodalsbreen.

Investigation of paraglacial activity on the Nigardsbre foreland was confined to the area between the snout and a moraine deposited in AD 1909 (Anderson and Sollid, 1971; Bickerton and Matthews, 1992; Figure 4.5). Although steep rock faces rise above the valley, only those cliffs southwest of the glacier snout channel meltwater directly from the Jostedalsbre ice cap onto the drift below. No slope surveys were carried out in the vicinity of Nigardsbreen, but drift on both valley sides and the valley floor appears to be thinner than at the other Jostedalsbre sites. Nevertheless, localised reworking of hillslope drift on the glacier foreland is manifest in gullying of the upper drift slope on the north side of Nigardsbrevatnet near the snout, and deposition of larger debris cones beneath rock clefts farther downvalley. Taken together, the extent of reworking around Nigardsbreen is represented by a gully density of 20 per kilometre of valley-side slope (Table 4.3).

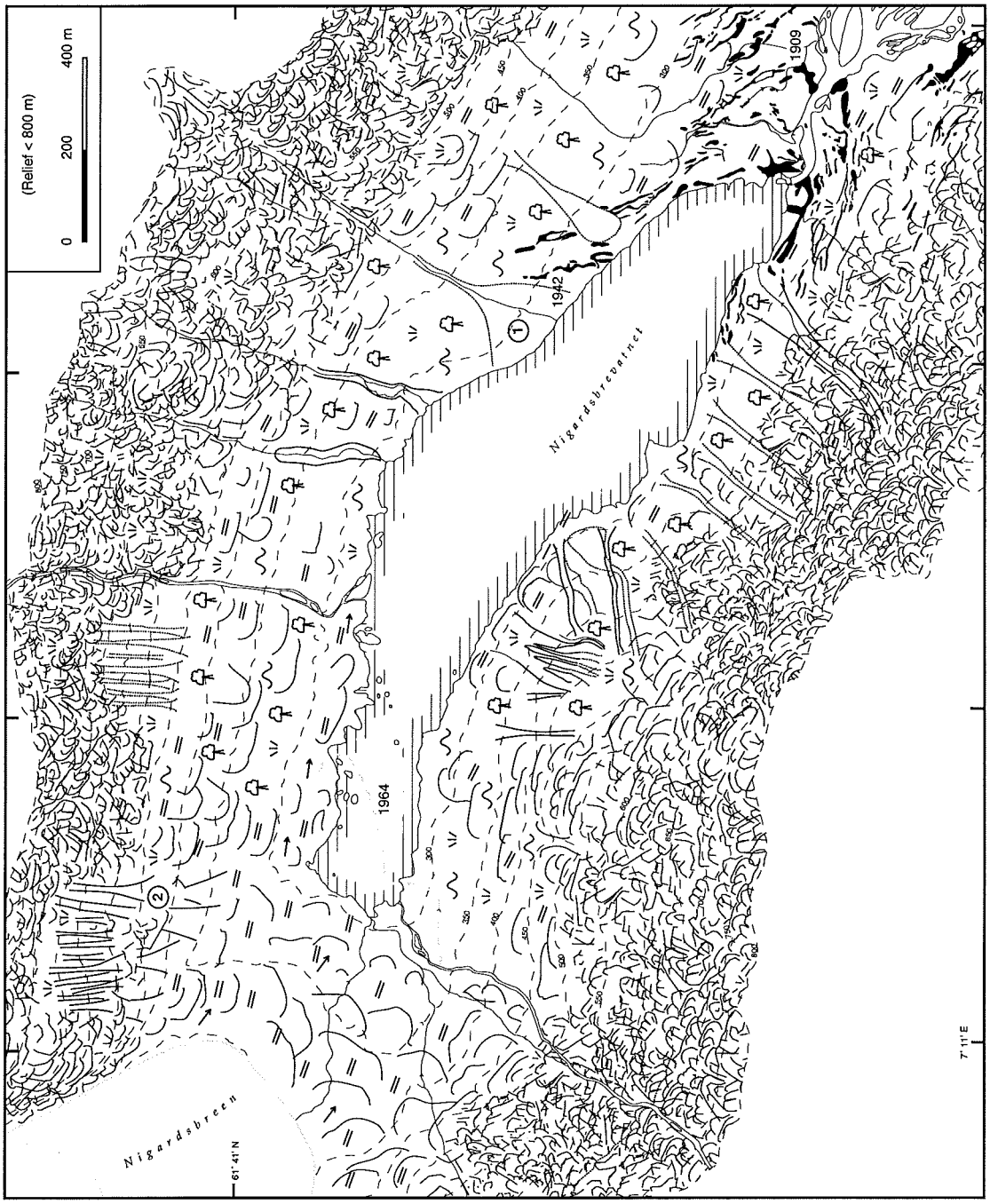


Figure 4.5. Paraglacial activity in the foreland of Nigardsbreen, Norway. Key in Figure 4.1.

4.3.2 Jotunheimen

The extent of paraglacial modification of steep hillslope drift at the Jotunheim field sites is illustrated in Figures 4.6 and 4.7, and summary indices are presented in Table 4.4. Mapping was carried out in Leirdalen from near the valley mouth southwards to within 2 km of Leirvassbu (Figure 4.6), and in Visdalen from the Spiterstulen mountain lodge to Visbreen (Figure 4.7). To permit the extent of drift reworking on recently-deglaciated terrain in Jotunheimen to be compared with that at the Jostedalsbre sites, summary indices of gully density, percentage reworked drift and gully volume were calculated for slopes within the limits of 'Little Ice Age' glacier expansion around Leirdalen (at søre Illåbreen) and Visdalen (at Heillstugubreen). Throughout both Leirdalen and Visdalen, the upper slopes are dominated by broken, gullied rock faces which support a thin but extensive cover of rockfall debris. This mid-slope debris mantle frequently overlies bedrock, except where talus has been transported downslope onto glacigenic drift. In both trunk valleys a break of slope is occasionally visible delineating the upper boundary of glacigenic drift from the talus apron at a height of *c.* 250 m above the valley floor. Gullying of the lower slope zone of glacigenic drift is usually initiated in shallow hollows below the break of slope, though the extent of modification is highly variable, reflecting variations in drift slope characteristics (considered below in section 4.5).

Within Leirdalen, gullying is more widespread on the eastern valley side (see Figure 4.8), particularly upvalley from Sauhøi, and in the extreme north of the field site, where recently-active and relict debris fans coalesce on the valley floor and bedrock is frequently exposed in gully floors. A few large cones and fans are found downslope of prominent rock gullies, or where proglacial streams join the trunk valley, such as east of Høgskridubreen. Paraglacial activity appears to be limited in the majority of glacier forelands around Leirdalen, and is largely

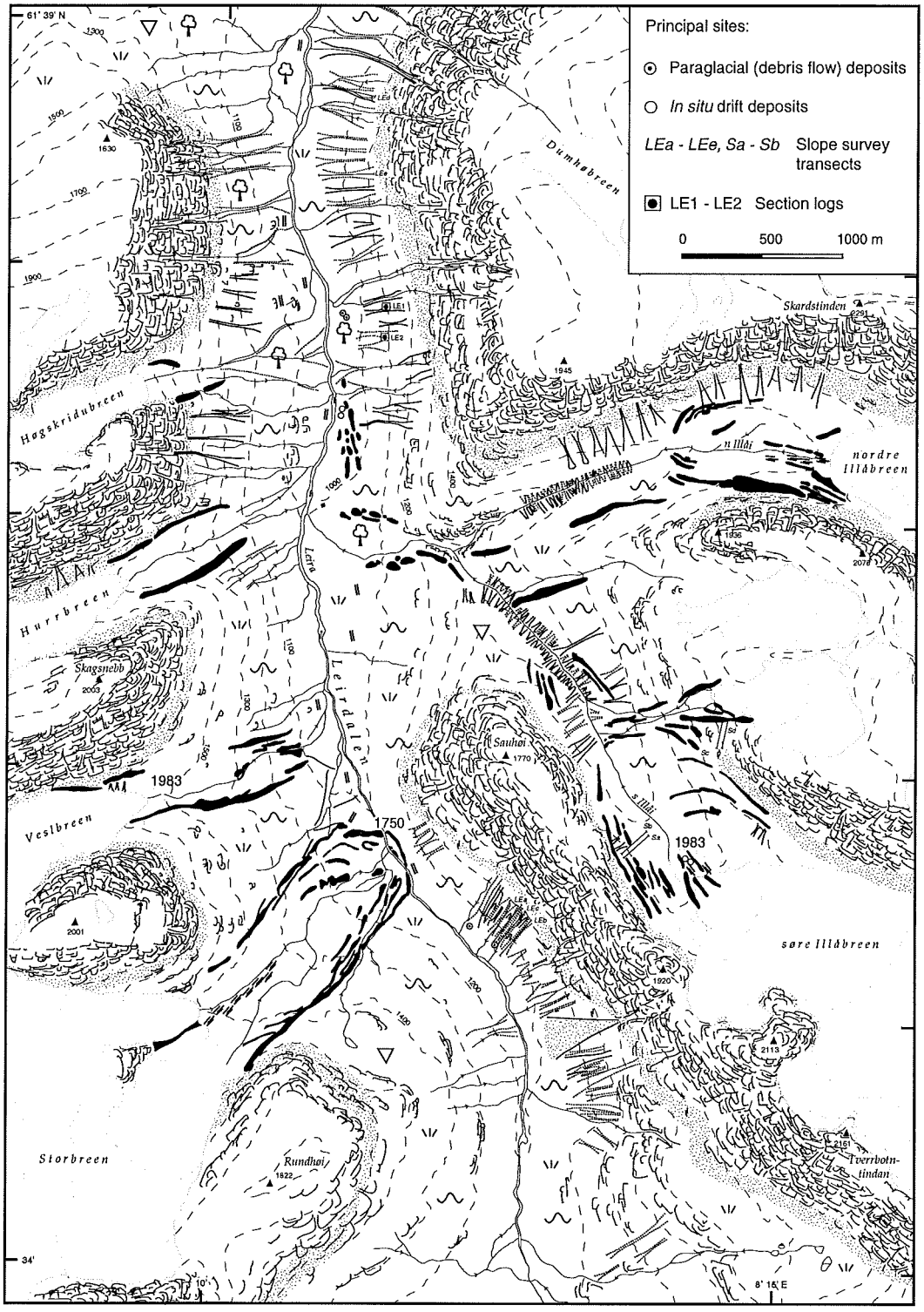


Figure 4.6. Paraglacial activity in and around upper Leirdalen, Norway. Key in Figure 4.1.

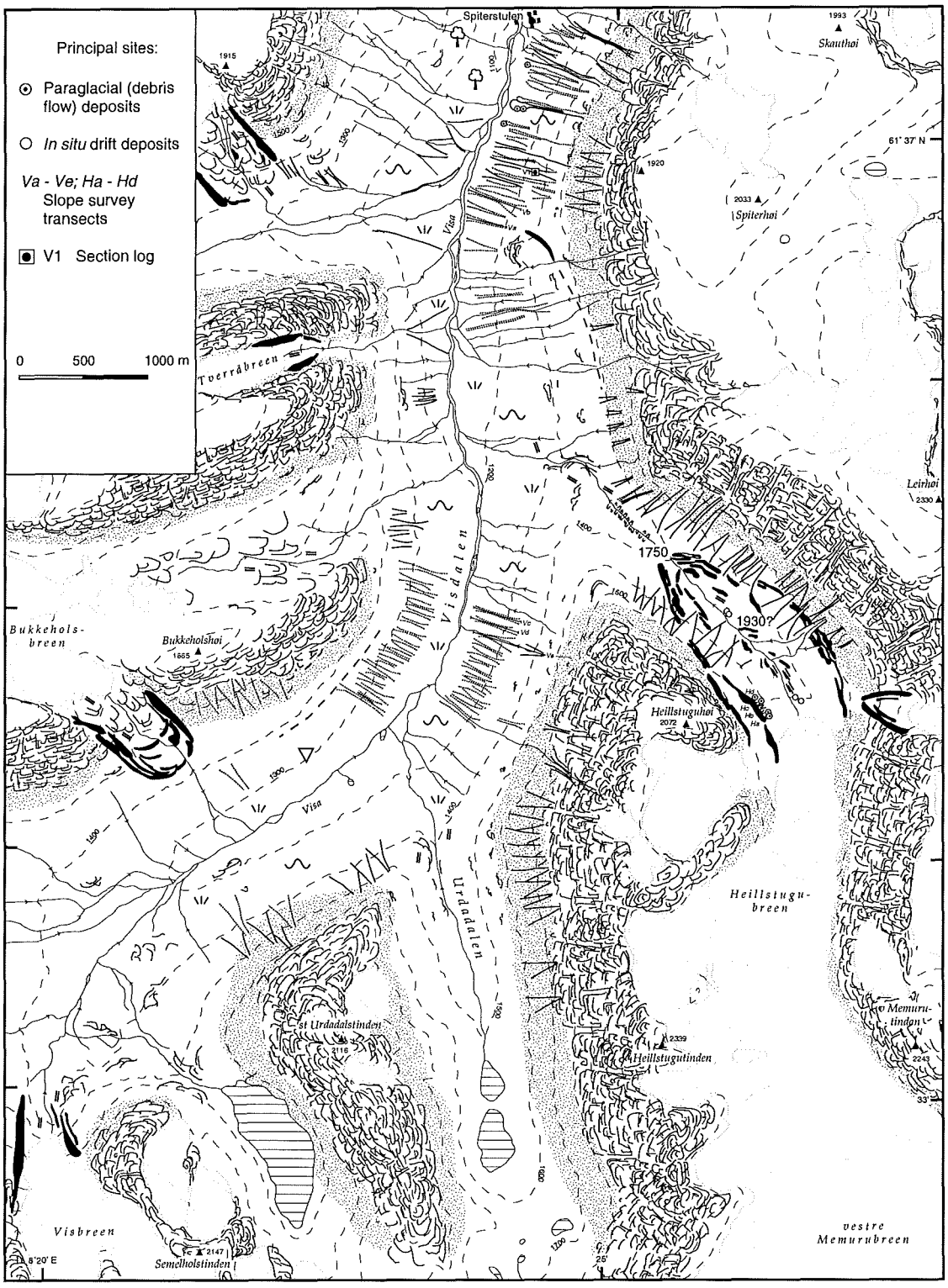


Figure 4.7. Paraglacial activity in and around upper Visdalen, Norway. Key in Figure 4.1.

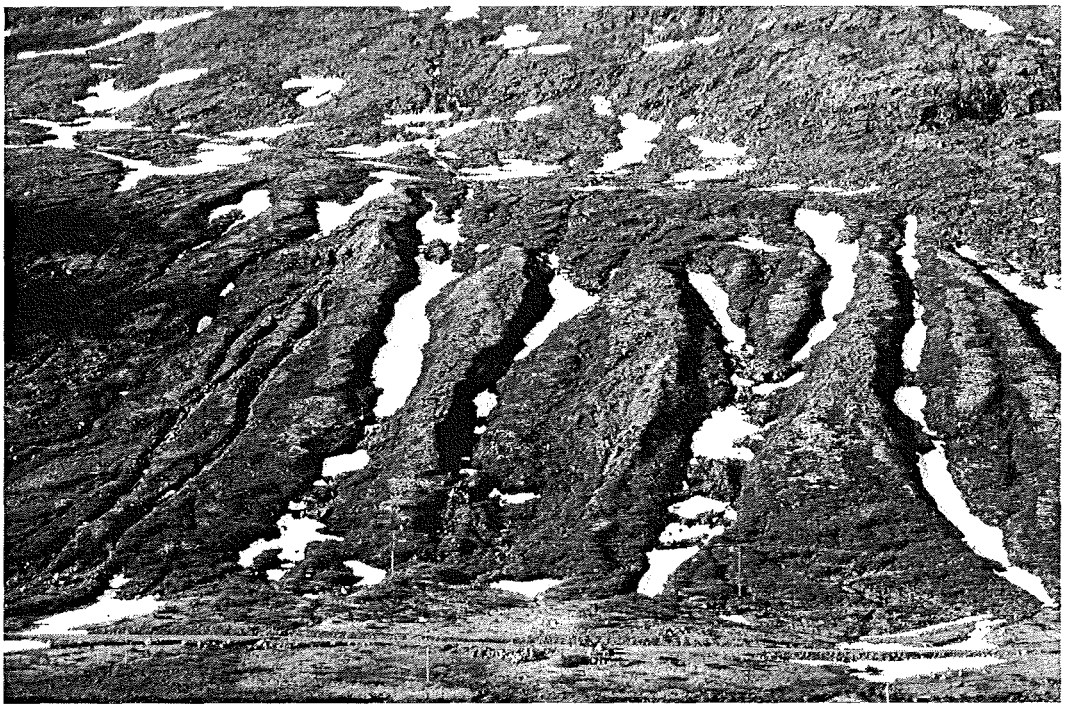


Figure 4.8. Largely-relict gully systems incising glacigenic drift in upper Leirdalen, Norway. Gullies emanate from a break of slope which delineates thick drift from a thin, upper mantle of drift and rockfall debris overlying bedrock. Notice bedrock exposed in gully floors and recent scarring of some intervening ridges of ungullied drift.

absent around Storbreen, Veslbreen and Hurrbreen, where lateral and end moraines show little sign of sediment reworking. Only in the forelands of nordre Illåbreen and søre Illåbreen is there notable modification of valley-side drift and deposition of debris aprons and cones on the valley floor. Both the nordre Illåi and søre Illåi rivers have downcut to bedrock, deeply incising valley floor drift, which is extensively gullied on either bank of both rivers. Five valley-side gullies in Leirdalen and two in the søre Illåbre foreland were surveyed. The volumes of the gullies surveyed in Leirdalen implied removal of *c.* 7,670 m^3 to *c.* 69,580 m^3 of sediment, with a mean estimate of *c.* 32,400 m^3 sediment removed per gully (Table 4.2). The valley-side gullies of the søre Illåbre foreland are very much smaller, and imply a mean of only *c.* 18 m^3 sediment removed per gully. For the trunk valley of Leirdalen, gully density was estimated to be 7.1 per kilometre, compared with 2.8 per kilometre for recently-deglaciated drift in the foreland of

søre Illåbreen (Table 4.4). The figures summarised in Table 4.4 suggest that around 6% of all glacigenic drift in Leirdalen has been redeposited downslope, compared to *c.* 0.01% around søre Illåbreen, although this latter figure excludes the volume of reworked sediment stored beneath ungullied slopes at this site.

Field site	Gully density (n km^{-1})	% reworked drift*	Mean gully volume (10^3 m^3)*
Leirdalen	7.1	6.42±5.25	32.4±26.5
søre Illåbreen	2.8	0.008±0.004	0.018±0.01
Visdalen	3.7	4.30±2.76	40.7±26.1
Heillstugubreen	8.0	0.07±0.06	0.092±0.08

Table 4.4. Extent of paraglacial modification at field sites in Jotunheimen, Norway. Values apply to the uppermost 1 km of drift slope at søre Illåbreen and Heillstugubreen, and a mean value per kilometre of valley-side slope for Leirdalen and Visdalen. * 95% confidence limits.

Modification of drift is less widespread in Visdalen than in Leirdalen, and is absent from all forelands except that of Heillstugubreen. Furthermore, gullies in the trunk valley are slightly shorter but deeper than those in Leirdalen (Table 4.2), and tend to commence from a break of slope only *c.* 200 m above the river Visa. Large fans of redeposited drift are found downvalley from Svellnosbreen and Tverråbreen, and are now incised by proglacial streams. In the Heillstugubre foreland distinct debris cones flank highly-concave drift slopes, or emanate from drift gullies lying beneath deep rock re-entrants. Occasionally these cones partially bury lateral moraines. The valley-floor deposits in this foreland are also fluvially-incised, creating steep, heavily-gullied drift slopes adjacent to the river. Throughout Visdalen, gully density is estimated to be 3.7 per kilometre, compared with 8.0 per kilometre for the recently-deglaciated Heillstugubre foreland. As gully density is lower in Visdalen than in Leirdalen, rather less drift (*c.* 4.3% ± 2.8%) has been reworked in the former. The dimensions of surveyed gullies in

Visdalen imply *c.* 8,260 m^3 to *c.* 67,780 m^3 of sediment removal, with an average of *c.* 40,700 m^3 (Table 4.2). Surveyed gullies on the Heillstugubre foreland, however, range from only 15.1 m^3 to 164.9 m^3 in volume, with an average of *c.* 92 m^3, suggesting that only *c.* 0.07% of drift has been reworked. Although this figure undoubtedly underestimates the true amount of drift resedimentation, it is similar in magnitude to that estimated for the foreland of søre Illåbreen. Both, values, however, are very much less than those estimated for the forelands of glaciers draining Jostedalsbreen.

4.3.3 Extent of paraglacial modification of drift: summary

The extent of paraglacial activity varies considerably, but may be extremely widespread where conditions are favourable. Gully density ranges from *c.* 5 gullies per kilometre at those sites exhibiting very limited reworking to *c.* 100 gullies per kilometre where paraglacial modification of hillslope drift is most active. This represents resedimentation of valley-side drift ranging from < 0.1%. at the least active sites in the Jotunheim to an estimated 44% (± 38% at 95% confidence) at Lodalsbreen. Possible factors explaining these variations in the extent of paraglacial reworking are examined in section 4.5 below.

4.4 Processes of sediment transfer.

At all of the Norwegian field sites, one of the most conspicuous legacies of drift slope modification takes the form of debris cones or fans located along the flanks of formerly glaciated valleys, reflecting paraglacial reworking of glacigenic drift on hillslopes. Several processes, including debris flow, snow avalanches and fluvial activity, have contributed to sediment redistribution, and the significance of each is assessed below.

4.4.1 Debris flows

Jostedalsbreen

In Bergsetdalen and at Nigardsbreen, the absence of exposures in the large debris cones precludes analysis of constituent sediments, and information on the nature of depositional processes has to be inferred from superficial deposits and surface topography. However, on several of the cones at Bergsetbreen and on all the cones and fans mapped at Nigardsbreen, the dominant microrelief consists of vegetated paired levées and terminal lobes indicative of former debris flow activity across the entire cone surfaces (*cf.* Kotarba and Strömquist, 1984; van Steijn *et al.*, 1988). This suggests that modification of valley-side drift by debris flows has been the dominant formative process on these cones. On the most recently-exposed drift beside both Bergsetbreen and Nigardsbreen, active debris flow tracks emanate from gullies cut in the drift. At Fåbergstølsbreen and Lodalsbreen numerous debris flow tracks descend from gullies cut into drift across cone and fan surfaces (Figure 4.9). As observed elsewhere, flow paths are marked by the deposition of parallel levées of unsorted debris (*cf.* Rapp and Nyberg, 1981; Larsson, 1982; Kotarba and Strömquist, 1984; van Steijn *et al.*, 1988). Ballantyne and Benn (1994) noted that in upper Fåbergstølsdalen later generations of flows have often cut across earlier flows, producing a complex microtopography of dissected levées up to 2 m high on fan surfaces. Some debris flow deposits in Fåbergstølsdalen are initially emplaced on late-lying snowbeds (Figure 4.10), resulting in a chaotic assemblage of ephemeral snow-cored debris mounds and ridges as the surrounding snow melts. These melt-out debris flow deposits are particularly common at the foot of gullies around the snout of Lodalsbreen. Farther upvalley, debris flows often discharge lobes of debris directly onto the surface of Lodalsbreen.

Figure 4.9. Debris flow tracks proceeding from gullies cut into steep, valley-side drift in the recently-deglaciated Fåbergstølsbre foreland, Norway. Debris flows descend across fan and cone surfaces depositing parallel levées of unsorted debris overlying *in situ* drift and bedrock.

Jotunheimen

Both ancient and recent modification of the lower valley-side drift in Leirdalen and Visdalen are dominated by debris flow activity. Clear exposures in gully sidewalls are rare, so information on the nature of depositional processes has to be inferred from surface topography and superficial deposits. Vegetated cones and fans at the foot of sections of gullied drift in the Leirdal and Visdal trunk

Figure 4.10. Hummocky debris flow deposits overlying snow-banks opposite the current snout position of Fåbergstølsbreen, Norway. Such deposits will result in a chaotic assemblage of ephemeral snow-cored debris mounds and ridges as the surrounding snow melts.

valleys support an irregular microrelief of hummocky levées and lobes indicative of former debris flow activity. On some slope-foot fans, particularly those at the northern end of either valley, and northeast of Rundhøi in Leirdalen, fresh debris flow deposits extend downslope from drift gullies, providing further support for the interpretation of cones and fans as primarily the product of repeated debris flow activity. Matthews *et al.* (1997) have described a fine-grained minerogenic layer overlying the valley floor in part of upper Leirdalen, and attribute its

deposition to a debris flow event in 1994. In Visdalen, fans often occur beneath gullies of moderate gradient, suggesting that some valley-confined debris flows may have followed stream channels (*cf.* Haeberli, 1996). Where valley-side drift slope modification has occurred within the 'Little Ice Age' glacier limits around Leirdalen and Visdalen, debris flow activity is much less extensive than in the trunk valleys. In the forelands of nordre Illåbreen and søre Illåbreen only a few debris flow tracks incise the valley-side drift, though at Heillstugubreen fresh debris flow deposits are more widespread along recently-exposed slopes. Only along the incised valley-floors of the nordre Illåi, søre Illåi and Heillstuguåi streams have debris flows removed and redeposited large amounts of drift in these valleys.

In many mountain areas, debris flows are triggered by intense rainstorms or rapid snowmelt through a rise in pore-water pressures (e.g. Sharp and Nobles, 1953; Theakstone, 1982; Innes, 1983b; Johnson and Rodine, 1984). At all the sites investigated during June or July in 1996 and 1997, snowpatches were commonly found at gully heads. In upper Fåbergstølsdalen five debris flows occurred during nine days of warm dry weather in 1992, and have been attributed to ground saturation and failure due to rapid melt of gully-head snowpatches (Ballantyne and Benn, 1994). This suggests that the main period of debris flow activity at these sites may correspond to the spring snowmelt period, though Nesje (cited in Ballantyne and Benn, 1994) has also observed the occurrence of debris flows in the Jostedalen area during intense autumn rainstorms.

Frequently, debris flows begin as shallow landslides that undergo progressive liquifaction, ultimately behaving as a viscous fluid (Rodine and Johnson, 1976). The widespread exposure of bedrock in broad funnel-shaped gully heads throughout the Norwegian field sites suggests that translational sliding of drift over bedrock has been instrumental in initiating debris flows in this

way (Ballantyne and Benn, 1994). Whether or not facilitated by fluvial incision at the top of the drift slope, debris flow tracks often appear to evolve by downslope channelisation and liquifaction of debris from a shallow translational failure. Nonetheless, fresh scarring is commonly evident on the steep gully sidewalls, indicating that loose debris accumulates on gully floors until it is evacuated by debris flows (*cf.* Statham, 1976).

4.4.2 Snow avalanches

Jostedalsbreen

Snow avalanches occur annually in many of the valleys around Jostedalsbreen. Although such avalanches have great erosive and transportational potential (e.g. Rapp, 1960; Luckman, 1977; Ward, 1985), entrainment of debris may be hindered by ground freezing, a stationary snow cover and/or a protective vegetative cover (Luckman, 1977; Gardner, 1983; Butler and Malanson, 1990). In Bergsetdalen occasional fresh angular boulders are strewn across the vegetated surfaces of cones 2 and 3 (Figure 4.2), indicating recent avalanche activity. On cone 6, damaged vegetation, numerous perched boulders, and drapes of fine material over some clasts provide evidence for recent avalanche transport. Ballantyne (1995a) noted snow avalanche deposition across the surface of this cone annually between 1988 and 1993, though the amount of debris transported by such avalanches now appears limited. Evidence for reworking of debris by snow avalanches is also common within Fåbergstølsdalen, in the form of debris tails on the lee (downslope) sides of large boulders and melt-out clasts perched in unstable positions atop boulders. Much of the avalanche debris lies beneath the larger gullies, whose heads provide source areas for snow accumulation, although damaged heath vegetation and fresh angular clasts on the valley floor in the central part of Fåbergstølsdalen suggests recent avalanches have also carried

debris over this ungullied part of the drift slope. Similar signs of avalanches sweeping debris down ungullied slopes are widespread along the southern flank of Stordalen. In contrast to the predominantly debris flow-reworked slope across the valley, the southern slope exhibits a very pronounced basal concavity, has a lower gradient, and is ungullied and largely drift-free. Toppled willows and birches extend for *c.* 2 km along the foot of the slope amid a chaotic assemblage of angular boulders and melt-out clasts. Small avalanches triggered by collapsing fragments of the ice cap at the crest of the cliff were observed to sweep part of the southern rock face during July 1997, and the presence of a large avalanche cone on the valley floor suggests that more substantial avalanches sweep over the lower drift slopes in the spring months. Avalanche debris is also evident at the foot of gullies above Lodalsbreen. On the foreland of Nigardsbreen, snow avalanche deposits are restricted to the foot of rock faces, and drift slopes appear to be less extensively modified by snow avalanches than at other sites around Jostedalsbreen.

Jotunheimen

Snow avalanches in upper Leirdalen and Visdalen appear to be restricted to the upper slopes, which predominantly consist of talus overlying thin drift cover. Nevertheless, the surfaces of some of the large cones at the northern end of both trunk valleys show signs of limited reworking by snow avalanches, such as the occasional presence of flattened vegetation and angular blocks transported from the rockfall debris above. However, where drift has been redeposited downslope in the tributary valley forelands, snow avalanching appears to be the primary process of drift modification. At nordre Illåbreen, søre Illåbreen and Heillstugubreen extensive aprons of debris flank concave, ungullied, avalanche-swept drift slopes, and valley-side lateral moraines have locally been buried by avalanche tongues that emanate from deep rock gullies. This is particularly

evident at Heillstugubreen, where reworking of drift is more extensive than at any other 'Little Ice Age' foreland mapped in Jotunheimen. Whilst the trunk valley-side drift slopes in Leirdalen and Visdalen support a closed vegetation cover which may impede avalanche activity, the absence of such a cover on the bare, open upper slopes at nordre Illåbreen, søre Illåbreen and Heillstugubreen probably favours effective removal of drift cover by snow avalanching at these localities.

4.4.3 Surface wash

Jostedalsbreen

Valley-side streams appear to play only a secondary role in transporting sediment at the Jostedalsbre sites, although some streams have cut shallow channels through debris flow deposits and have locally deposited thin spreads of sand and silt on valley floors. In Fåbergstølsdalen a number of gully floors are incised by ephemeral streams, and sediment removal and exposure of bedrock has resulted from surface runoff within some gully heads. However, a proglacial stream in upper Bergsetdalen has cut a deep gully through fresh valley-side drift upslope from cone 5 (Figure 4.2). The size of this gully implies the removal of at least 90,000 m^3 sediment. In addition, Ballantyne (1995a) describes miniature boulder berms in the vicinity of stream channels at the apices of two cones in Bergsetdalen, suggesting localised reworking of debris by flood torrents. In contrast, little evidence of fluvial reworking of drift is apparent at Lodalsbreen and Nigardsbreen.

Jotunheimen

The number of drainage channels cutting through drift at the Jotunheim sites is significantly less than on the Jostedalsbre forelands, and hence the

significance of surface wash processes in modifying drift slopes in and around Leirdalen and Visdalen appears less pronounced. However, at both sites several vegetated gully floors have been incised to bedrock by small streams draining the upper rock slopes, resulting in terraced gully cross-profiles and the deposition of fine-grained colluvium on cones and fans. Localised accumulations of fine sediment immediately downslope from shallow erosional scars at or near the crest of ungullied drift in upper Leirdalen and upper Visdalen also suggest recent slopewash. Several large, low-angled fans have been deposited by proglacial streams on the valley floors of Leirdalen and Visdalen, and these have subsequently been entrenched by the parent streams. These landforms are interpreted as paraglacial alluvial fans emplaced shortly after glacier thinning or retreat exposed drift deposits to fluvial reworking. Within the foreland areas, however, surface wash processes appear to have had limited effect in reworking valley-side sediment, although valley floor fills have been significantly entrenched by the nordre Illåi, søre Illåi and Heillstuguåi outlet streams.

4.4.4 Processes of sediment transfer: summary

The planimetric area of drift resedimented by debris flows, snow avalanches and surface wash was obtained by digitising cartographic data from Figures 4.2-4.7 on the basis of detailed field observations. Table 4.5 outlines the relative areal extent of debris flow, snow avalanche and surface wash deposits as a percentage of the total reworked drift cover at each site. At the majority of sites, debris flow deposits account for more than 50% of reworked sediment by area, though on the forelands of søre Illåbreen and Heillstugubreen snow avalanche deposits are more widespread. However, it should be noted that the relative cover of each type of deposit is not an accurate reflection of the volume of sediment transported by each of the three processes, nor of the geomorphologic importance of each process in terms of drift remobilisation. As snow avalanche debris often

Field site	% Debris flow	% Snow avalanche	% Surface wash
Jostedalsbreen			
Bergsetbreen	49	39	12
Fåbergstølsbreen	67	27	6
Lodalsbreen	52	43	12
Nigardsbreen	65	18	17
Jotunheimen			
Leirdalen	58	19	23
søre Illåbreen	23	72	5
Visdalen	53	21	26
Heillstugubreen	20	74	6

Table 4.5. Percentage areal extent of debris flow, snow avalanche and surface wash deposits as components of total reworked drift at field sites in Norway.

consists of no more than a thin surface veneer whilst debris flows produce thick terminal lobes and substantial lateral levées, the widespread presence of snow avalanche deposits at any site need not necessarily imply that snow avalanching has been the dominant paraglacial process. Taking into account the generally much greater thickness of debris flow deposits, together with the spatial dominance of such deposits at most sites, it is evident that (in terms of volume of sediment transported) debris flow activity is undoubtedly the most effective process in the paraglacial sediment transfer system.

4.5 Constraints on paraglacial modification of drift.

Little research has hitherto been devoted to investigation of the physical constraints that govern paraglacial modification of steep drift slopes. Spatial variations in the intensity of paraglacial reworking of hillslope drifts probably reflect the complex interaction of both intrinsic and extrinsic controls, but few such controls have been specifically identified. Whilst topographic setting, relief

amplitude and bedrock structure undoubtedly dictate the energy available to all slope processes and the availability of entrainable sediment mantling steep slopes within a basin, the dataset reported in this thesis is insufficiently large to establish the significance of these controls on paraglacial activity. Ballantyne (1995a) suggested that slope thresholds may be important, and Ballantyne and Benn (1994) inferred that the thickness of hillslope drift may also be influential in determining whether or not extensive translocation of sediment takes place, but neither inference has been rigorously tested. Here the possible importance of a number of intrinsic and external controls of paraglacial reworking are considered, using gully density as a surrogate measure of intensity of paraglacial reworking of drift.

4.5.1 Intrinsic controls

To aid identification of the potential intrinsic controls on gully density at the Jostedalsbre and Jotunheim field sites, four hypotheses were tested using data amalgamated from both areas. These were: (1) that gully density increases with drift-slope gradient; (2) that gully density increases with drift thickness; (3) that gully density is inversely related to sediment packing; and (4) that gully density is related to matrix granulometry. To facilitate such analyses, gully density at each site was plotted against upper drift slope gradient, drift thickness, drift matrix void ratio, and percentage of sand (60-2000 μm) and silt (2-60 μm) within fine-grained (< 2 mm) samples of drift (Figure 4.11). Where appropriate, the significance of individual bivariate relationships was tested using Spearman's rank correlation coefficient (r_s).

Slope gradient is widely recognised as a major intrinsic control of hillslope debris flow initiation (e.g. Takahashi, 1981; Innes, 1983a; Costa, 1984). Whilst the correlation between gully density and slope gradient values from the

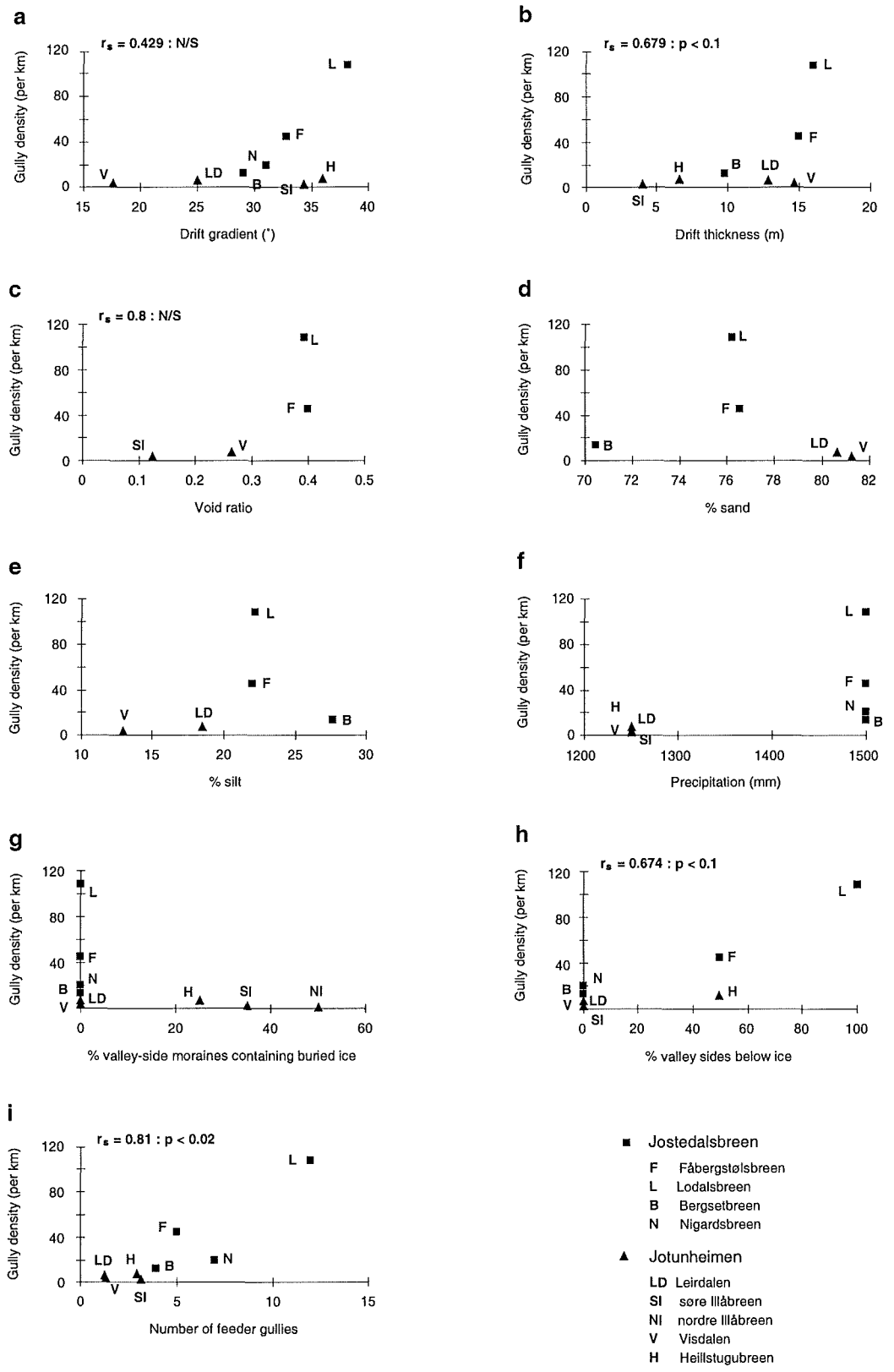


Figure 4.11. Gully density relations with (a) drift gradient, (b) drift thickness, (c) matrix void ratio, (d) % sand content, (e) % silt content, (f) regional precipitation, (g) the extent of buried ice within valley-side drift, (h) % of valley sides lying immediately beneath ice and (i) the number of feeder gullies at field sites around Jostedalsbreen and in the Jotunheim. Spearman's rank correlation coefficient is denoted by r_s. p describes the significance level of the correlation. N/S: not significant.

Norwegian field sites is not statistically significant at $p < 0.1$, the two variables appear positively related to each other. Figure 4.11a suggests that a threshold gradient exists at *c.* 30°, below which gully density values are low (< 20 km^{-1}). This seems to imply that a drift-slope gradient exceeding *c.* 30° represents a prerequisite for extensive paraglacial modification of valley-side drift, but that other factors determine whether extensive gullying actually occurs on drift slopes steeper than 30°. At both søre Illåbreen and Heillstugubreen gully densities are low (< 8 km^{-1}), even though upper drift slope gradients appear sufficiently steep (34.4° and 36.1° respectively) for widespread paraglacial modification.

Another possible control on the extent of paraglacial activity may be drift thickness. Plotting gully density from seven of the Norwegian field sites against drift thickness (Figure 4.11b) reveals a positive relationship between the two variables significant at $p < 0.1$. This evidence seems to contradict the suggestion by Ballantyne and Benn (1994) that the lack of gullies in the central part of the Fåbergstølsbre foreland may reflect thicker drift cover, as the converse appears to be the case. A drift thickness threshold of 10-15 m appears to differentiate those sites where gully density is low (>20 km^{-1}) from those where gully spacing is dense. However drift thickness *per se* does not seem to be the sole control on gully density, for although drift thickness averages 14 m in Leirdalen and Visdalen, gully density values at these sites are very low indeed (7.1 km^{-1} and 3.7 km^{-1} respectively), possibly reflecting the low drift slope gradients in these trunk valleys.

The third possible intrinsic control on gully density suggested above is drift matrix void ratio. It was hypothesised that where drift is loosely packed (high void ratio), higher infiltration rates may lead to a more rapid rise in pore-water pressures during intense precipitation, causing a greater reduction in shearing resistance than where drift is tightly packed (low void ratio). Mean void

ratio data are available for only four sites. As a result, the relationship between gully density and void ratio is not statistically significant, but the plot (Figure 4.11c) suggests a positive correlation between the two variables, with a possible threshold void ratio of 0.30-0.35 separating sites that exhibit dense gullying from those with very low gully densities.

The fourth hypothesis suggested above concerns the possibility that matrix granulometry exerts a control on the extent of paraglacial reworking of drift. On the basis of extensive granulometric analysis of frost-weathered regoliths in the Scottish Highlands, Innes (1982; 1983c) observed that debris flow activity appears to be largely restricted to lithologies yielding a sand-rich regolith, and less widespread on silty regoliths. Susceptibility of areas of sandy drift or regolith to debris flow may reflect associated high infiltration rates (Ballantyne, 1986). However, Figures 4.11d and 4.11e reveal no clear relationship between gully density and either percentage sand or percentage silt content, suggesting that granulometric composition is not a primary intrinsic control on the degree of paraglacial reworking of valley-side drift.

In sum, hypothesis testing based on the relationships between gully density and possible intrinsic controls suggests that paraglacial reworking of drift is very limited (1) where upper drift slope gradient is less than *c.* 30°, (2) where drift is less than 10-15 m thick, and (3) where the void ratio of drift matrix material is less than 0.30-0.35. The analyses summarised in plots (a) - (e) in Figure 4.11, however, consider only intrinsic controls related to drift gradient, thickness and sediment properties. The influence of possible extrinsic controls is considered below.

4.5.2 Extrinsic controls

Four hypotheses relating to possible extrinsic controls of paraglacial activity at the Norwegian field sites were considered. Each of these evaluates possible hydrological controls on the extent of paraglacial activity, again using gully density as a surrogate for degree of drift reworking. The four possible extrinsic controls considered are: (1) that gully density increases with regional precipitation; (2) that gully density is greatest where drift overlies buried ice; (3) that gully density increases where an ice cap above the drift slope drains onto the slope; and (4) that gully density increases with the number of feeder gullies delivering water to the drift. To evaluate these possible controls, gully density was plotted against mean annual precipitation (Figure 4.11f), the extent of buried ice within valley-side drift (Figure 4.11g), the percentage of valley sides lying immediately downslope of glacier ice (Figure 4.11h) and the number of rockwall feeder gullies immediately upslope of valley-side drift (Figure 4.11i). Calculations for Figure 4.11i were limited to the uppermost kilometre of each foreland field site and represented by a mean value per kilometre in Leirdalen and Visdalen. Two cautionary points need to be noted here. The first concerns possible auto-correlation amongst these extrinsic variables. Unlike the intrinsic controls considered above, which are essentially independent of each other (except for the granulometric variables, which may relate to void ratio), the four possible extrinsic controls may be inter-related. Second, since gully maturation is generally self-enhancing (*cf.* Horton, 1945), gully density at any site may reflect past extrinsic constraints at the time of gully initiation as well as present controls.

A positive relationship clearly exists between gully density and regional mean annual precipitation (Figure 4.11f). Although the regional precipitation data do not take into account local precipitation patterns, in general the wetter climate of the Jostedalsbre forelands appears to favour much more intense reworking of

drift slopes. This relationship is consistent with the results of studies that highlight the necessity of abundant water for the initiation of debris flows through an increase in pore-water pressure and consequent decrease in shearing resistance (e.g. Caine, 1980; Rapp and Strömquist, 1976; Lawson, 1982; Wieczorek and Jäger, 1996). However, it should be borne in mind that mean annual precipitation *per se* is not necessarily a direct influence on the frequency of slope failure and debris flow events, which is essentially controlled by the frequency of prolonged, intensive rainstorms and antecedent moisture conditions (Watanabe, 1985; Church and Miles, 1987; Kotarba, 1987; Zimmermann and Haeberli, 1992).

The second hypothesis considered was that variations in the degree of paraglacial drift reworking may be explained, at least in part, by the extent of buried ice within valley-side drift (*cf.* Lewkowicz, 1987a, 1987b; Mattson and Gardner, 1991). Valley-side drift accumulations underlain by ice may be expected to exhibit greater instability because of the potential for saturation of drift during melt of such ice. However, gully density appears to be negatively correlated with estimated extent of buried ice (Figure 4.11g), suggesting that the latter plays little or no part in controlling the degree of paraglacial drift reworking.

The third hypothesis considered is that valley-side drift lying directly downslope of glacier ice would be expected to experience protracted wetting and saturation during the ablation season, and would thus be potentially more prone to failure and flow. Analysis of the data appears to support this suggestion. Figure 4.11h indicates a moderate but significant (at $p < 0.1$) correlation between gully density and the percentage of valley sides lying immediately downslope of glacier ice. This relationship is strongly influenced, however, by the high gully density at just two sites (Fåbergstølsbreen and Lodalsbreen); on the Heillstugubre foreland, 50% of valley sides lie downslope of glacier ice, yet gully density is only 8 km^{-1}.

An apparent link between hydrology and the extent of paraglacial activity is also evident in a moderate correlation (significant at $p < 0.02$) between gully density and the number of feeder gullies on the rockwall upslope (Figure 4.11i). Feeder gullies not only act as source areas for snow avalanches, but also focus water onto the drift below. However, despite the significant positive relationship between gully density and number of feeder gullies, the latter are much less numerous that the former. This suggests that the link between the two is indirect, and that other variables are of greater importance in influencing drift gully spacing. Field observations suggest that though drift gullying often occurs directly downslope of individual feeder gullies, in some locations it is also widespread where water is delivered to the drift between large feeder gullies, or through 'sheetflow' over steep, massive, ungullied cliffs, as occurs at the Lodalsbre snout (section 4.3.1 above).

Although possible auto-correlation of independent variables does not permit firm conclusions to be reached concerning the extrinsic controls on paraglacial drift modification, the above relationships nonetheless suggest that delivery of water to the slope is of critical importance. In particular, gully intensity is positively correlated with regional mean annual precipitation, the degree to which slopes are overlooked by glacier ice, and the density of feeder gullies in the rockwall above drift slopes, though appears to be unrelated to the presence or absence of buried ice. The above analyses suggest that both the amount of precipitation and focusing of water delivered onto valley-side drift are instrumental in conditioning gullying intensity, and may determine the presence or absence of widespread paraglacial reworking at sites where the intrinsic constraints on drift reworking are satisfied.

4.6 Timing of paraglacial modification of drift.

Although the processes of paraglacial slope modification have been described in several mountain environments, the factors determining the timing and duration of activity remain largely unexplained. This section considers the temporal dimension of paraglacial modification of valley-side drift at the Norwegian field sites in terms of both the initiation and termination of activity, and assesses possible causes for each.

4.6.1 Initiation of paraglacial activity

Jostedalsbreen

The timing of the onset of paraglacial activity associated with recent retreat of the outlet glaciers of Jostedalsbreen has been established through morphostratigraphic relationships between paraglacial landforms and lichenometrically-dated moraine sequences, on the grounds that accumulation of paraglacial cones and fans cannot have commenced before the terrain they occupy was deglaciated. As Ballantyne (1995a) warned, however, it is possible that the upper parts of some depositional features began to accumulate whilst glacier ice still occupied the ground immediately downslope. Conversely, it is possible that reworking of valley-side drift did not immediately follow deglaciation, implying that the moraine ages are maximal for the onset of paraglacial slope modification.

At Bergsetbreen (Figure 4.2), the apices of valley-side cones 1-6 lie within the limit of the 'Little Ice Age' advance of Bergsetbreen. Cone accumulation at this site must therefore post-date AD 1750-1775, the probable age of the 'Little Ice Age' limit (Bickerton and Matthews, 1993). The lichenometric ages of moraine fragments buried beneath or adjacent to paraglacial debris cones in

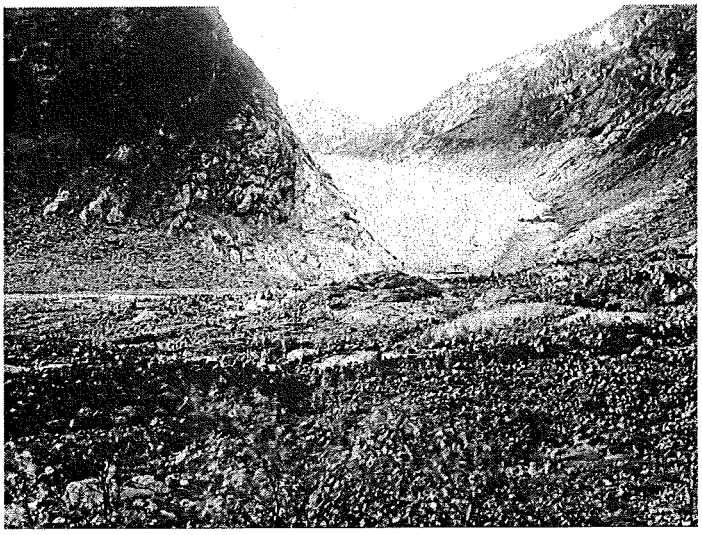
Bergsetdalen suggest that cones 2 and 6 began to accumulate between AD 1750 and AD 1857 and that the onset of accumulation of cone 3 occurred between AD 1857 and AD 1908 (Ballantyne, 1995a). In each case, however, it should be noted that the relationships between the cones and moraines provide only a maximal age estimate for the onset of cone accumulation. Although valley-side drift deposits are likely to be least stable immediately after deglaciation (i.e. prior to vegetation colonisation), it is possible that drift reworking and cone accumulation was delayed for some time after deglaciation.

On the basis of historical evidence cited by Ballantyne and Benn (1994), the innermost moraine ridge in Fåbergstølsdalen records an advance that culminated in AD 1930 (Figure 4.3). The valley floor upvalley from this moraine was therefore deglaciated between AD 1930 and 1996, when the present research programme started. Photographs taken of the field site in 1933 and 1943 (Figures 4.12a and 4.12b) show that despite marked thinning of Fåbergstølsbreen as it retreated from the 1930 moraine ridge, the easternmost valley-side drift slopes experienced little or no gullying prior to 1943, and that the outermost Little Ice Age lateral moraine (now largely obliterated) extended continuously downvalley at that time. These photographs demonstrate that the gully erosion evident in Figures 4.3 and 4.12c in the eastern part of Fåbergstølsdalen postdates AD 1943, and represents at most 53 years (1943-1996) of paraglacial activity. Aerial photographs taken in 1966 reveal the position of the frontal margin of Fåbergstølsbreen roughly midway between the 1930 moraine and the position of the snout in 1996 (Figure 4.3), which suggests that gully erosion above the snout in the western part of Fåbergstølsdalen postdates 1966 and represents no more than 30 years (1966-1996) of activity.

All paraglacial deposits at the Lodalsbre site lie below the extrapolated level of the Little Ice Age glacial maximum, and hence must have begun to

a

b



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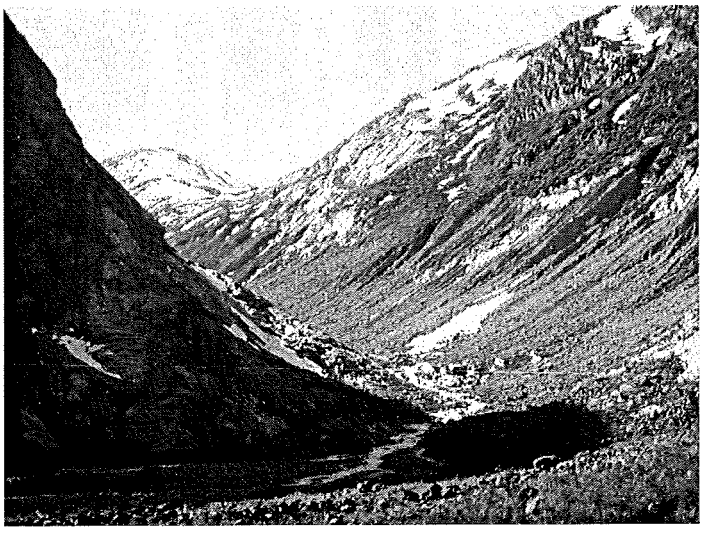


Figure 4.12. Photographs of upper Fåbergstølsdalen in (a) 1933 (K. Fægri), (b) 1943 (K. Fægri) and (c) 1996. Despite the substantial thinning of the glacier 1933-43, these photographs reveal an absence of gully erosion of the drift slopes on the north side of the valley, and show that the outermost 'Little Ice Age' moraine originally extended continuously across the valley side. This evidence demonstrates that paraglacial reworking of valley-side drift postdates 1943.

accumulate after *c.* AD 1826, the lichenometric age obtained for the outermost moraine by Bickerton and Matthews (1993). However, cone deposits in area 1 (Figure 4.4) overlie a moraine ridge dating to *c.* 1837, indicating that the bulk of the paraglacial deposits at this location were not deposited until after that time. Burial of other dated moraine ridges demonstrates the upvalley younging of paraglacial gully initiation associated with the retreat of Lodalsbreen and resulting exposure of steep drift deposits. The apices of cones in area 2 lie only a short distance downslope from a glacier limit dated to *c.* 1904, implying that cone accumulation could have commenced shortly after that date. However, the position of moraines dated to *c.* 1938 at their distal end demonstrates that area 2 did not become completely ice-free until then, implying that the onset of debris accumulation in this zone probably began between *c.* 1904 and *c.* 1938. Upvalley extrapolation of the ice surface from a moraine deposited in *c.* 1957 and from the 1976 snout position suggests that the debris cones in area 3 could not have accumulated until after 1957, and that those in areas 4 and 5 only began accumulating since 1976. Thus modification of drift near the 1996 terminus of Lodalsbreen has occurred within a period of 20 years (1976-1996).

Dated moraines marking the retreat of Nigardsbreen from its Little Ice Age limit demonstrate that all reworked drift within the field site lies below the level of the Little Ice Age glacial maximum, and must therefore have accumulated after *c.* AD 1745 (Matthews *et al.*, 1986; Bickerton and Matthews, 1992). Aerial photograph and morphostratigraphic evidence show that the onset of cone accumulation in the eastern part of the site (Figure 4.5) postdates *c.* AD 1942, and reconstructions of the surface profile of Nigardsbreen based on terrestrial photogrammetry demonstrate that area 2 (in the western part of the foreland) was still buried beneath ice in AD 1937 (Østrem, 1988). Both cone 1 and the gullying of drift in area 2 are clearly visible in a photograph taken of the Nigardsbreen foreland in AD 1954. It can be assumed that reworking of drift in the eastern part

of this foreland represents no more than 54 years (1942-1996) of paraglacial activity.

Jotunheimen

The Leirdal and Visdal trunk valleys (Figures 4.6 and 4.7) were last occupied by glacier ice in Late Weichselian times (Vorren, 1977), prior to widespread deglaciation at *c.* 9 ka BP in the Preboreal chronozone (*cf.* Vorren, 1973). As is discussed further in section 6.4.2, reworked glacigenic deposits have been identified on valley-side slopes in Leirdalen, immediately overlying *in situ* till, and are inferred to represent paraglacial reworking of hillslope drift during or shortly after Preboreal deglaciation, prior to vegetation colonisation of the slope surface. The initial emplacement of some or all of the slope-foot debris cones and fans in the trunk valleys at both Leirdalen and Visdalen sites may therefore date to this period of paraglacial activity. Morphostratigraphic relationships between paraglacial landforms and dated moraine sequences within tributary glacier (Little Ice Age) forelands, however, demonstrate much more recent drift-slope reworking in these areas. The timing of the onset of drift slope reworking within such forelands can only be constrained temporally with reference to the estimated ages of the outermost Little Ice Age moraines. McCarroll (personal communication, 1996) considers that the age of *c.* AD 1750 estimated by Matthews (1974, 1977a) for the outermost moraine at Storbreen corresponds approximately to the age of other moraines representing Little Ice Age glacial maxima around Leirdalen and Visdalen. Accordingly, it is inferred that the onset of paraglacial resedimentation of valley-side drift in the forelands of nordre Illåbreen, søre Illåbreen and Heillstugubreen postdates *c.* 1750. On the Heillstugubre foreland a large, predominantly lichen-free moraine ridge lying *c.* 700 m within the Little Ice Age limit may correspond to the regional readvance dated at *c.* 1920-1930 (McCarroll, personal communication, 1996). If so, the gullying immediately west of the

present snout of Heillstugubreen must postdate *c.* 1920 and represents no more than about 75 years of erosion.

Possible causes

As stated by Church and Ryder (1989), what distinguishes paraglacial modification of steep drift slopes from non-paraglacial activity is not the process of sediment reworking, but the 'direct glacial conditioning' of such processes (Church and Ryder, 1972). Thus the mechanisms triggering the onset of paraglacial reworking are no different from those in other environments where 'glacial conditioning' does not apply. Models of slope failure on unconsolidated sediments (e.g. Chandler, 1977) indicate that failure and consequent flow of sediment is usually attributable to a rise in pore-water pressure causing a reduction in shearing resistance.

The onset of paraglacial reworking of valley-side drift slopes during or immediately after deglaciation is favoured by the abundance of freshly-exposed unconsolidated sediment on steep, unvegetated slopes, and the supply of meltwater (Ryder, 1971a). It is likely that such conditions favoured reworking of recently-exposed drift by debris flows, avalanches or valley-side streams at the Norwegian field sites investigated. At Fåbergstølsbreen, however, extensive paraglacial modification of drift by gullying was evidently delayed by at least 13 years after the retreat of the glacier from the 1930 moraine (Figure 4.12b). It is possible that a similar delay in hillslope response to deglaciation may have also occurred at other sites. The delay in drift-slope erosion at Fåbergstølsbreen (and possibly elsewhere) seems to imply that though deglaciation exposed slopes in a state of critical stability (metastability), subsequent failure may have been triggered by some specific destabilising event.

This could have happened in a number of ways. First, initial widespread paraglacial debris flow activity may have been triggered by a high-magnitude rainstorm (*cf.* Rapp and Nyberg, 1981; Larsson, 1982; Watanabe, 1985; Zimmermann and Haeberli, 1992). Such a storm caused widespread erosion of valley-side drift in the Jostedalen area in the summer of 1979 (Watson, personal communication, 1997). Subsequent debris flows may be triggered by lower-magnitude rainstorm events, as gully development tends to be self-enhancing following initial channel incision. Alternatively, triggering of initial failure by build-up of high pore-water pressures could have resulted from exceptionally rapid snowmelt (Ballantyne and Benn, 1994). Secondly, melt-out of ice buried within steep drift accumulations may progressively destabilise the sediment until pore-water pressure is sufficiently high for delayed failure to occur (Harris and Gustafson, 1993), though the analysis of the association of buried ice and the distribution of paraglacial reworking (section 4.5.2 above) suggests that this is unlikely to have been important at the sites investigated. Thirdly, physically-based modelling of the role of climate, vegetation and pedogenesis in affecting slope stability has demonstrated that delayed failure of particular regoliths may partly reflect a gradual decline in slope stability due to progressive pedogenesis (Brooks *et al.,* 1993b, 1995; Brooks and Richards, 1993, 1994; Brooks, 1997), though this seems unlikely to be relevant in the case of drift slopes exposed over a period of only a few decades. Fourthly, earthquakes are known to have triggered drift-slope failures in Norway and elsewhere (e.g. Solonenko, 1963; Grove, 1985; Kotarba, 1992; Lägerback, 1992). It is likely that enhanced seismicity was more prevalent during the period of ice-sheet deglaciation than the Late Holocene, due to differential glacio-isostatic uplift of crustal blocks along faults, and may have contributed to the onset of Preboreal paraglacial activity. However, from consultation of records spanning the period AD 1497-1975, Husebye *et al.* (1978) identified a surge in seismic energy release in Norway between AD 1863 and 1913, which suggests that seismic activity cannot be discounted as a possible

trigger of failure on steep drift slopes exposed by the retreat of glaciers from their Little Ice Age maxima.

4.6.2 Termination of paraglacial activity

Jostedalsbreen

Although limited accumulation (particularly of avalanche debris) occurs at present on cones 1, 2, 3, 5 and 6 in Bergsetdalen (Figure 4.2), sources of readily entrainable drift above each of these cones are largely exhausted, and hence the main phase of paraglacial activity may be regarded as over (Ballantyne, 1995a). Moreover, all of these debris cones support a widespread vegetation cover, indicating that sediment accumulation on cone surfaces is now limited. Ballantyne (1995a) estimated the effective termination of sediment accumulation on cones 2, 3 and 6 by dendrochronological analyses of mature birch and willow growing on cone surfaces. Because of uncertainty in assuming (i) that tree colonisation could only proceed after the cone surfaces had stabilised, and (ii) that colonisation occurred soon after stabilisation, age estimates based on the ages of tree rings must be regarded as minimal. Nevertheless, his results suggested that sediment accumulation had largely ceased over much of the surfaces of these cones by AD 1965 at the latest. On the most recently deglaciated part of the Fåbergstølsbre foreland, however, paraglacial activity is clearly still active, with debris flows and avalanches occurring annually (Ballantyne and Benn, 1994), numerous recent debris flow lobes on the valley floor and an abundant reservoir of drift on upper slopes available for future reworking (Figures 4.9 and 4.10). In the Lodalsbre foreland, much of the valley-side drift farthest from the glacier terminus has stabilised and supports a closed vegetation cover, but more recently-exposed drifts, like those in upper Fåbergstølsdalen, are clearly subject to current reworking. Similarly, the older parts of the Nigardsbre foreland are vegetated and

have the appearance of stability. In general, therefore, although the most recently exposed terrain on the Jostedalsbre forelands exhibits evidence for abundant recent drift reworking on steep slopes, older terrain often supports an extensive or complete vegetation cover, and shows little sign of recent drift reworking, except locally by avalanches. Collectively, these sites seem to suggest that the initial phase of paraglacial slope modification and drift reworking on steep valley-side slopes lasts for a timescale of several decades, and no more than a century or two.

Jotunheimen

In the trunk valleys of Leirdalen and Visdalen, recently-active valley-side gullies and debris cones are superimposed on relict gullies and cones of variable size and possibly age. Lichenometric and radiocarbon ages have been published for a number of active and relict cone surfaces in Leirdalen (Innes, 1982; Matthews *et al.*, 1997) and are considered in section 4.6.3. The continued resedimentation of Preboreal-age glacigenic material demonstrates that paraglacial activity has not yet terminated in these valleys. Where paraglacial reworking of valley-side drift has occurred in the recently-deglaciated tributary forelands in Jotunheimen, there is also no evidence for termination.

Possible causes of termination

Although the active life of paraglacial gullies and associated debris cones will inevitably vary according to conditioning factors, the localised termination of paraglacial activity within some of the Norwegian sites appears to have been caused primarily by exhaustion of sediment in the source areas upslope, though slope stabilisation may also reflect other causes. The relict cones investigated at the Jostedalsbre and Jotunheim sites are all characterised by a depletion of drift upslope, frequently indicated by widespread exposure of bedrock in gully heads

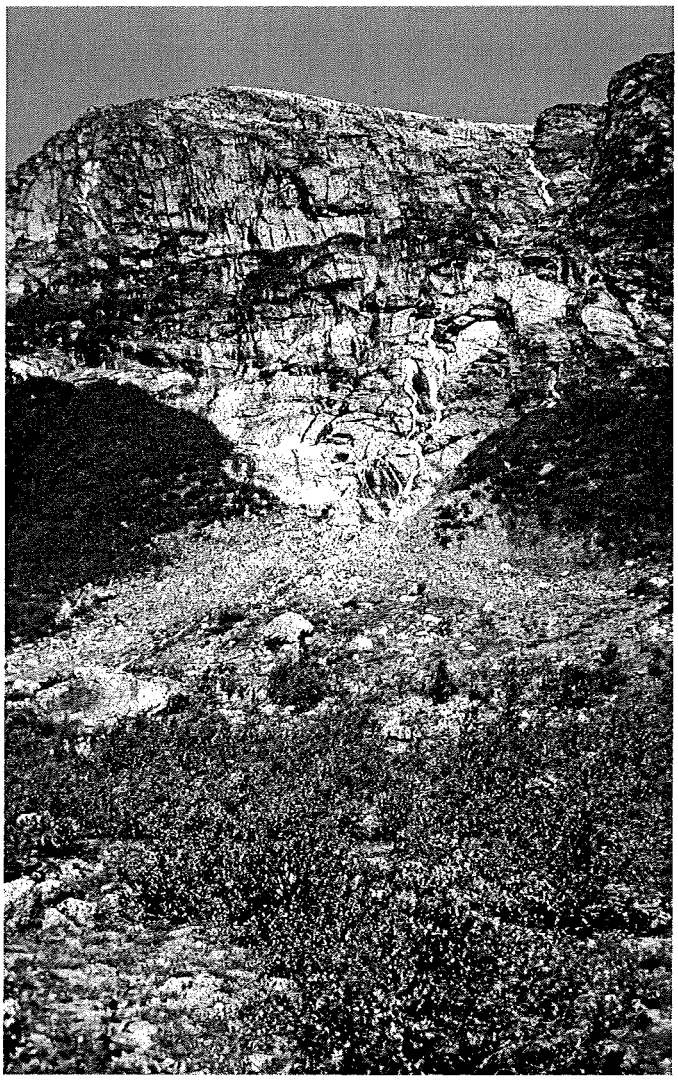


Figure 4.13. Largely-relict debris cone 2 in the Bergsetbre foreland, Norway, showing the area of bedrock from which drift has been stripped and is now sediment-exhausted.

and floors (Figure 4.13). Moreover, the dating control on both initiation and termination of paraglacial activity in Bergsetdalen has demonstrated that sediment exhaustion at that site occurred within a few decades or centuries of deglaciation (Ballantyne, 1995a). In addition to the depletion of sediment supply, it is probable that paraglacial drift reworking at some sites may have locally ceased as vegetation colonisation stabilised valley-side drift slopes, though this represents something of a 'chicken and egg' problem: although vegetation colonisation may

increase slope stability, slope stability also favours vegetation colonisation. A further likely control on termination of drift reworking, at least by debris flows, is gradient (Ballantyne and Benn, 1996). Whilst a minimum slope gradient seemingly limits the initiation of hillslope debris flows, upslope removal of drift and resedimentation downslope may progressively lower the overall slope gradient below the threshold of flow initiation (see section 7.3.1 below). If this is so, preferential termination of paraglacial debris flow activity in Bergsetdalen, lower Stordalen and the trunk valleys of Leirdalen and Visdalen may reflect the slightly gentler valley-side gradients at these sites compared with those of the Fåbergstølsbre and upper Lodalsbre forelands.

4.6.3 Delayed or renewed paraglacial activity

In both the Leirdal and Visdal trunk valleys, the picture of a pulse of paraglacial activity immediately following regional deglaciation is complicated by evidence for phases of reworking of valley-side drift deposits within the past few millennia (*cf.* Innes, 1982; Matthews *et al.*, 1997). This suggests that the stability attained after an initial episode of rapid drift reworking was no more than conditional, and vulnerable to changes in both intrinsic conditions (shearing resistance) or extrinsic conditions (such as climate change, vegetation change or specific, high-magnitude destabilising events). Matthews *et al.,* (1997) have interpreted radiocarbon-dated deposits on the valley floor in Leirdalen as representing the onset of renewed debris flow activity at *c.* 7.5 ka BP and an apparent increase in debris flow frequency after *c.* 3.8 ka BP. They also suggested a further increase in debris flow activity at this site during recent centuries, possibly associated with Little Ice Age climatic deterioration. Evidence for an apparent recent acceleration in debris cone aggradation in Leirdalen was also detected by Innes (1982) through lichenometric dating of superficial debris flow deposits. Innes suggested that debris began to reaccumulate on the sampled

cones at *c.* AD 1640, though his data must be treated with caution as older flow deposits may be buried by more recent ones, so that only the latter were sampled.

A point of interest here is that recent paraglacial activity acting on the gullies and fans of possible Preboreal age in Leirdalen appears to differ in magnitude from the initial paraglacial processes that created these landforms. Not only are recent debris flow lobes small in comparison with the volume of reworked sediment implied by the ancient paraglacial cones, but the calibre of sediment also differs. The mean *a*- axis length of 100 clasts removed from two fresh debris flow deposits is 10.9 cm, in contrast to a mean of 28.5 cm from 100 clasts measured on two relict cones, a difference significant at $p < 0.01$ (Mann-Whitney two sample test). This suggests that more recent debris flows have been of lower magnitude and competence in comparison with the reworking events that built the cones.

4.7 Summary.

This chapter has addressed questions concerning the extent, nature, constraints and timing of paraglacial activity in two areas of southern Norway. Whilst caution should be exercised in applying the results presented above to other environments, paraglacial activity does appear to be the 'normal' geomorphic response to deglaciation where favourable intrinsic controls and extrinsic triggers exist. In sum, four principal findings are identified; these are outlined below.

1. At the four recently-deglaciated sites investigated around Jostedalsbreen, gully density is in the order of *c.* 10-100 gullies per kilometre, reflecting resedimentation of an estimated 44% (± 38% at 95% confidence) of all valley-side drift under propitious circumstances. Drift slopes at the sites in the Jotunheim massif are generally much less extensively modified by

paraglacial processes (gully density nowhere exceeds 8 km^{-1} in either site), and no more than *c.* 11% of hillslope drift appears to have been resedimented *via* gully erosion.

2. The overall preponderance of landform assemblages comprising paired levées, thick terminal lobes and poorly sorted deposits on lower slopes downslope of gullies indicates that debris flow is the dominant process of paraglacial sediment transfer. However, flattened vegetation, perched clasts and debris tails below highly concave valley-side slopes locally testify to the operation of snow avalanches as a secondary paraglacial process, particularly in the foreland valleys investigated in Jotunheimen.

3. Factors controlling paraglacial drift modification by debris flows include gradient, sediment availability and water supply. At the Norwegian field sites, gully density generally exceeds 20 gullies per kilometre where drift is steeper than *c.* 30° and thicker than *c.* 10 m, and where the void ratio of unreworked sediment exceeds *c.* 0.35. Widespread gullying is also favoured at sites of high and focused water input, in particular where melting snow and ice are involved.

4. The paraglacial landforms observed around Jostedalsbreen and in Jotunheimen represent different stages of paraglacial activity. Ancient debris cones in Leirdalen and Visdalen probably reflect paraglacial reworking of drift deposits during or shortly after deglaciation in Preboreal times. On recently-deglaciated terrain paraglacial deposits have accumulated very rapidly following recent deglaciation, often within decades of exposure to nonglacial processes. Although sediment remobilisation continues to the present at most of the Jostedalsbre sites, debris cones in Bergsetdalen have stabilised within *c.* 200 years of ice

retreat through exhaustion of sediment supply. In Leirdalen and Visdalen there is evidence for phases of paraglacial reworking of valley-side drift long after initial cone formation.

Chapter 5

Delayed or renewed paraglacial modification of drift slopes in Scotland

5.1 Introduction.

The legacy of paraglacial drift slope adjustment is evident not only in presently-glacierized environments such as Norway, but also in mid-latitude mountain environments that were deglaciated at the end of the Pleistocene epoch. One such area is the Highlands of Scotland, where many Late Pleistocene paraglacial landforms flank the sides of deglaciated valleys (e.g. Peacock, 1986; Brazier *et al.*, 1988; Auton, 1990; Benn, 1990, 1991). In Scotland, however, there is evidence not only for rapid reworking of hillslope drift deposits immediately after deglaciation, but also for delayed or renewed reworking of valley-side drift deposits within the past few centuries. As part of a growing body of literature focusing on accelerated rates of landscape change in upland Britain during the latter half of the Holocene (Ballantyne, 1991b), many authors have drawn attention to modification of drift slopes in the Scottish Highlands by debris flows (e.g. Strachan, 1976; Innes, 1983b, 1985a; Ballantyne and Eckford, 1984; Ballantyne, 1986a; Brazier *et al.,* 1988; Brazier and Ballantyne, 1989; Ballantyne and Benn, 1996; Hinchliffe *et al.,* 1998) and surface wash (e.g. Tipping, 1995). Nevertheless, surprisingly little has been unequivocally established concerning the extent, precise timing and causes of this delayed or renewed slope activity, the issues that form the principal concerns of this chapter. After a description of field and analytical methods employed (5.2), the chapter assesses the extent (5.3), processes (5.4), constraints (5.5) and timing and possible causes (5.6) of delayed or renewed paraglacial modification of drift at the four Scottish field sites. Principal results are summarised in the concluding section (5.7).

5.2 Methods.

As with the field sites in Norway, investigation of the characteristics of delayed or renewed paraglacial hillslope modification at the Scottish field sites involved (1) geomorphological mapping of the distribution and morphology of drift slopes, (2) instrumental survey of slope form, (3) measurement of the dimensions of gullies incised into hillslope drift, and (4) analysis of the sedimentological characteristics of *in situ* glacigenic drift deposits. For field mapping purposes 1:5,000 scale base maps were produced from Ordnance Survey 1:25,000 and 1:10,000 sheets. Consultation of ground and aerial photographs of each of the four field sites aided accurate mapping of gullies and other features. A key to all geomorphological maps is provided in Figure 4.1.

A number of the terms and methods employed below have been defined above in section 4.2. At the Scottish sites gully density was calculated as a mean frequency per kilometre of slope for the entire field site. In the Western Red Hills, Glen Einich and the Pass of Drumochter, slope profile measurements and other survey data were obtained using a Wild T1000 EDM (Electronic Digital Mapping) unit, and slopes in Glen Docherty were surveyed using an abney level, ranging rods and a 30 m tape. Gully dimensions were also surveyed at these four sites to determine drift thickness, the volume of remobilised sediment and mean gully volume, expressed within 95% confidence limits (Table 5.1). To assess the possible role of sediment characteristics on the distribution of delayed or renewed paraglacial slope reworking, laboratory granulometric analyses were conducted on fine-grained (< 2 mm) sediment removed from valley-side drift deposits, and matrix void ratios were calculated as a measure of sediment packing.

Mapping of the Scottish field sites revealed natural sections in drift deposits both along the sides of gullies and within debris cones. Depositional

facies underlying drift slopes at the field sites locally include organic-rich layers, many of which appear to represent *in situ* buried palaeosols. Sections through reworked drift were photographed and logged to record the precise stratigraphic position of such buried palaeosols prior to sampling for radiocarbon assay and analysis of constituent palynomorphs. The top and bottom 3-5 mm of selected *in situ* buried palaeosols were sampled for radiocarbon assay. All samples were collected by removing overlying or underlying deposits so that the surface of the organic horizon was exposed. Each layer was then sampled using a fresh razor, sealed in aluminium foil, labelled and placed in an air-tight polythene sampling bag for cold storage. Samples were submitted to the NERC Radiocarbon Laboratory at East Kilbride for either conventional ^{14}C radiometric dating, or pre-treatment and forwarding to the NSF-AMS facility at the University of Arizona where aliquots of prepared gas underwent accelerator-beam measurement. Radiocarbon ages were transformed to calendar years using the calibration programme CALIB rev. 3.0 of Stuiver and Reimer (1993) and the calibration data of Stuiver and Becker (1993). Calendar age ranges are quoted in the text as 95% confidence limits ($\pm 2\sigma$). For convenience of comparison, previously published radiocarbon dates have also been converted to calendar age ranges using the CALIB rev. 3.0 programme.

Buried palaeosols were also sampled for analysis of constituent sub-fossil pollen using 10 cm and 30 cm stainless steel monolith tins, and sealed in clearly labelled, air-tight bags for storage at 4°C. Samples were measured for 0.5 cm^3 volume by displacement, and tablets of *Lycopodium* were added to allow pollen and charcoal concentration values to be calculated. Pollen preparation followed the standard HCl, NaOH, HF and acetolysis treatment (Fægri and Iversen, 1989). The resulting pollen samples were mounted unstained in silicone oil, and routine pollen identification and counting were undertaken at a magnification of x600. The basic counting sum was a minimum of 300 total land pollen (TLP) grains, all

of which were recorded against categories of preservation status. Pollen type nomenclature follows Stace (1991) and Bennett *et al.*, (1994). Microscopic charcoal concentrations are expressed in values of $cm^2\ cm^{-3}$, and the charcoal to pollen ratio was also determined in order to overcome problems of false charcoal peaks caused by differential rates of sediment accumulation (Swain, 1973; Patterson *et al.*, 1987). Selected palynomorph data are illustrated as a percentage of total land pollen. Occasional reference is also made to calculated pollen concentrations to verify peaks in the pollen percentage totals of individual taxa. Pollen diagrams were produced using the computer programmes TILIA and TILIA·GRAPH (Grimm, 1991). For convenience of representation, calculated pollen percentages of < 2% for individual taxa at each level are indicated by crosses, the number of which signifies the actual number of grains encountered. To assess variations in the organic content and nature of the minerogenic component of the buried palaeosols, sub-samples were dried, weighed, combusted at 550° for 4 hours and re-weighed to calculate the percentage weight loss on ignition, and the residue subjected to particle size analysis using a Coulter LS100 laser granulometer.

5.3 Extent of delayed or renewed paraglacial modification of drift.

Although Late Holocene modification of drift-covered hillslopes has been described at several sites in Scotland (e.g. Innes, 1983b; Brazier *et al.,* 1988; Brazier and Ballantyne, 1989; Ballantyne and Benn, 1996), little attempt has hitherto been made to assess the extent of delayed or renewed paraglacial reworking of valley-side sediment. Innes (1983b) measured thirty individual recent (< *c.* 500 years old) debris flow deposits from sites in the Scottish Highlands, and found that these vary in volume from *c.* 5 m^3 to 300 m^3, and volume calculations for three debris cones in Glen Feshie (western Cairngorms)

Profile	Maximum depth (m)	Length (m)	Mean width (m)	Volume (10^3 m^3)
Western Red Hills				
Wa	3.3	152	7.5	2.51
Wb	4.8	67	14.6	1.30
Wc	6.8	247	10.4	4.41
Wd	8.8	104	22.2	11.2
We	4.2	250	9.2	3.66
Wf	9.8	156	17.0	7.82
Mean	*6.3*	*163*	*13.5*	*5.2±3.3*
Glen Docherty				
Da	10.4	373	23.5	38.9
Db	4.0	340	20.1	14.3
Dc	10.0	331	26.4	41.4
De	7.0	244	17.7	17.0
Df	5.6	80	19.2	5.21
Mean	*6.2*	*274*	*21.4*	*23.4±16.0*
Glen Einich				
Ea	13.2	305	34.9	121
Eb	6.4	149	10.9	1.92
Ec	4.7	197	6.9	1.74
Ed	5.9	66	10.7	0.45
Ee	6.6	148	10.9	1.47
Mean	*7.4*	*173*	*14.9*	*25.3±53.5*
Pass of Drumochter				
PDa	8.3	118	23.3	8.18
PDb	5.5	191	12.0	4.86
PDc	6.7	118	25.6	6.96
PDd	6.0	106	32.3	9.84
Mean	*6.6*	*133*	*23.3*	*7.46±2.42*

Table 5.1. Dimensions of gully systems surveyed at field sites in the Northwest Highlands and the Grampian Highlands. Mean volumes are expressed with 95% confidence limits.

yielded values of 22,300-23,300 m^3, 2400-3400 m^3 and 31,800-35,600 m^3 (Brazier, 1987). The bulk of these cones has been shown to have accumulated through reworking of glacial drift upslope over the last 400 years, though these

cone sediments rest on earlier debris flow deposits that predate *c.* 2000 yr BP (Brazier and Ballantyne, 1989). In this section the distribution of delayed or renewed paraglacial resedimentation of valley-side drift deposits is considered for each of the field sites in the Scottish Highlands to establish the extent of drift reworking, and whether this is a common or infrequent attribute of drift-mantled slopes in this area.

5.3.1 Northwest Highlands

The distribution of landforms indicative of delayed or renewed paraglacial modification of steep hillslope drift at the two field sites in the Northwest Highlands is shown in Figures 5.1 to 5.4, and quantitative indices summarising the extent of delayed or renewed paraglacial sediment reworking at these sites are presented in Table 5.2. To help identify patterns in the extent of delayed or renewed paraglacial activity, both sites have been divided into two areas. The limits of areas 1 and 2 in the Western Red Hills site (Figures 5.2 and 5.3) are defined in Figure 5.1. The uppermost slopes of the Western Red Hills comprise a veneer of frost-shattered diamictic regolith interrupted only locally by bedrock outcrops. Below *c.* 350-400 m, the slopes are gullied, vegetated and predominantly drift-mantled, though a shallow cover of granitic scree frequently extends some distance downslope onto the glacial drift. On the lower drift slopes numerous debris cones extend downslope from gullies and debris tracks. Some are completely vegetation-covered, but others support covers of unvegetated debris indicative of recent activity. Survey data for six gullies indicate that gully volumes range from *c.* 1,300 m^3 to *c.* 11,200 m^3, with a mean value of 5,150 m^3. For the whole field site, the extent of delayed or renewed reworking is represented by a mean gully density of 21 gullies per kilometre of valley-side slope, and an estimated 6.5% (± 4% at 95% confidence) of valley-side drift has been remobilised downslope (Table 5.2). However, resedimentation of drift is most

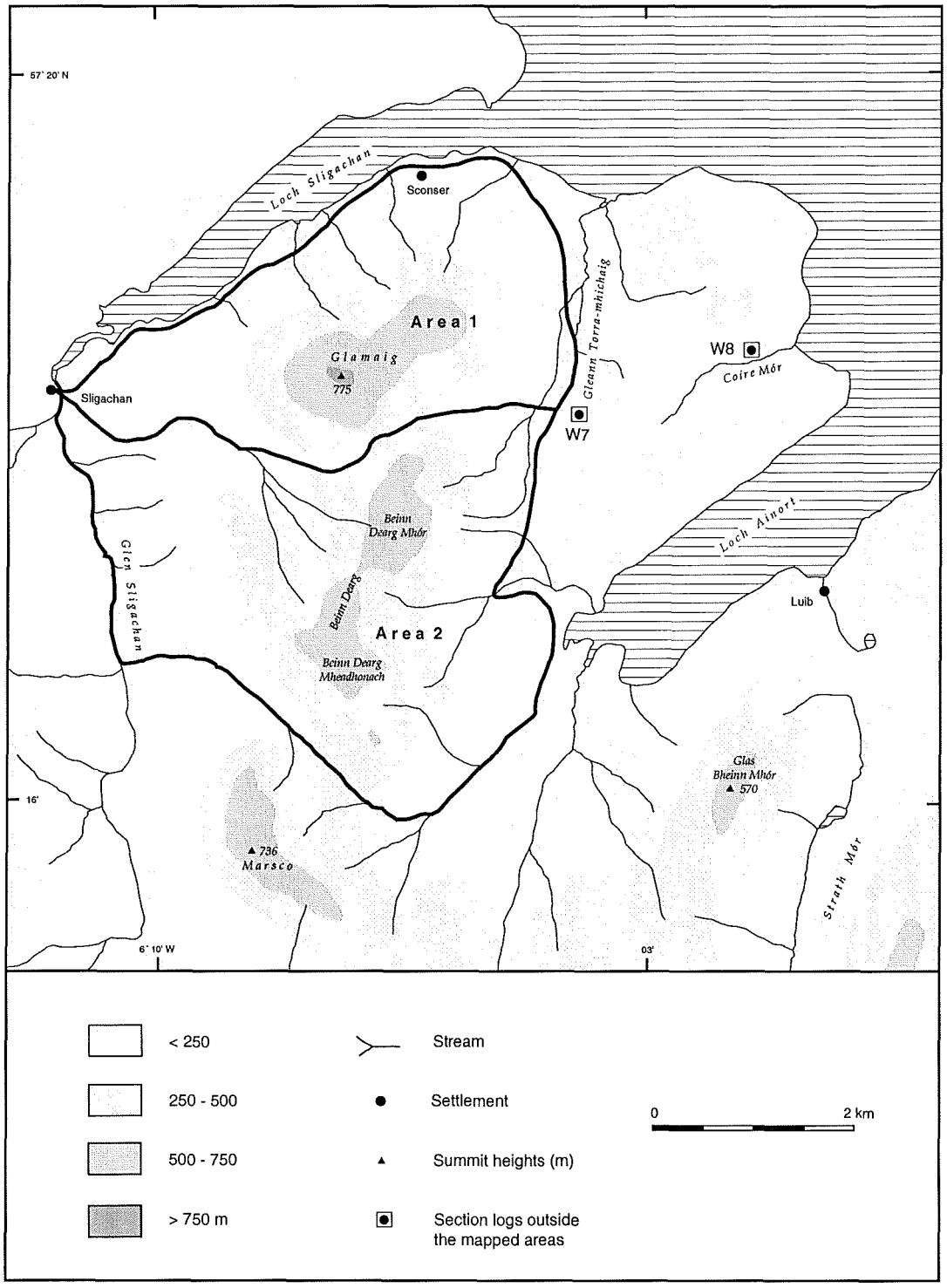


Figure 5.1. The Western Red Hills field site, Scotland showing location of Areas 1 and 2, and sections W7 and W8 outside the mapped area.

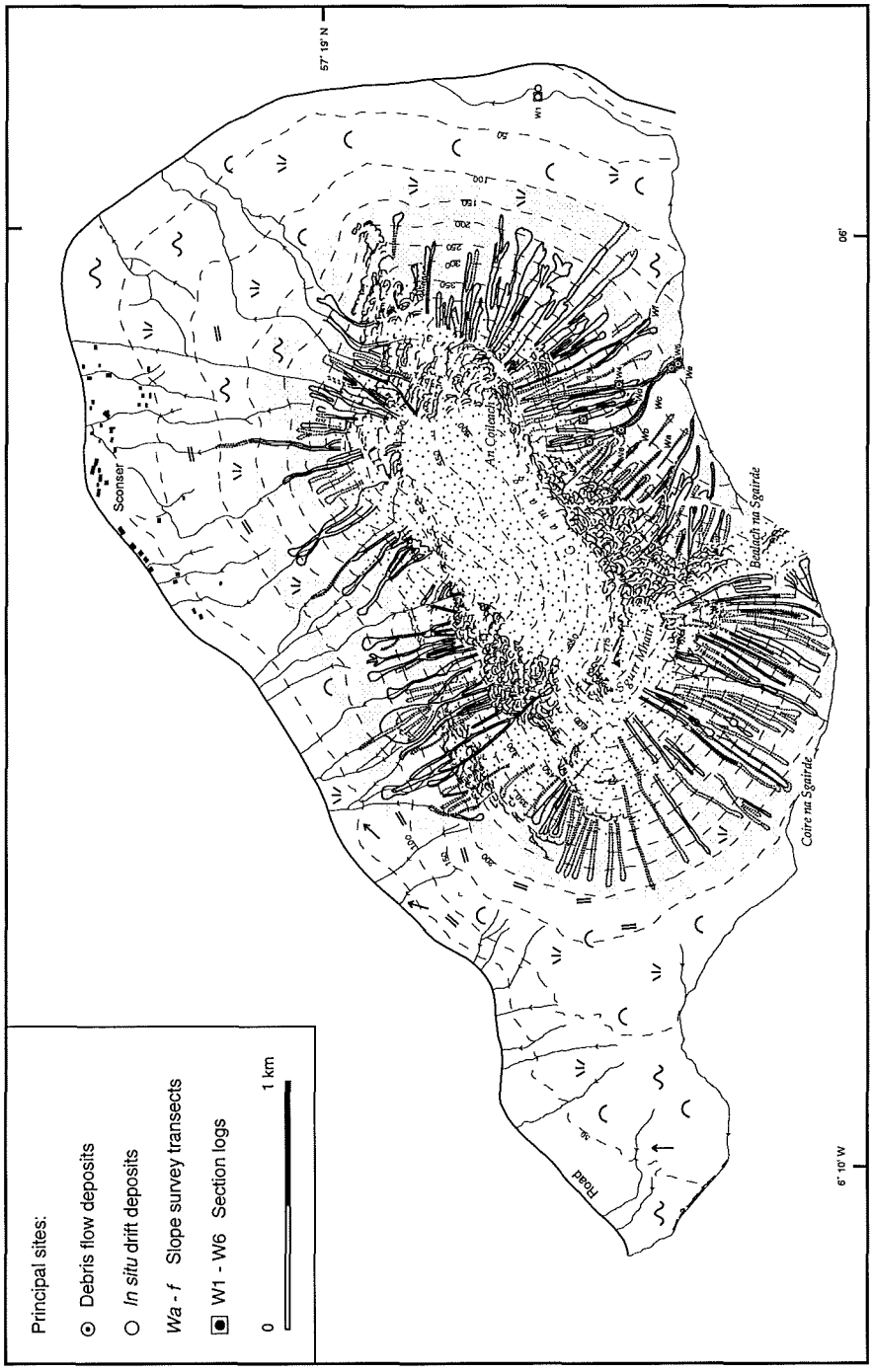


Figure 5.2. Delayed or renewed paraglacial activity in the Western Red Hills (Area 1), Scotland. Key in Figure 4.1.

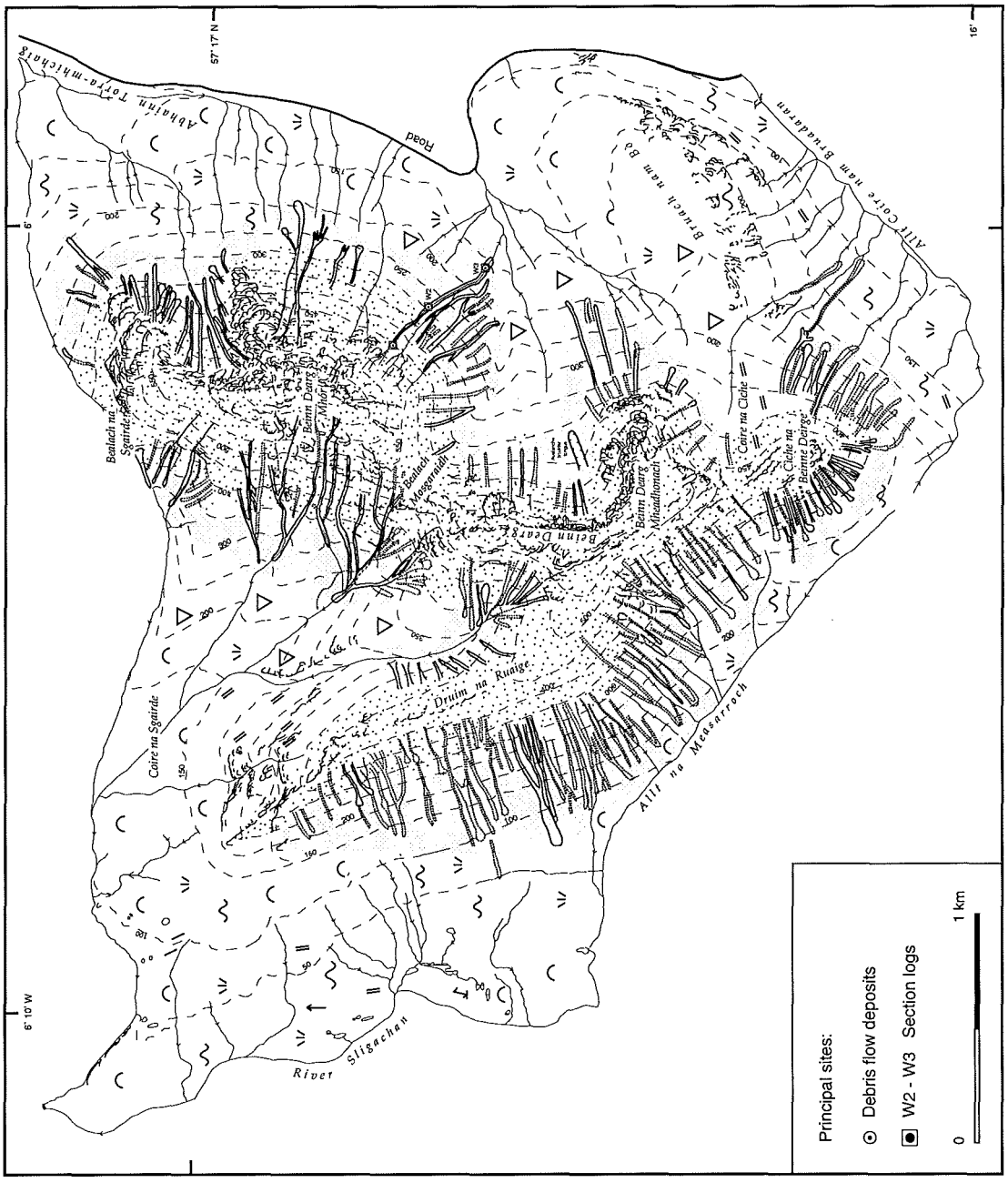


Figure 5.3. Delayed or renewed paraglacial activity in the Western Red Hills (Area 2), Scotland. Key in Figure 4.1.

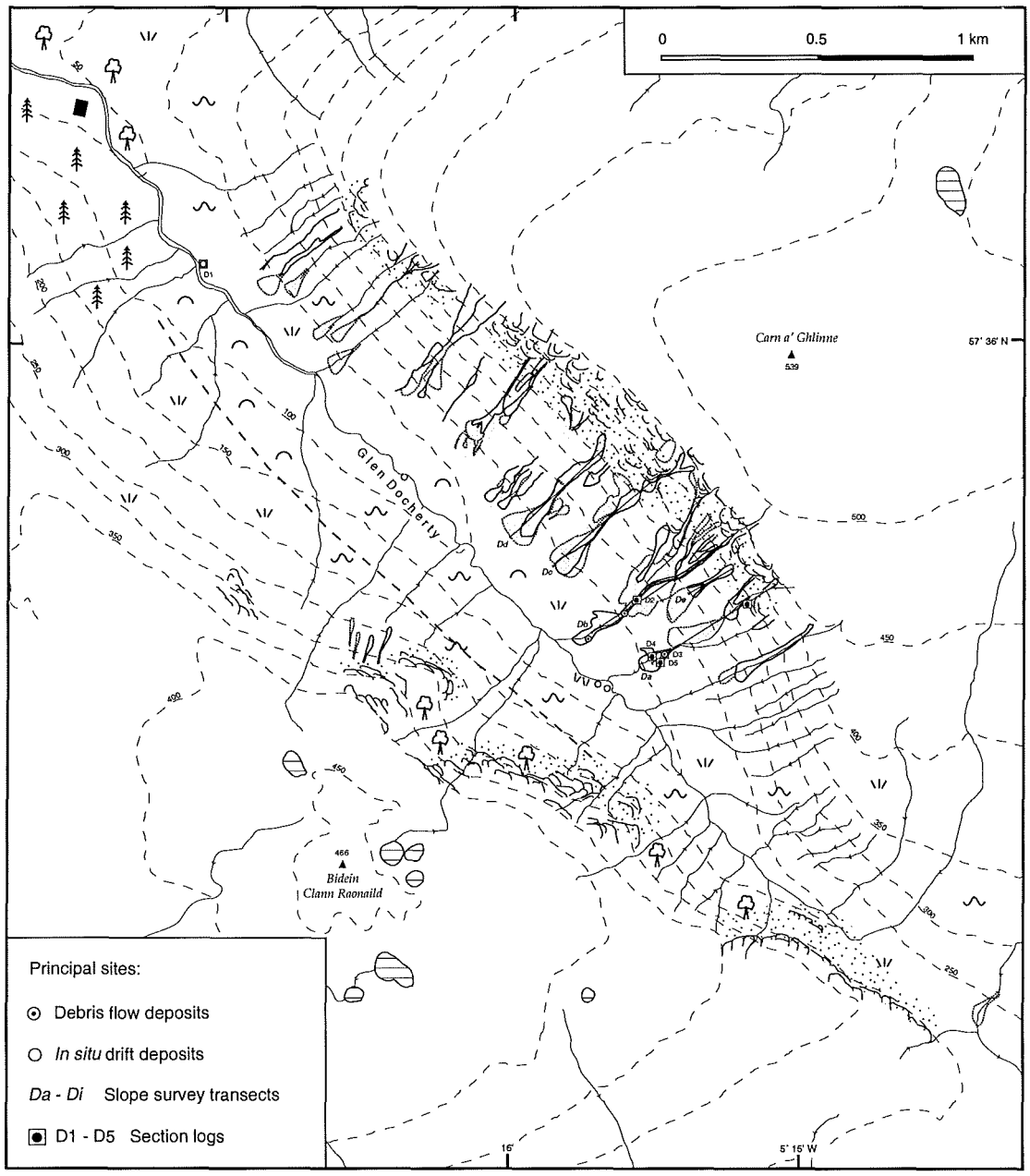


Figure 5.4. Delayed or renewed paraglacial activity in Glen Docherty, Scotland. Key in Figure 4.1.

extensive on the slopes of Glamaig (Figure 5.2 and 5.5), where gully density is 25 gullies per kilometre, and approximately 8% (± 5% at 95% confidence) of valley-side drift has been reworked.

Field site	Gully density (n km^{-1})	% reworked drift*	Mean gully volume (10^3 m^3)*
Western Red Hills	21	6.5±4.2	5.2±3.3
Area 1	25	7.6±4.9	
Area 2	20	5.9±3.8	
Glen Docherty	3.0	3.0±1.3	23.4±16.0
Area 1	5.4	5.4±2.4	
Area 2	0.6	0.22±0.22	

Table 5.2. Extent of delayed or renewed paraglacial modification in the Northwest Highlands field sites, Scotland. * 95% confidence limits.

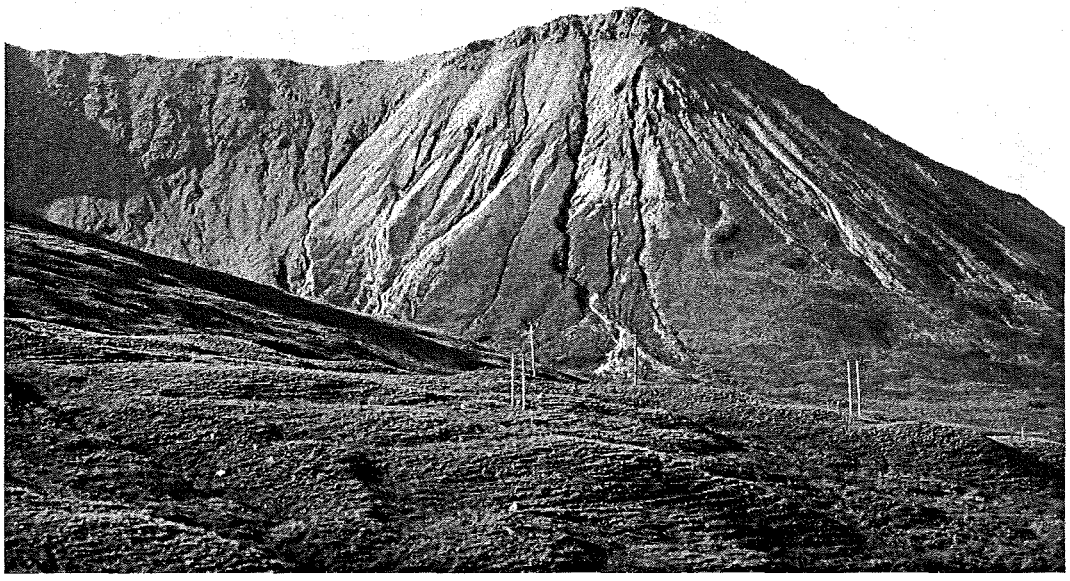


Figure 5.5. Extensive renewed paraglacial resedimentation of drift on Glamaig, Western Red Hills. Notice the widespread gully erosion of upper slopes and deposition of reworked debris on cone surfaces downslope.

Investigation of the characteristics of delayed or renewed paraglacial modification in Glen Docherty was confined to the upper part of the glen overlooked by Carn a' Ghlinne and Bidein Clann Raonaild, where areas 1 and 2 represent the north and south sides of the glen respectively (Figure 5.4). Both valley-side slopes support vegetated glacigenic drift, though bedrock appears extensively at the crest of the south-facing slope. The valley-floor deposits in Glen Docherty have locally been fluvially-incised, creating steep, gullied drift exposures adjacent to the river. Gullies cut into hillslope drift are most numerous on the north side of the glen beneath the broken headwall (area 1), and lead downslope onto cone and fan surfaces on the valley floor. Gully density in area 1 is 5.4 gullies per kilometre, but just 0.6 gullies per kilometre in area 2 (Table 5.2). Although the mean gully density for all Glen Docherty is much lower (3 gullies per kilometre) than that for the Western Red Hills (21 per kilometre), most of the gullies in Glen Docherty are considerably longer and wider than those measured on Skye. Survey of five gullies on the north side of Glen Docherty yielded estimates of sediment removed of *c.* 5,210 m^3 to *c.* 41,360m^3, with a mean of 23,360 m^3 (Table 5.1). The estimated proportion of valley-side drift that has been reworked on the north side of Glen Docherty (5.4% (± 2.4% at 95% confidence) is consequently similar to that estimated for the sites on Skye (6.5 (±4.2 at 95% confidence)), though because of the small proportion of drift reworked on the south side of the glen, the overall percentage reworked drift in all Glen Docherty is markedly less (3% (± 1.3% at 95% confidence)).

5.3.2 Grampian Highlands

The distribution of landforms indicative of delayed or renewed paraglacial modification of steep hillslope drift at the Grampian Highlands field sites is illustrated in Figures 5.6 and 5.7, and summary indices of the extent of such activity are presented in Table 5.3. Both of the field sites in the Grampian

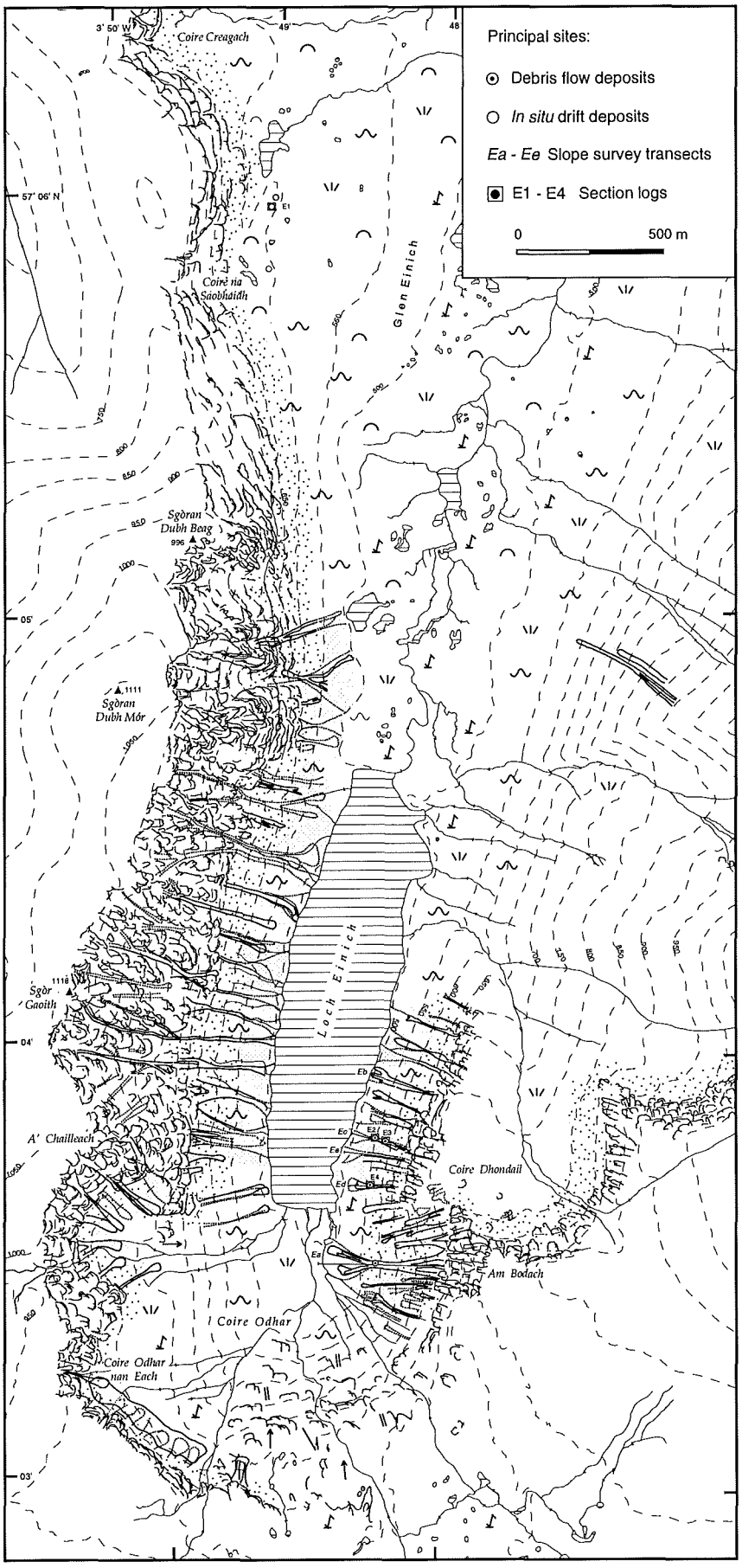


Figure 5.6. Delayed or renewed paraglacial activity in Glen Einich, Scotland. Key in Figure 4.1.

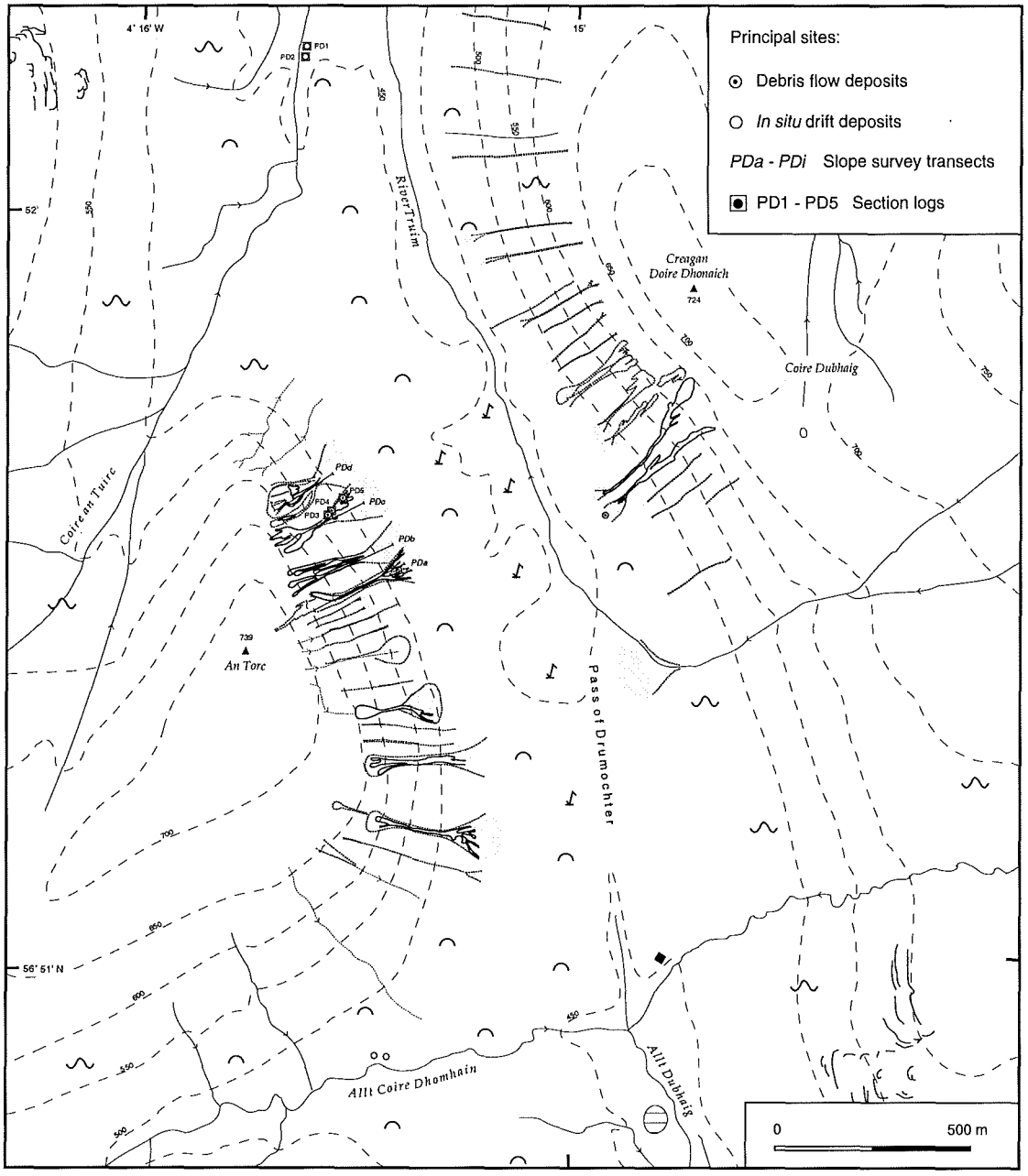


Figure 5.7. Delayed or renewed paraglacial activity in the Pass of Drumochter, Scotland. Key in Figure 4.1.

Highlands have also been divided into two areas to aid comparison of the extent of and constraints on delayed or renewed paraglacial activity. Mapping was carried out in the upper 4.5 kilometres of Glen Einich (Figure 5.6). Area 1 covers the terrain above Loch Einich, and area 2 extends downvalley from the northern end of Loch Einich. Valley-wall rock slopes in area 1 are flanked by an apron of steep, vegetated glacigenic drift which is locally mantled with a veneer of relict talus. In this area gullies feed numerous debris cones which commonly abut the loch-side, particularly on the east-facing slopes beneath Sgòr Gaoith. Data from five slope surveys yielded gully volumes of *c.* 450 m^3 to *c.* 120,990 m^3, with a mean of *c.* 25,310 m^3 (Table 5.1). Gully density in area 1 is approximately 14 gullies per kilometre, and an estimated 8.4% (± 4.8% at 95% confidence) of valley-side drift has been reworked through delayed or renewed paraglacial activity. In contrast, Glen Einich is more open in area 2, where its valley-side rock slopes are less dramatic and gully density is low (1.8 gullies per kilometre), suggesting that only *c.* 0.1% of drift has been reworked.

Field site	Gully density (n km^{-1})	% reworked drift*	Mean gully volume (10^3 m^3)*
Glen Einich	8.2	4.9±2.8	25.3±53.5
Area 1	13.7	8.4±4.8	
Area 2	1.8	1.1±0.6	
Pass of Drumochter	2.6	0.84±0.15	7.5±2.4
Area 1	7.3	4.1±1.3	
Area 2	0.9	0.12±0.12	

Table 5.3. Extent of delayed or renewed paraglacial modification in the Grampian Highlands field sites, Scotland. * 95% confidence limits.

In the Pass of Drumochter (Figure 5.7), area 1 includes the facing slopes of An Torc and Creagan Doire Dhonaich, and area 2 covers the remainder of the

mapped slopes. Unlike the slopes of Glen Einich, the Drumochter Hills are rounded and vegetation-covered, with few rock outcrops. Upper slopes are mantled with a veneer of gelifluctate, but lower valley-side slopes supported a thick cover of glacial drift, the upper boundary of which marks the limit of Loch Lomond Stade glaciation (Sissons, 1974). Active gully erosion is almost entirely confined to area 1, where redeposition of sediment has resulted in the accumulation of several large slope-foot debris cones, but similar landforms are absent in area 2. This within-site variation in the extent of gullying is reflected in the contrasting gully density values between area 1 (7.3 gullies per kilometre, Table 5.3) and area 2 (0.9 gullies per kilometre). The dimensions of four surveyed gullies in the Pass of Drumochter imply *c.* 4,860 m^3 to *c.* 9,840 m^3 of sediment removal, with an average of *c.* 7,460 m^3 (Table 5.1). Whilst these gullies are wider than all but the largest gully surveyed in Glen Einich, the Drumochter gully systems are on average slightly shorter than those measured in the Cairngorms, and their dimensions and frequency suggest that *c.* 4% of all hillslope drift has been reworked in area 1, with reworking of less than *c.* 1% of hillslope drift in the Drumochter site as a whole.

5.3.3 Extent of delayed or renewed paraglacial modification of drift: summary

The extent of delayed or renewed paraglacial activity varies considerably at the Scottish field sites, but is generally very much less widespread than at the most active sites of paraglacial drift reworking and hillslope modification in western Norway (section 4.3). Gully density ranges from 1 gully per kilometre at those sites exhibiting very limited reworking to 25 gullies per kilometre where delayed or renewed paraglacial modification of hillslope drift has been most active. Spatially, the proportion of hillslope drift that has been remobilised by such activity ranges from *c.* 1% in the Pass of Drumochter to *c.* 6.5% in the

Western Red Hills of Skye. Possible factors explaining these variations in the extent of drift reworking are considered in section 5.5 below.

5.4 Processes of sediment transfer.

At each of the four field sites in the Scottish Highlands, debris flow tracks, debris cones, fans and gravel spreads located on the lower slopes of valley sides are a prominent legacy of delayed or renewed paraglacial drift slope modification. Several processes, including debris flow, fluvial activity and snow avalanches, have contributed to downslope redistribution of glacigenic drift, and the significance of each is assessed below.

5.4.1 Debris flows

The dominant microrelief on the majority of debris cones in the Western Red Hills and Glen Docherty consists of levées and terminal lobes indicative of debris flow activity across entire cone surfaces (*cf.* van Steijn *et al.*, 1988; Ballantyne, 1995a). Numerous fresh debris flow tracks descend from gullies cut into the drift on the Western Red Hills, and parallel levées of unsorted sediment often extend several hundred metres upslope from terminal debris lobes. Particularly good examples fringe the slopes of Beinn Dearg Mhór (Figure 5.8) and Glamaig, where debris flows have often been observed in recent years (H. MacLeod, personal communication, 1996). In 1968 a number of debris flows were also triggered by an intense rainstorm on the northern side of Glen Docherty (Strachan, 1976). Moreover, several debris cones at both sites have been freshly incised. The sections cut in these cones reveal stacked sediment units that are usually aligned sub-parallel to the slope surface. These units are dominated by poorly-sorted diamictons that are interpreted as debris flow diamictons (see section 6.3.5). Together, these morphological and sedimentological lines of

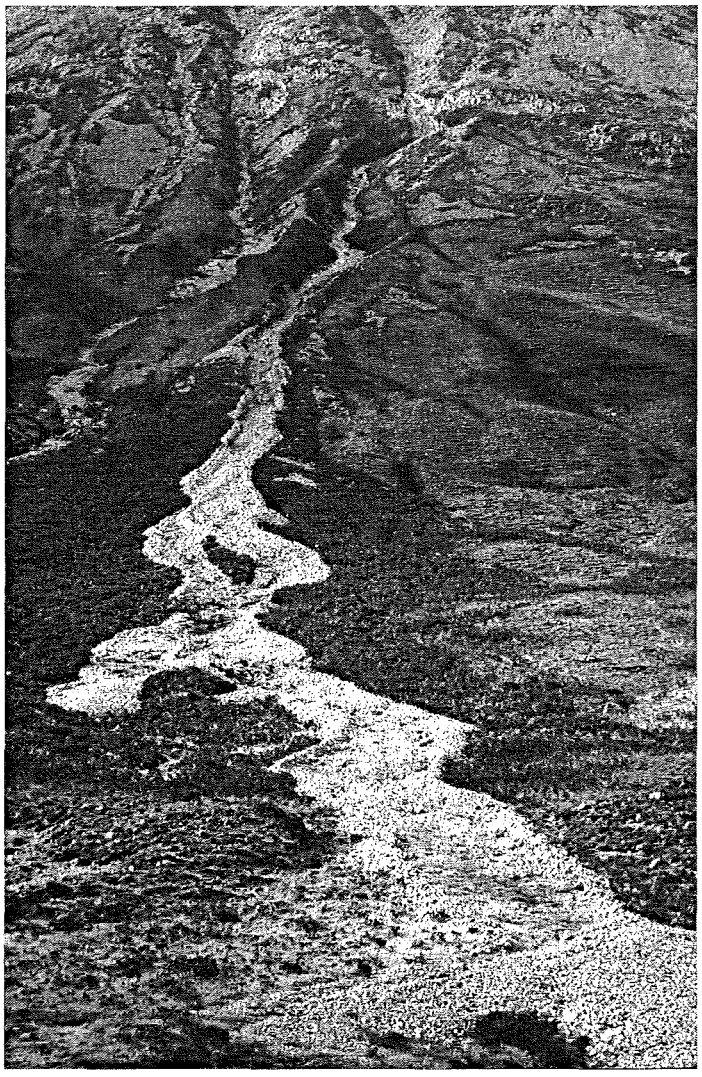


Figure 5.8. Recently-active debris flow tracks emerging from gullies cut into the drift on Beinn Dearg Mhór in the Western Red Hills. Parallel levées of unsorted deposits mark the flow path for several hundred metres upslope from the terminal lobe.

evidence indicate that modification of valley-side drift by successive debris flows has been the dominant formative process of sediment reworking and redeposition on debris cones at the field sites in the Northwest Highlands.

The evidence provided by superficial deposits and surface microtopography supports a similar interpretation for the debris cones and fans mapped in the Grampian Highlands. Several of the debris cones below An Torc and Creagan Doire Dhonaich in the Drumochter Pass exhibit signs of recent debris

flow activity, including an irregular microrelief of fresh hummocky levées up to 1 m high (Figure 5.9) and dissected lobes. Where a sharp break of slope occurs at the slope foot, recent debris flows have been deflected around older, vegetated cone deposits. Sections exposed in profile PDc at Drumochter (Figure 5.7) reveal stacked debris flow deposits, again suggesting that debris flow has been the dominant process responsible for sediment reworking and cone formation. In upper Glen Einich, subdued, irregular ridges and hummocks cover the surfaces of many partially-vegetated cones, and fresh levées extending several hundred metres upslope from a debris cone delineate the path of a recent debris flow from the largest gully below Am Bodach (Figure 5.6).

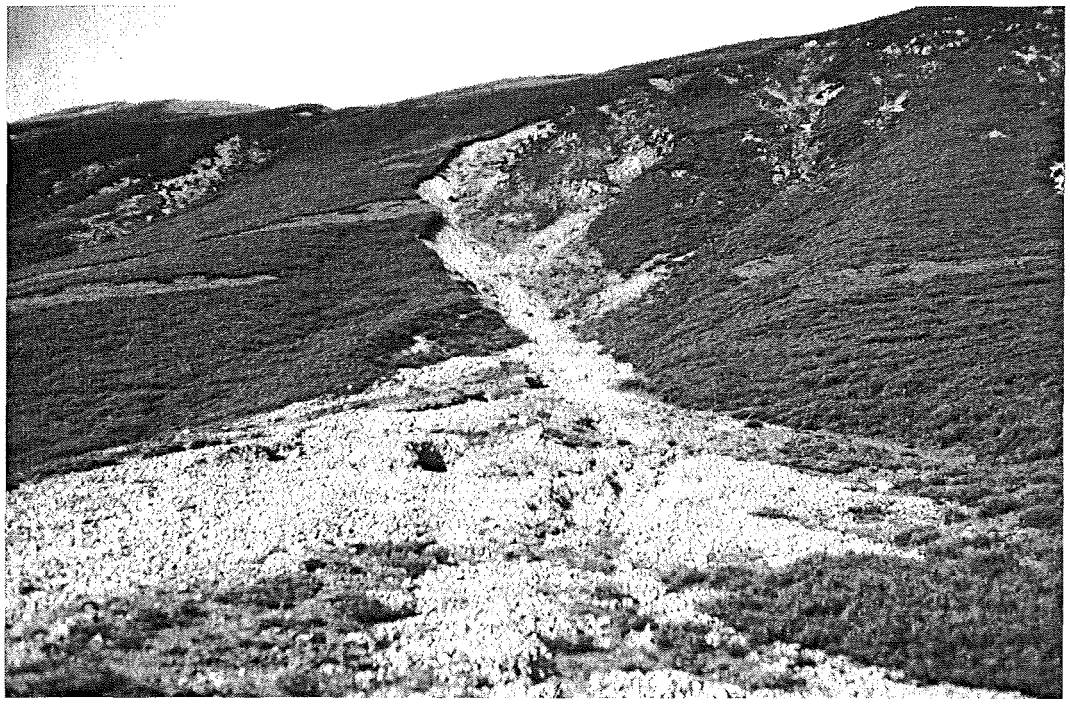


Figure 5.9. Debris cone below An Torc (Pass of Drumochter) showing signs of recent debris flow activity: notice the irregular microrelief of fresh hummocky levées up to 1 m high and dissected lobes.

5.4.2 Surface wash

Surface wash appears to play only a secondary role in transporting sediment at the Scottish field sites. The geomorphological maps (Figures 5.2-5.4

Figure 5.10. Fresh fine-grained sediment splay emanating from debris flow deposits on a largely-vegetated cone surface in the Pass of Drumochter indicative of reworking of glacigenic debris by slopewash processes. Pen for scale.

and 5.6-5.7) suggest that valley-side streams are more widespread at the field sites in the Northwest Highlands than at those in the Grampians: when expressed as the number of valley-side streams per kilometre of valley-side drift, stream density in the Western Red Hills and Glen Docherty is 4.7 and 5.1 streams per kilometre respectively, compared to 3.0 and 2.8 streams per kilometre for the Pass of Drumochter and Glen Einich. In most cases these streams occupy shallow gullies, particularly within the tracks of debris flows, and have been responsible for further downslope reworking of sediment, often resulting in exposure of bedrock

within gully heads. Localized reworking of fresh debris flow deposits is evident in many sections in the form of slopewash facies intercalated between debris flow units (*cf.* Eyles *et al.*, 1988; Eyles and Kocsis, 1988; Brazier and Ballantyne, 1989; Derbyshire and Owen, 1990), especially at the two sites in the Northwest Highlands. Between-site differences in the importance of surface wash as an agent of drift reworking is also suggested by comparison of the extent of thin splays of fine-grained sediment that locally overlie debris cones or emanate from debris flow lobes (Figure 5.10). Such wash deposits are apparently absent in Glen Einich, but widespread in the Western Red Hills and Glen Docherty. Indeed, flood torrent deposits during recent years have frequently infilled 0.5 m deep ditches at the foot of the northern slope of Glamaig (H. MacLeod, personal communication, 1996).

5.4.3 Snow avalanches

Heavy snowfall, prolonged snow-lie and the accumulation of massive cornices at the crest of corries and glacial troughs make the Cairngorms more susceptible to snow avalanching than probably any other upland area in Britain (Ward, 1984). It is not surprising, then, that evidence of snow avalanche activity at the Scottish field sites is mostly confined to upper Glen Einich. Recent avalanche activity is evident in the form of fresh, angular perched boulders and chipped clasts resting amid lichen-covered rocks on several cones overlooking the eastern shore of Loch Einich (*cf.* Ward, 1985; Luckman, 1992; Ballantyne, 1995a). Where topography permits, these cones appear to possess a marked downslope concavity, characteristic of slopes modified by snow avalanches (e.g. Luckman, 1978, 1992; Huber, 1982; Nyberg, 1989). One such debris cone supporting a chaotic assemblage of avalanche detritus amongst unsorted debris flow deposits is the large cone at the base of profile Ea, beneath the cliffs of Am Bodach (Figures 5.6 and 5.11). There is also very localized evidence for former

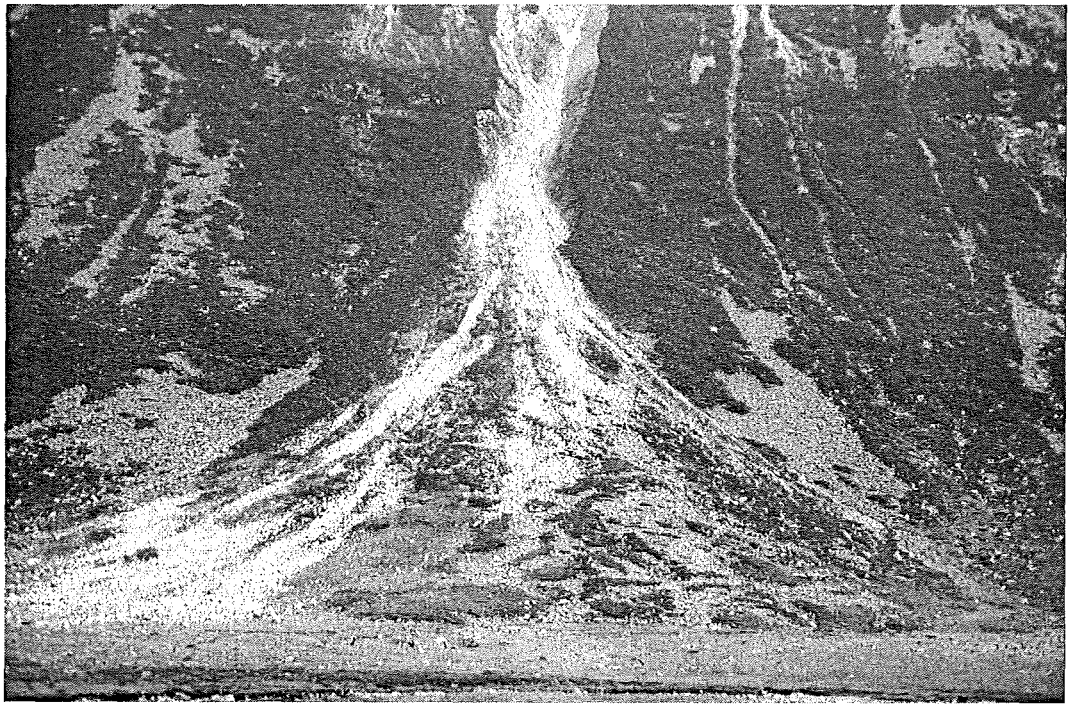


Figure 5.11. Large debris cone in upper Glen Einich supporting numerous perched clasts indicative of debris reworking by recent snow avalanche activity and fresh debris flow deposits emanating from the drift gully upslope. Two tents provide scale in the foreground.

modification of drift and frost-weathered regolith by snow avalanches in the Western Red Hills. Below the cliffs of Beinn Dearg Mheadhonachin (Figure 5.3), a well-developed but vegetated avalanche boulder tongue terminates downslope at a low ramp of boulders that has been interpreted by Benn (1990) as a relict avalanche impact rampart. This interpretation implies that powerful avalanche activity occurred at this site after the wastage of the Loch Lomond Stade glaciers, though there is no geomorphic evidence for recent snow avalanches in the Western Red Hills.

Snow avalanche paths frequently show evidence of debris flow activity. (Kostaschuk *et al.*, 1986; Sauchyn, 1986). Luckman (1992) considered that the geomorphic effectiveness of snow avalanching in the Lairig Ghru glacial breach adjacent to Glen Einich might be enhanced by a symbiotic relationship between debris flow and snow avalanche activity, whereby debris flows supply avalanche

tracks with loose debris for later removal. The composite nature of reworked drift deposits on some debris cones in Glen Einich would appear to support this view. In general, however, this site is exceptional in a Scottish context. The lack of geomorphic evidence for avalanche activity at the other field sites implies that process has been of very localized importance in terms of the remobilisation of valley-side drift.

5.4.4 Processes of sediment transfer: summary

The planimetric area of drift resedimented by debris flows, surface wash and snow avalanches was obtained by digitising cartographic data from Figures 5.2-5.4 and 5.6-5.7 on the basis of detailed field observations. Table 5.4 outlines the relative areal extent of debris flow, surface wash and snow avalanche deposits as a percentage of the total reworked drift cover at each site. At all sites, debris flow deposits account for more than 60% of reworked sediment by area, whilst areal cover of surface wash deposits is between 20% and 35%. Only at Glen Einich are snow avalanche deposits significant, accounting for an estimated 19% of reworked sediment by area. However, it should be noted that the relative cover of each type of deposit is not an accurate reflection of the volume of sediment

Field site	% Debris flow	% Snow avalanche	% Surface wash
Northwest Highlands			
Western Red Hills	67	1	32
Glen Docherty	65	0	35
Grampian Highlands			
Glen Einich	61	19	20
Pass of Drumochter	75	0	25

Table 5.4. Percentage areal extent of debris flow, surface wash and snow avalanche deposits as components of total reworked drift at field sites in the Scottish Highlands.

transported by each of the three processes, nor of the relative geomorphologic importance of these processes in terms of drift remobilisation. Taking into account the generally much greater thickness of debris flow deposits than slopewash or snow avalanche debris, together with the spatial dominance of such deposits at all sites, it is evident that (in terms of volume of sediment transported) debris flow activity is undoubtedly the most effective process in the delayed or renewed paraglacial remobilisation of valley-side drift at the study sites in the Scottish Highlands.

5.5 Constraints on delayed or renewed paraglacial modification of drift.

As the data presented in section 5.3 demonstrate, the field sites in the Scottish Highlands are not equally endowed with landforms representing delayed or renewed paraglacial modification of drift slopes. These spatial variations in evidence for delayed or renewed paraglacial reworking of hillslope drift covers probably reflect the complex interaction of both intrinsic and extrinsic controls. Whilst topographic setting, relief amplitude and bedrock structure undoubtedly dictate the energy available to all slope processes and the availability of entrainable sediment mantling steep slopes within a basin, the dataset reported in this thesis is insufficiently large to establish the significance of these controls on paraglacial activity. However, the influence of certain intrinsic controls has been considered in a number of previous studies relating to debris flow activity in the Scottish Highlands. Innes' (1982, 1983b) work on the distribution of hillslope and valley-confined debris flows, for example, illustrated the influence of local lithology on debris flow activity, which tends to be more prominent on lithologies that have weathered to produce cohesionless sand-rich regolith, such as sandstone and granite, than in areas of silt-rich regolith cover. Ballantyne (1991c) came to similar conclusions regarding the influence of lithology on the distribution of

debris flows on Skye. A further important control on the distribution of debris flow activity is sediment availability (Statham, 1976a; Strachan, 1976; Brazier and Ballantyne, 1989). Glacially-scoured areas of Scotland such as Knoydart, Morar and Morvern support only a patchy drift cover on most hillslopes, and hence there is only very localised evidence for debris flow activity (Ballantyne and Harris, 1994). The possible importance of a number of intrinsic and extrinsic controls on the degree of delayed or renewed paraglacial reworking evident in different areas are considered below, using gully density as a surrogate measure of intensity of reworking of drift, as in section 4.5.

5.5.1 Intrinsic controls

To investigate possible intrinsic controls on gully density at the Scottish field sites, four hypotheses were tested using data amalgamated from all four sites. These are: (1) that gully density increases with drift-slope gradient; (2) that gully density increases with drift thickness; (3) that gully density is inversely related to sediment packing; and (4) that gully density is related to matrix granulometry. Gully density for both areas in each site was therefore plotted in turn against various topographic and sedimentological parameters, namely upper drift slope gradient, drift thickness, drift matrix void ratio, inclusive graphic skew, and percentage of sand (60-2000 μm) and silt (2-60 μm) within fine-grained (< 2 mm) samples of drift (Figure 5.12). Where appropriate, the significance of individual correlations was tested using Spearman's rank correlation coefficient (r_s).

Slope gradient is widely recognised as a major intrinsic control of hillslope debris flow initiation (e.g. Takahashi, 1981; Innes, 1983a; Costa, 1984). Although he was referring to Lateglacial paraglacial activity, Benn (1990) found that on Skye, the maximum slope angle on which moraines are preserved is *c.* 20°. Figure 5.12a suggests that a threshold gradient exists between 25° and 29°: at lower

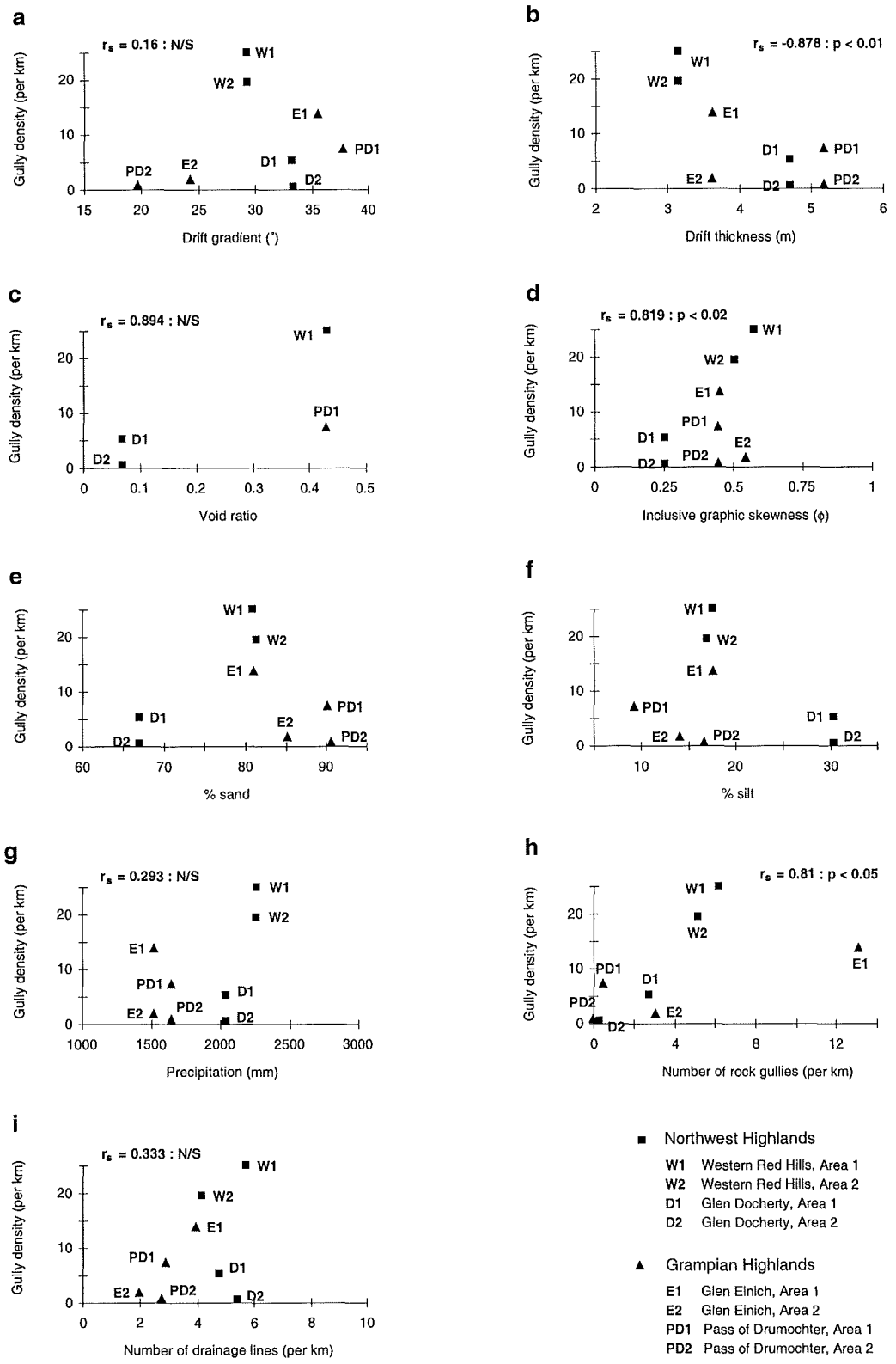


Figure 5.12. Gully density relations with (a) drift gradient, (b) drift thickness, (c) matrix void ratio, (d) graphic skewness, (e) % sand content, (f) % silt content, (g) regional precipitation, (h) the number of feeder gullies and (i) the number of valley-side streams at field sites in the Northwest Highlands and the Grampian Highlands. Spearman's rank correlation coefficient is denoted by r_s. p describes the significance level of the correlation. N/S: not significant.

gradients, gully density is very low (< 2 km^{-1}), but at higher gradients gully density is much higher, though not at all sites. This seems to imply that a drift-slope gradient exceeding 25-29° represents a prerequisite for extensive paraglacial modification of valley-side drift, but that other factors determine whether extensive gullying actually occurs on drift slopes steeper than 25-29°. In both Glen Docherty area 1 and the Pass of Drumochter area 1 gully densities are low (< 8 km^{-1}), even though upper drift slope gradients appear sufficiently steep (33.2° and 37.8° respectively) for widespread debris flow activity and associated gully development.

Another possible control on the extent of drift reworking is drift thickness. Figure 5.12b reveals a significant ($p < 0.01$) negative relationship between the two variables: gullying is most widespread at sites where drift is no more than *c.* 3-4 m thick, suggesting that initial failure and consequent flow of debris may reflect more rapid build-up of pore-water pressures during rainstorms at sites with relatively shallow drift cover. However, the negative relationship between gully density and drift thickness conflicts with the analysis of similar data from the Norwegian sites (see section 4.5), suggesting that there is no simple relationship between the two variables.

A third possible intrinsic control on gully density is drift matrix void ratio: where drift is loosely packed (high void ratio), not only is shearing resistance likely to be lower, but also higher infiltration rates may lead to a more rapid rise in pore-water pressures during intense precipitation, causing a greater reduction in shearing resistance than where drift is tightly packed. Mean void ratio data are available for only four areas, but Figure 5.12c suggests a positive correlation between the two variables. This may explain the relatively high density of gullies on the rather loose sandy regolith on upper slopes in the Western Red Hills.

The fourth hypothesis suggested above concerns the possibility that matrix granulometry exerts a control on gully density. On the basis of extensive granulometric analysis of frost-weathered regoliths in the Scottish Highlands, Innes (1982, 1986) noted that the grain-size distributions of regoliths susceptible to debris flow activity tend to be more fine-skewed than those of regolith covers on slopes where debris flows are rare or absent, and inferred an inverse relationship between susceptibility to debris flow activity and field moisture capacity. Figure 5.12d reveals a moderate but significant (at $p < 0.02$) positive relationship between gully density and the inclusive graphic skew of particle size distributions at the field sites, also suggesting a possible link between the negative skewness of particle-size distributions of drift cover and more extensive gully development. Innes (1982, 1983c) also observed that debris flow activity appears to be much more common on lithologies that support a coarse, sand-rich regolith such as Torridonian Sandstone and granite, and less widespread on finer, silt-rich regoliths. Susceptibility of areas of sandy drift or regolith to debris flow may reflect high infiltration rates, and hence more rapid build-up of pore-water pressures during intense rainstorms (Ballantyne, 1986). However, though gully densities are generally higher at the field sites underlain by granite (the Western Red Hills and Glen Einich) than those underlain by schists (Glen Docherty and the Pass of Drumochter), Figures 5.12e and 5.12f reveal no clear relationship between gully density and either percentage sand or percentage silt content, suggesting that neither variable represents a primary intrinsic control on the degree of delayed or renewed paraglacial reworking of valley-side drift.

In sum, hypothesis testing based on the relationships between gully density and possible intrinsic controls suggests that delayed or renewed paraglacial reworking of drift tends to be limited where upper drift slope gradient is less than 25-29°, and where drift thickness exceeds *c.* 4 m. The influence of sediment properties is more difficult to evaluate. Though no simple relationship between

gully density and percentage sand or percentage silt was detected, there is evidence that gully density tends to be relatively high on drift where the grain-size distribution is fine-skewed and where void ratio is relatively high. Both attributes are characteristic of drift derived mainly from granitic rocks, and possibly account for the generally higher densities of gullies on granite terrain.

5.5.2 Extrinsic controls

Three possible extrinsic controls on gully density at the Scottish field sites were considered, namely: (1) that gully density increases with regional (mean annual) precipitation (Figure 5.12g); (2) that gully density increases with the number of feeder gullies delivering water to the drift cover from upslope (Figure 5.12h); and (3) that gully density increases with the number of drainage lines (stream valleys) developed on the drift slope (Figure 5.12i). The number of feeder gullies and drainage lines are expressed as the average frequency per kilometre of slope. As observed earlier (section 4.5.2), these possible extrinsic controls are probably inter-related and may differ from past extrinsic controls at the time of gully initiation. Consequently, the following analysis should be treated with caution.

No significant correlation was detected between gully density and mean annual precipitation, though it is notable that the wettest sites (on Skye) support the highest gully density. This suggests that high rainfall favours gully development where other (intrinsic) conditions are propitious, a relationship that might be expected in view of the importance of high pore-water pressures in initiating debris flows (e.g. Caine, 1980; Rapp and Strömquist, 1976; Lawson, 1982; Wieczorek and Jäger, 1996). However, it should be noted that mean annual precipitation is not necessarily a direct influence on the frequency of slope failure and debris flow events, which are essentially controlled by the intensity, duration

and frequency of high-magnitude rainstorms, and by antecedent moisture conditions (Watanabe, 1985; Church and Miles, 1987; Zimmermann and Haeberli, 1992; Kotarba, 1997).

As at the sites in Norway, an apparent link between slope hydrology and gully density is reflected in a weak but significant (at $p < 0.05$) correlation between gully density and the number of feeder gullies upslope (Figure 5.12h). Such gullies focus water onto drift downslope, promoting gully incision by running water and indirectly by slope failure and consequent flow of debris (van Steijn *et al.*, 1988; Ballantyne and Benn, 1994; Ballantyne, 1995a; Benn and Evans, 1998). The influence of rock gullies on the density of gullies cut in drift downslope is particularly evident in Glen Docherty, where differences in gully density on the north side (5.4 gullies per kilometre) and south side (0.6 gullies per kilometre) of the glen appear directly related to the frequency of feeder gullies cut in bedrock upslope (2.7 per kilometre on the north side of the glen, 0.3 per kilometre on the south side). However, despite the significant positive relationship between gully density and number of feeder gullies at the Scottish sites, the latter are much less numerous than the former, except in Glen Einich area 1. This discrepancy is also evident in the relationship between drift gully density and feeder gully density at the Norwegian sites (section 4.5.2), and suggests that the link between the two is indirect; for although many drift gullies are positioned immediately downslope from rock-cut gullies, others are not. Nevertheless, the largest gullies cut into valley-side drifts in the Western Red Hills and Glen Einich often occur directly downslope of pronounced rock-cut gullies or complex rock-cut gully systems. An outstanding example is gully Ea in Glen Einich, which has a volume of *c.* 120,990 m^3, and is fed by a network of five rock-cut gullies upslope.

The third hypothesis considered is that gully density might be expected to be positively correlated with drainage density, which in turn is related to precipitation. Figure 5.12i indicates that although there is no significant correlation between the two variables, areas of low drainage density are associated with low gully density, but not *vice versa.* This implied that though the wettest areas support the densest gullying, not all wet areas are densely-gullied.

Although possible auto-correlation of independent variables does not permit firm conclusions to be reached concerning the extrinsic controls on gully density and thus on delayed or renewed paraglacial drift modification, the above relationships nonetheless suggest that delivery of water to the slope is of critical importance. In general, the highest gully densities are associated with the wettest areas (Skye), but in equally wet areas where intrinsic controls are less favourable, gully density may be low (Glen Docherty). At a more local level, the positive relationship between density of rock-cut feeder gullies and that of gullies cut in drift downslope illustrates the critical role of the former in the development of some of the largest gullies by focusing delivery of water onto the drift slope, though no more than about 50% of drift gullies in the mapped areas are directly associated with rock-cut gullies in this way.

In general, no single variable appears to determine the distribution or frequency of gullies incised in valley-side drift deposits. Some factors, such as gradient and (perhaps less convincingly) depth of drift appear to constrain gully development, whilst others (sediment packing, granulometry, mean annual precipitation and density of rock-cut gullies upslope) appear to render particular slopes either more or less susceptible to gully development. The critical factor would appear to be the susceptibility of individual slopes to failure and remobilisation of sediment, particularly during extreme rainstorm events. Although the above analyses provide an indication of the factors that make

particular slopes more vulnerable to failure and consequent gullying, the location, frequency, intensity and duration of extreme rainstorms - and indeed the nature of antecedent soil moisture conditions at the time of their occurrence - represent essentially random and unpredictable variables in both space and time.

5.6 Timing and possible causes of delayed or renewed para-glacial modification of drift.

At the four sites investigated in the Scottish Highlands, there is evidence for both recent and ancient reworking of valley-side drift deposits. Recent activity is evident in the form of unvegetated gully systems and fresh accumulations of reworked sediment at the foot of slopes, and vegetated, stable gullies and debris cones imply a much older phase or phases of drift reworking. The juxtaposition of stable, vegetated drift slopes and recently-active gullies, often located within larger, vegetated gullies, suggests recent reactivation of previously stable valley-side drift (Ballantyne and Benn, 1996). Moreover, at some sites evidence for previous episodes of sediment remobilisation is evident in natural sections cut through debris cones, in the form of buried soil or peat horizons intercalated with stacked sequences of debris flow and slopewash deposits. The aim of the research reported in this section is to investigate the timing and possible causes of such delayed or renewed paraglacial reworking of glacigenic valley-side drift, through a combination of radiocarbon dating of buried organic horizons and analysis of sub-fossil pollen assemblages and charcoal content within such horizons.

5.6.1 The buried organic horizons

Several sections cut through debris cones at the Scottish field sites exhibit thin organic-rich horizons intercalated with debris flow and slopewash sediment units. Such organic-rich horizons are sub-parallel to the cone surface and are

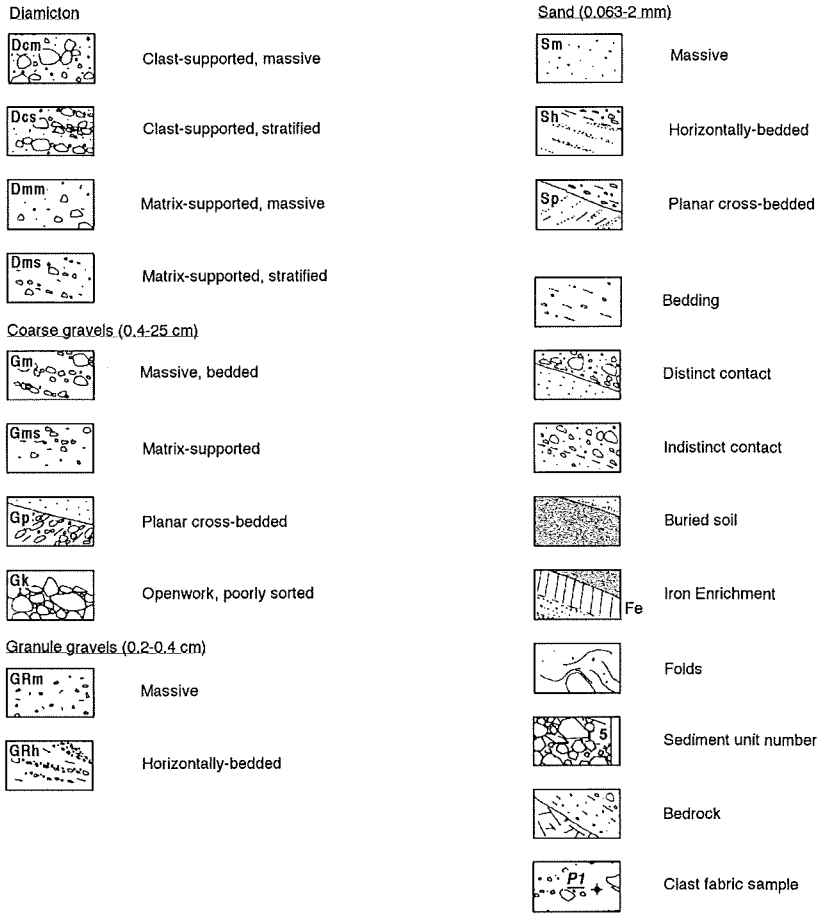


Figure 5.13. Key to symbols used in the lithofacies logs (adapted from Benn, 1990).

often continuous for several metres in a downslope direction. Five of these sections were excavated to produce a clean vertical face, then photographed and logged. The key to the symbols employed in the section logs is given in Figure 5.13, and the five sections are illustrated in Figure 5.14. In all sections, the dominant minerogenic units are clast- or matrix- supported massive or stratified diamictons, which are interpreted as stacked debris-flow deposits that represent resedimentation of glacigenic deposits derived from upslope (see chapter 6). Occasional thinner interbeds of sand or granule gravels are interpreted as slopewash horizons, possibly representing eluviation of fine sediment from debris

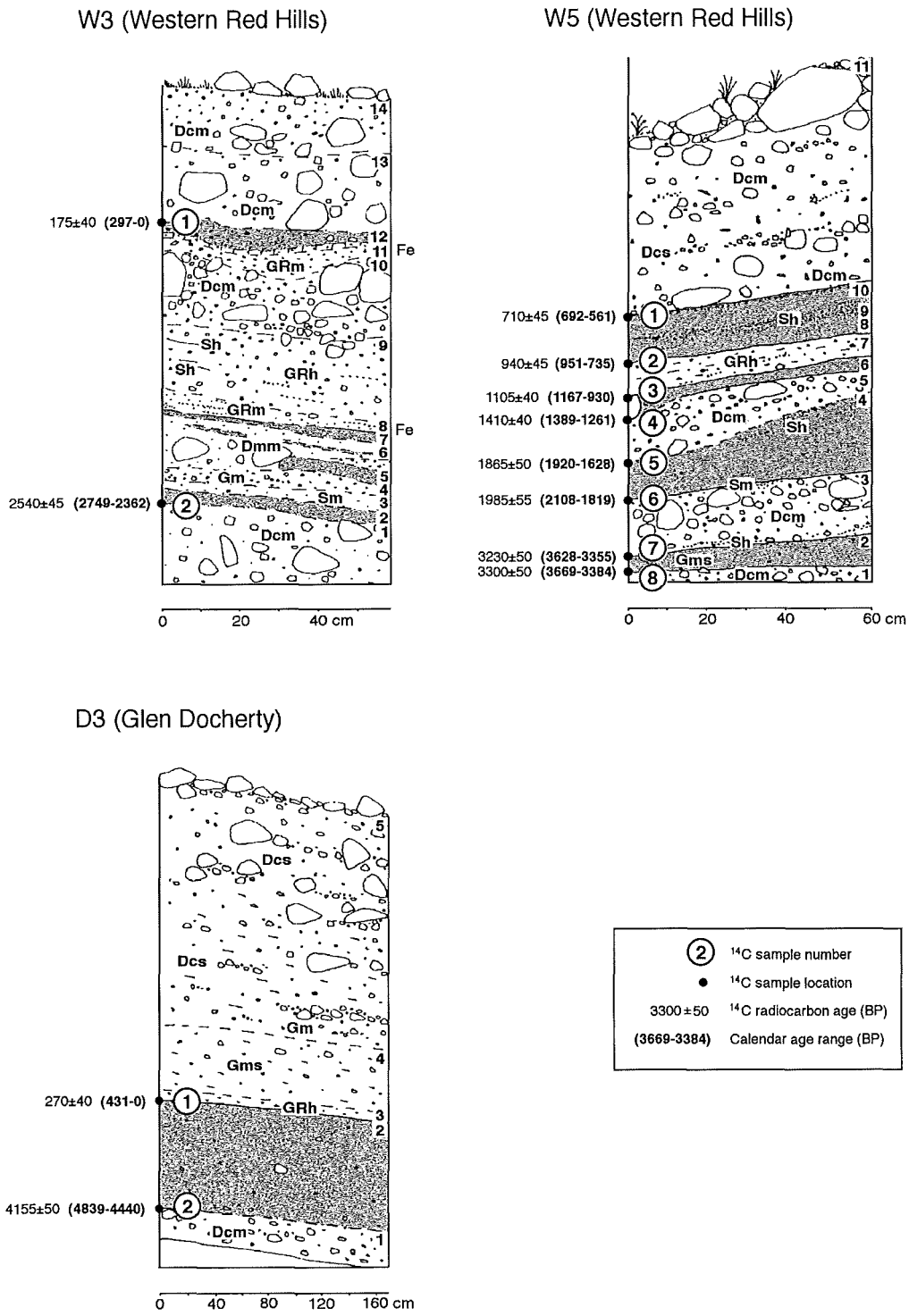
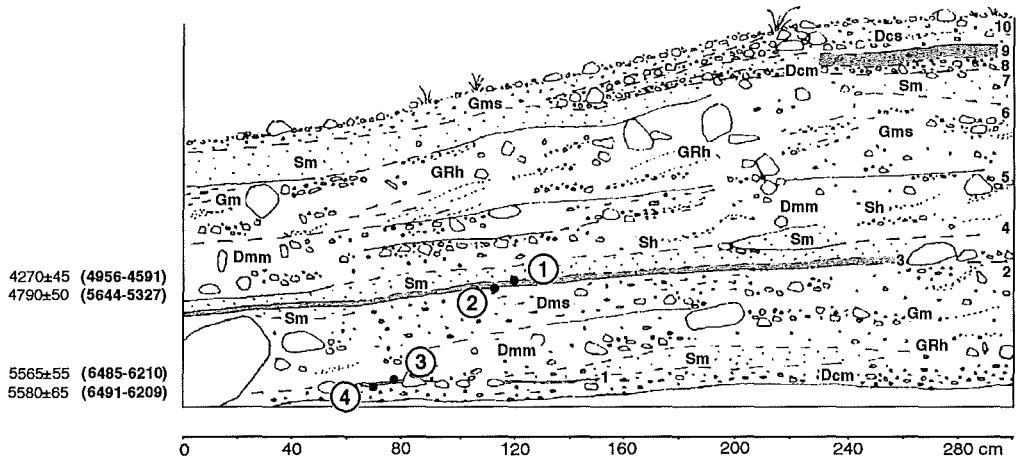


Figure 5.14. ^{14}C radiocarbon dates and calibrated calendar ages for buried soils exposed in sections through debris cones in the Scottish Highlands. ^{14}C dates were transformed to calendar dates using the conversion programme CALIB 3.0 of Stuiver and Reimer (1993) and the calibration data of Stuiver and Becker (1993). Calendar age ranges are given as 95% confidence limits. No vertical exaggeration. Key in Figure 5.13. Bold numerals along the right-hand margin of each section column represent the numbering of individual sediment units from the base upwards. Section D3 is < 5 m upslope from section D4, and is located within the same debris flow track.

D4 (Glen Docherty)



PD3 (Pass of Drumochter)

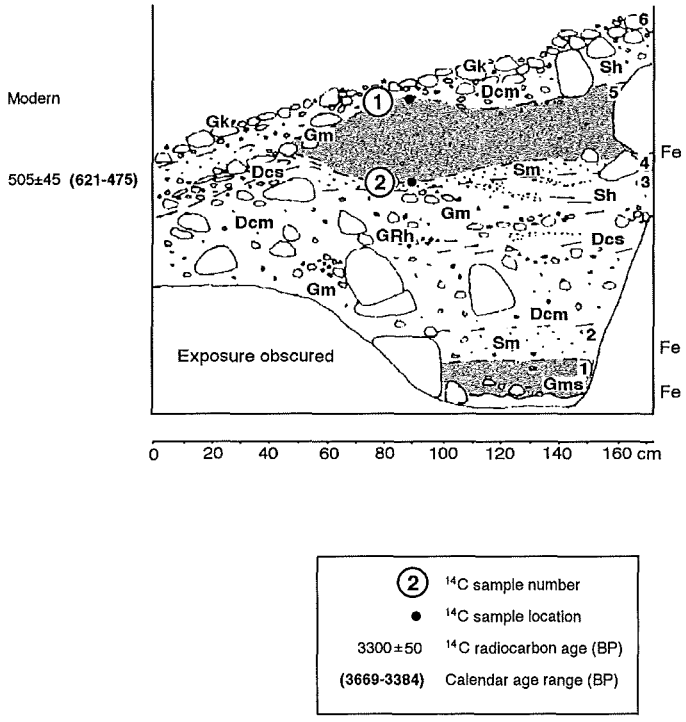


Figure 5.14. ^{14}C radiocarbon dates and calibrated calendar ages for buried soils exposed in sections through debris cones in the Scottish Highlands. ^{14}C dates were transformed to calendar dates using the conversion programme CALIB 3.0 of Stuiver and Reimer (1993) and the calibration data of Stuiver and Becker (1993). Calendar age ranges are given as 95% confidence limits. No vertical exaggeration. Key in Figure 5.13. Bold numerals along the right-hand margin of each section column represent the numbering of individual sediment units from the base upwards. Section D4 is < 5 m downslope from section D3, and is located within the same debris flow track.

flow deposits or overbank deposition by streams.

The organic-rich layers range from thin buried soils to well-humified peat layers up to 0.8 m thick. These soil and peat layers are interpreted as representing former surface organic horizons that developed during periods of stability. Their variable thickness and characteristics, however, suggest different degrees of soil development. Some organic horizons, such as unit 2 at section W5, are highly organic (up to 73% by weight loss on ignition), whereas others yielded loss-on-ignition values of only *c.* 10% by weight. Some horizons exhibit staining indicative of iron enrichment, and *in situ* weathering of clasts at their lower contact (e.g. unit 5 at section PD3), but others reveal no evidence for either process. Such contrasting characteristics probably reflect differences in palaeosol maturity, parent material, local drainage and preservation. The nature of the minerogenic component of such layers largely reflects the composition of the underlying sediment unit, or localised inwash of sediment from upslope. Some organic layers (notably unit 5 in section PD3) exhibit truncation or thinning in an upslope or downslope direction, indicating erosion prior to deposition of overlying minerogenic sediments, and it is possible that the absence of buried palaeosols in several excavated sections (notably in upper Glen Einich) reflects complete removal of such soils prior to the emplacement of overlying sediment units. Alternatively, drift slopes at these locations may not have been stable over sufficiently prolonged periods to permit the formation of mature organic-rich soils.

5.6.2 Timing of delayed or renewed paraglacial modification of drift

Radiocarbon dating of buried palaeosols was undertaken to define the timing of episodes of slope instability resulting from reworking of glacigenic drift by successive debris flow and slopewash events, and that of intervening periods of

slope stability associated with soil development and peat growth (Figure 5.14 and Table 5.5). However, radiocarbon dating of buried soils can be problematic. Soils are a complex mixture of organic components which may have experienced a long history of mixing and translocation prior to burial. The radiocarbon age for a buried soil therefore reflects not only time elapsed since burial, but also the radiocarbon age of constituent organic carbon prior to burial (Matthews, 1980, 1992b, 1993). The latter is termed the 'apparent mean residence time' (AMRT) of the soil (Campbell *et al.*, 1967), and increases approximately linearly with soil depth (Scharpenseel, 1972; Matthews, 1980; Matthews and Caseldine, 1987). Consequently, though a radiocarbon age determination for the top of a buried soil provides only a maximal estimate of the date of burial, the possible effect of AMRT on the age of soil burial can be minimised by collecting very thin (in this case 3-5 mm thick) samples of organic material from the very top of buried soil horizons (*cf.* Ballantyne, 1986c; Harris *et al.,* 1987; Scharpenseel and Becker-Heidmann, 1992). Moreover, AMRT effects are likely to be minimal for well-humified peat horizons or peaty soils, and for very thin, immature soils that have developed over short time periods. However, if the upper surface of a soil has been eroded prior to emplacement of overlying minerogenic sediments, it is likely that the resultant date will considerably overestimate age of burial. Conversely, radiocarbon age determinations for the base of buried soils reflect the 'average' age of organic material present and thus tend to be minimal for the onset of soil development, particularly as there may have been a temporal hiatus between surface stabilisation and the beginning of soil (or peat) formation. Of the eighteen samples submitted for radiocarbon assay, nine were collected from the top of buried palaeosols or peat layers and thus yield maximal ages for the deposition of overlying sediment units. The remaining nine samples correspond to the base of buried palaeosols, and thus represent minimal ages for the onset of peat growth and soil formation, and hence for periods of slope stability.

Section	Sample No.	Unit No.	Sample depth (cm)	Laboratory code	Radiocarbon age ^{14}C yr BP	$\delta^{13}C$ ‰	Calibrated age* calendar yr BP	
Western Red Hills								
W3	1	12	41	AA-25617	175±40	-27.6	297-0	†
W3	2	2	113	AA-25618	2540±45	-26.2	2749-2362	§
W5	1	10	60	AA-25615	710±40	-27.6	692-561	†
W5	2	8	74	AA-25616	940±45	-28.1	951-735	§
W5	3	6	82	AA-25613	1105±40	-27.8	1167-930	†
W5	4	6	83	AA-25614	1410±40	-27.7	1389-1261	§
W5	5	4	89	SRR-5931	1865±50	-28.0	1920-1628	†
W5	6	4	109	SRR-5932	1985±55	-27.8	2108-1819	§
W5	7	2	127	SRR-5933	3230±50	-27.9	3628-3355	†
W5	8	2	135	SRR-5934	3300±50	-28.2	3669-3384	§
Glen Docherty								
D3	1	2	276	AA-28992	270±40	-28.4	431-0	†
D3	2	2	369	AA-28995	4155±50	-29.2	4839-4440	§
D4	1	3	111	AA-28229	4270±45	-27.5	4956-4591	†
D4	2	3	113	AA-28994	4790±50	-27.3	5644-5327	§
D4	3	1	140	AA-28993	5565±55	-26.4	6485-6210	†
D4	4	1	141	AA-28996	5580±65	-26.4	6491-6209	§
Pass of Drumochter								
PD3	1	5	30	SRR-6188	Modern	-27.5	-	†
PD3	2	5	70	AA-28230	505±45	-28.2	621-475	§

Table 5.5. Radiocarbon ages for buried organic-rich palaeosols exposed in sections through debris cones at field sites in the Scottish Highlands. * radiocarbon ages were transformed to calibrated calendar age ranges using the conversion programme of Stuiver and Reimer (1993) and the calibration data of Stuiver and Becker (1993). The calibrated age range is given as 95% confidence limits. † denotes maximal age for the burial of a soil horizon; § denotes minimal age for the onset of pedogenesis following surface stabilisation.

Timing of periods of slope stability

Radiocarbon dates from the base of peat layers or buried palaeosols provide minimal ages for the onset of peat growth or pedogenesis, and thus commencement of a period of slope stability. However, radiocarbon ages for organic soil or peat on the surface of a reworked drift deposit also inevitably reflect the age of emplacement of the underlying minerogenic sediment. Consequently, a minimal age for the onset of peat growth or pedogenesis at a specific stratigraphic level is likely to be determined by the timing of the preceding phase of sediment reworking. Moreover, radiocarbon ages obtained for buried palaeosols may represent only localised peat or soil development rather than widespread drift slope stability. Thus although radiocarbon dates for samples collected from the bottom of buried organic horizons may be employed in association with dates for the tops of such horizons to establish a minimal duration for phases of local drift slope stability at individual sections, they do not necessarily indicate phases of widespread slope stability.

Samples from the bottom of each of the four buried organic horizons exposed in section W5 in the Western Red Hills yielded dates of 3300±50 yr BP (3669-3384 cal yr BP), 1985±55 (2108-1819 cal yr BP; unit 4), 1410±40 yr BP (1389-1261 cal yr BP; unit 6) and 940±45 yr BP (951-735 cal yr BP; unit 8), each representing minimal ages for the onset of slope stability at these levels. Together with the dates for the samples from the tops of these horizons, these dates define the minimum duration of four periods of local slope stability: from 3300±50 yr BP (3669-3384 cal yr BP) to 3230±50 yr BP (3628-3355 cal yr BP); from 1985±55 yr BP (2108-1819 cal yr BP) to 1865±50 yr BP (1920-1628 cal yr BP); between 1410±40 yr BP (1389-1261 cal yr BP) and 1105±40 yr BP (1167-930 cal yr BP); and from 940±45 yr BP (951-735 cal yr BP) to 710±45 yr BP (692-561 cal yr BP). Samples from section W3 in the same area were taken from the base of the lowest

exposed organic horizon (2540 ± 45 yr BP (2749-2362 cal yr BP)) and the top of the uppermost organic horizon (175 ± 40 yr BP (297-0 cal yr BP)). These two samples are separated stratigraphically by a complex sequence of wash, debris flow and intercalated organic-rich horizons, indicating a complex history of stability and instability between these two dates. In Glen Docherty dates of 4155±50 yr BP (4839-4440 cal yr BP) and 270±40 yr BP (431-0 cal yr BP) obtained from respectively the top and base of a thick peat deposit in section D3 indicate that a period of prolonged stability lasting over 4000 years ended within the past few centuries with emplacement of the overlying gravel and diamict horizons. At section D4 in Glen Docherty, paired dates from the base and top of two thin organic horizons suggest minimum periods of stability from 5580±65 yr BP (6491-6209 cal yr BP) to 5565±55 yr BP (6485-6210 cal yr BP) and 4790±50 yr BP (5644-5327 cal yr BP) to 4270±45 yr BP (4956-4591 cal yr BP), separated and succeeded by emplacement of debris flow deposits. In Drumochter Pass, samples from the top and bottom of a buried organic in section PD3 indicate stability from 505±45 yr BP (621-475 cal yr BP) until recent burial under a shallow cover of debris flow deposits.

Surprisingly, these results suggest that most of the implied periods of stability at sections W5 and D4 are very much briefer than the intervening periods of sediment accumulation. Indeed, in two instances the dates from the top and base of particular organic horizons are virtually identical, so that the calculated 95% confidence limits for equivalent calendrical age strongly overlap. This is true for the dates bracketing the lowermost organic horizon in section W5 (3300±50 yr BP and 3230±50 yr BP, equivalent to 3669-3384 cal yr BP and 3628-3355 cal yr BP respectively), and the lowermost organic horizon in section D4 (5580±65 yr BP and 5565±55 yr BP, equivalent to 6491-6209 cal yr BP and 6485-6210 cal yr BP respectively). Conversely, some dates appear to imply very prolonged periods of uninterrupted instability and sediment accumulation. The

layer of debris flow deposits overlying the lowermost organic layer in section W5, for example, is bracketed by samples dating to 3250±50 yr BP (3628-3355 cal yr BP) and 1985±55 yr BP (2108-1819 cal yr BP), implying that this 20 cm thick minerogenic layer accumulated over at least 1200 years (and possibly as long as 1800 years) without interruption by a period of stability sufficient to allow renewed organic soil development. This is clearly not feasible, and is manifestly inconsistent with the nature of debris flow deposition, which is rapid and sporadic. Indeed, it might be expected that dates bracketing such individual debris flow units should be almost identical, given the rapidity with which deposition and ensuing stabilisation occur. Such inconsistencies imply either that one or more erosional events have truncated the stratigraphic record between dated samples bracketing minerogenic sediments, or that pedogenesis or peat growth following emplacement of individual minerogenic units was delayed for centuries. Several lines of evidence favour the former interpretation, notably (1) the closeness of the top and basal radiocarbon ages of some organic horizons; (2) the long time intervals separating the dates bracketing some minerogenic horizons, where these appear to record only a single debris flow event (e.g. units 3 and 5 in section W5); and (3) the clear evidence for truncation or downslope thinning of both minerogenic layers (e.g. units 6 and 7 in section D4) and particularly organic horizons (e.g. units 2 and 4 in section W5; unit 1 in section D4; and unit 5 in section PD3). These truncated or thinning organic layers provide compelling evidence of surface erosion prior to the emplacement of the overlying sediment units. Conversely, there is no direct evidence for delay in the onset of pedogenesis following sediment deposition. Unpublished observations by C.K. Ballantyne (personal communication, 1998) on debris flow lobes deposited in July 1978 in Drumochter Pass and in September 1985 below the Lomond scarp in Fife indicate that both had become completely vegetated within twenty and ten years of deposition respectively, suggesting that vegetation development and associated pedogenesis is, on a millennial timescale, virtually instantaneous.

The implication of this conclusion is that whereas radiocarbon dates from the base of organic horizons may provide a reasonably accurate (if minimal) age estimate for the onset of renewed stability and consequent organic soil or peat development, dates from the top of organic horizons provide an approximate age for the deposition of the overlying sediment only where there is reasonable stratigraphic evidence to suggest that the two are conformable, and not separated by one or more episodes of erosion. This is probably the case where the organic horizon is of consistent width both upslope and downslope, as with the thick peat horizon in section D3, the uppermost organic horizon in section W5, and the thin but extensive buried soil horizon represented by unit 3 in section D4. All other organic horizons may be considered suspect in this respect, and radiocarbon ages from the tops of such horizons provide no more than a maximal limiting age for the timing of deposition of the overlying sediment; such ages, however, may actually pre-date the depositional event by several centuries. An interesting implication of this argument is that the dates obtained from the base of organic horizons are likely in some instances to provide a much more accurate (if slightly young) estimate of the age of emplacement of the underlying deposits than the ages obtained from the tops of truncated organic horizons that underlie minerogenic sediment units.

Timing of depositional events

The above arguments imply that the duration of depositional events indicative of slope instability and delayed paraglacial reworking of glacigenic drift cannot in all cases simply be inferred from the radiocarbon ages obtained for samples from the base of the overlying organic horizon and top of the underlying organic horizon, as erosion of the latter may lead to erroneous inferences. Furthermore, it is likely that the depositional sequences evident at individual sections represent a sample of the total number of depositional events on the

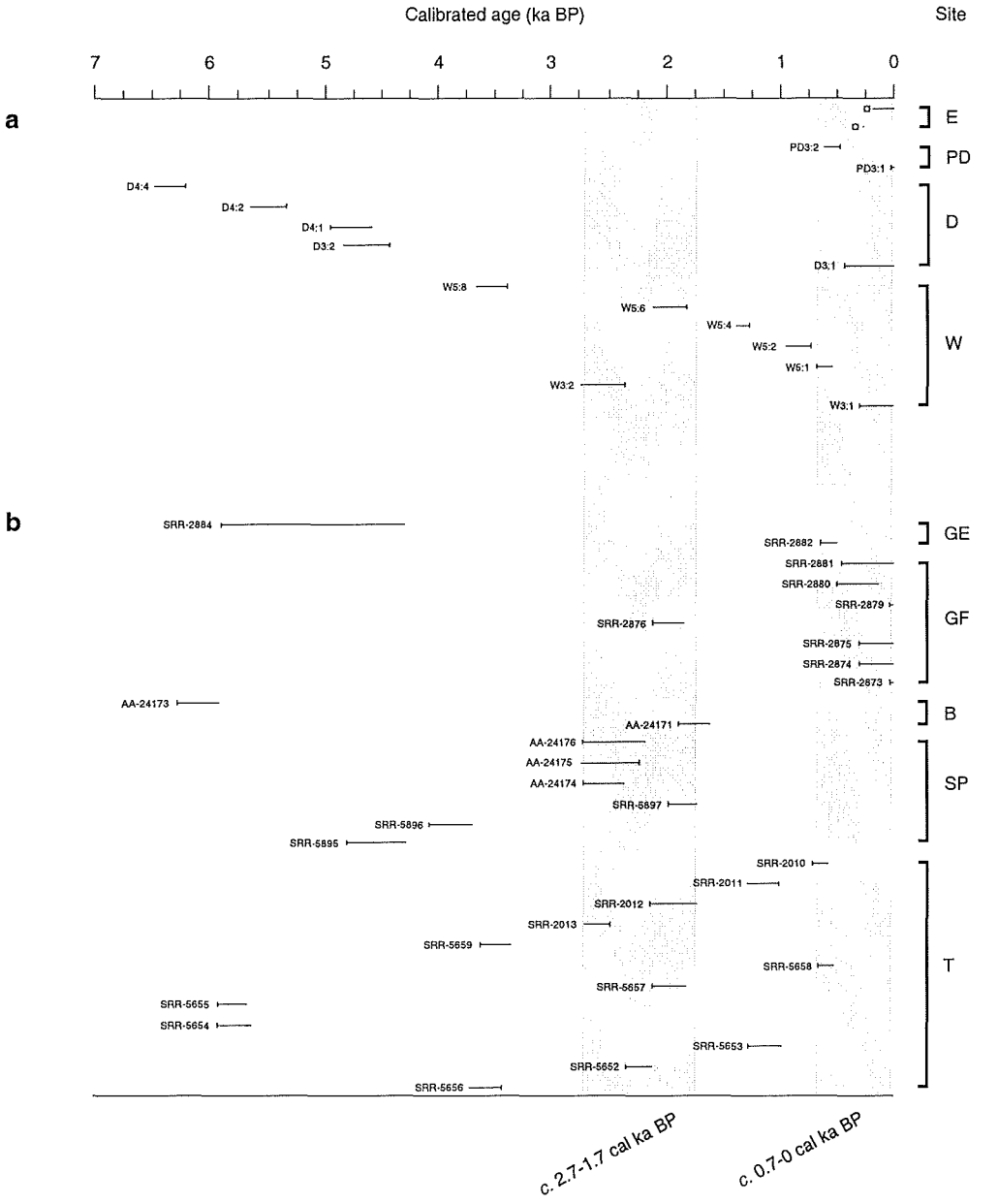
slope, thus sounding a cautionary note for the reconstruction of drift reworking from internal structure evident at only a limited number of sections. Bearing these caveats in mind, and avoiding inferences based on dates from the tops of organic horizons that may have experienced erosion, the dating evidence for section W5 in the Western Red Hills of Skye suggest that episodes of slope instability occurred immediately prior to 3300±50 yr BP (3669-3384 cal yr BP), 1985± 55 yr BP (2108-1819 cal yr BP), 1410±40 yr BP (1389-1261 cal yr BP) and 940±45 yr BP (951-735 cal yr BP), and shortly after 710±45 yr BP (692-561 cal yr BP). The dating evidence from section W3 indicates sediment emplacement immediately prior to 2540±45 yr BP (2749-2362 cal yr BP) and sometime after 175±40 yr BP (297-0 cal yr BP); the stratigraphy at this site indicates also several intervening episodes of slope instability and sediment reworking, but no dates are available for these. The radiocarbon ages from the base and top of the thick and apparently intact buried peat at section D3 in Glen Docherty imply deposition of debris flow deposits immediately prior to 4155±50 yr BP (4839-4440 cal yr BP), and a further episode of sediment deposition at c. 270±40 yr BP (431-0 cal yr BP). Those from section D4 imply sediment deposition immediately prior to 5580±65 yr BP (6491-6209 cal yr BP) and 4790±50 yr BP (5644-5327 cal yr BP), and at *c.* 4270±45 yr BP (4956-4591 cal yr BP). The two dates obtained from the base and top of the truncated organic horizon in section PD3 in the Drumochter Pass imply an episode of slope instability and sediment deposition immediately prior to 505±45 yr BP (621-475 cal yr BP), and a single period of recent debris flow activity within the past few decades.

Reworking of steep drift slopes in the Scottish Highlands

To determine the synchroneity or otherwise of these depositional events, calendar age ranges were plotted for fourteen radiocarbon dates considered to be reliable maximal age estimates for emplacement of reworked debris and

reasonably accurate minimal age estimates for the onset of renewed slope stability (Figure 5.15a). Two estimated lichenometric ages of debris cone deposits in Glen Einich (Innes, 1982) are also included. Initial inspection of these data suggests asynchronous drift slope reworking at the four sites, with barely any overlap in the age ranges. This pattern suggests that delayed or renewed paraglacial modification of drift slopes at the Scottish field sites represents randomly-occurring instability throughout the Late Holocene.

A fuller picture emerges, however, by combining data from the Scottish field sites with other radiocarbon age data for reworking of glacigenic drift-mantled slopes and rockfall talus slopes by debris flow and slopewash activity elsewhere in the Scottish Highlands, yielding a total of 43 radiocarbon dates (Figure 5.15b). Although there is a general tendency for the calendar age ranges to be widely scattered, two periods of temporal clustering may nonetheless be identified, the first at *c.* 2.7-1.7 cal ka BP, and the more recent after *c.* 0.7 cal ka BP. A cluster of twelve calendar ages for the burial of palaeosols by reworked sediment and the onset of slope stabilisation in the Western Red Hills, Glen Feshie, at Baosbheinn, Stac Pollaidh and Trotternish at *c.* 2.7-1.7 cal ka BP presents an argument for enhanced reworking of hillslopes in the Scottish Highlands within this time interval. Similarly, a further fourteen calendar age ranges for emplacement of debris and the onset of pedogenesis following slope stabilisation in the Pass of Drumochter, Glen Docherty, the Western Red Hills, Glen Etive, Glen Feshie and Trotternish and two lichenometric ages of debris flow deposits in Glen Einich cluster within the period 0.7-0 cal ka BP, again suggesting a period of generally enhanced slope instability. Although there are numerous problems associated with lichenometric-dating of debris flow deposits, evidence for a second phase of enhanced hillslope reworking in Scotland at this time is further supported by Innes' (1983b) lichenometric analysis of Scottish debris flows, from which he concluded that the majority of hillslope debris flows in



D4:1 —— Maximal age for emplacement of reworked debris overlying buried palaeosol (section:sample no.)

SRR-5656 —— Minimal age for the onset of pedogenesis following slope stabilisation (laboratory code).

□ —— Lichenometric age of surficial debris flow deposits.

Possible phase of enhanced hillslope reworking in the Scottish Highlands.

Sites and sources: Western Red Hills (W); Glen Docherty (D); Glen Einich (E), Innes, 1982; Pass of Drumochter (PD); Glen Etive (GE), Brazier *et al.*, 1988; Glen Feshie (GF), Brazier and Ballantyne, 1989; Baosbheinn (B), Hinchliffe, 1998; Stac Pollaidh (SP), Hinchliffe, 1998; and Trotternish (T), Innes, 1983c; Hinchliffe, 1998.

Figure 5.15. A comparison of calendar age ranges for reworking of slopes by debris-flow and slopewash activity (a) in the four Scottish field sites and (b) elsewhere in the Scottish Highlands. ^{14}C dates were tranformed to calendar ages using the CALIB 3.0 programme of Stuiver and Reimer (1993) and the calibration data of Stuiver and Becker (1993). The calibrated age range is given as 95% confidence limits.

Scotland have probably occurred within the last 500 years, and particularly within the last 250 years. The implications of these observed patterns are outlined in section 5.6.3 below.

Timing of delayed or renewed paraglacial modification of drift: summary

In sum, analysis of eighteen radiocarbon dates in their stratigraphic context suggests that delayed or renewed paraglacial modification of valley-side drift at the Scottish field sites represents localised, intermittent activity since *c.* 6.5 cal ka BP, and implies a random pattern of debris flow and slopewash activity throughout the Late Holocene. However, when combined with a larger dataset of hillslope reworking of glacigenic drift and rockfall talus slopes throughout the Scottish Highlands, broad coincidences of timing are evident at *c.* 2.7 to 1.7 cal ka BP and after *c.* 0.7 cal ka BP. Consultation of aerial photographs dating to AD 1946 indicates that all sites have experienced debris flow activity over the past 50 years, though extensive gullying and debris flow deposition relating to the most recent phase of enhanced activity clearly pre-date 1946.

5.6.3 Possible causes of delayed or renewed paraglacial modification of drift

Reworking of valley-side drift slopes by debris-flow activity at the Scottish field sites is seemingly favoured by the abundance of sediment on steep, unvegetated slopes, and the focused supply of water (section 5.5). Whilst widespread erosion and reworking of valley-side drift immediately after deglaciation is readily understood in terms of the steepness of some drift deposits, abundance of water and absence of a protective vegetation cover, the onset of delayed or renewed drift-slope failure within the past 6.5 ka, over six millennia after deglaciation, seems to imply triggering of failure by some specific destabilising event and/or progressive reduction in the shearing resistance of

valley-side drift. Possible causes of such delayed or renewed paraglacial erosion and resedimentation of valley-side drift are assessed below.

Seismic activity

Earthquakes are known to have triggered debris flows in many mountain environments (e.g. Solonenko, 1963; Grove, 1985; Kotarba, 1992; Lägerback, 1992). Although the diminishing rate of glacio-isostatic recovery during the Lateglacial and Holocene is likely to have progressively reduced the probability of high-magnitude earthquake events in the Scottish Highlands (Ballantyne *et al.*, 1998), it remains possible that even a relatively low-order seismic event occurring at a time of high pore-water pressures could have initiated slope failures and debris flows. Seismic activity therefore cannot be totally discounted as a possible trigger of delayed or renewed paraglacial drift reworking on steep drift slopes in Scotland. In this context it is notable that though earthquakes of estimated magnitude 6.5 - 7.0 are thought to have accompanied deglaciation at the end of the Loch Lomond Stade, Davenport *et al.* (1989) estimated that the Western Highlands of Scotland may have experienced magnitude 5.0 - 6.0 events as late as *c.* 3.4 cal ka BP.

Progressive pedogenesis

Another possible cause of drift-slope reworking in the Scottish Highlands after *c.* 6.5 cal ka BP is a slow reduction in the shearing resistance of valley-side soils. Innes (1982) suggested that progressive changes in the grain-size distribution of valley-side drift or regolith may render slopes more vulnerable to failure. More recently, numerical modelling of the behaviour of free-draining podzolic soils in the Scottish Highlands by Brooks and her co-workers (Brooks, 1997; Brooks *et al.,* 1993b, 1995; Brooks and Richards, 1993, 1994) implies that

failure and reworking of valley-side drifts during the Holocene may partly reflect a gradual decline in the stability of such deposits due to progressive soil development, making drift slopes increasingly vulnerable to failure induced by extreme rainstorm events. Significantly, an increased occurrence of hillslope failures at around 2.5 ka BP is predicted by such modelling (Brooks *et al.*, 1993b). Whilst the concept of progressive decline in slope stability is plausible, it is not certain that such intrinsic changes outweigh the influence of two possible extrinsic influences, namely climate change (and particularly an increase in the magnitude of storm events) or anthropogenically-induced vegetation change.

Anthropogenic interference

Destabilisation of drift-mantled hillslopes in upland Britain has, in some instances, been attributed to anthropogenic interference with natural vegetation covers (e.g. Durno and McVean, 1959; Fairburn, 1967; McVean and Lockie, 1967; Harvey *et al.*, 1981; Harvey and Renwick, 1987; Harvey, 1992). In particular, Innes (1982, 1983b, 1983d, 1997) cited land-use changes (particularly burning and overgrazing) as possible causal factors of intensified Late Holocene debris flow activity on Scottish mountains, but his arguments are unsupported and seem unlikely to apply at locations where debris flows originate in rock gullies. However, in Glen Etive in the Western Grampians, pollen analysis and radiocarbon dating of buried soils within a paraglacial debris cone have demonstrated a causal relationship between woodland clearance and fluvial reworking of early Holocene debris flow deposits at *c.* 550 yr BP (Brazier *et al.*, 1988). Moreover, land-use changes may not be the only possible anthropogenic cause of slope instability. Several authors (e.g. Flower and Battarbee, 1983; Battarbee *et al.*, 1985) have shown that increased acidification of upland lochs during the nineteenth and early twentieth centuries AD apparently reflects industrially-acidified precipitation. Innes (in Ballantyne, 1991b) has suggested

that an associated reduction in the extent of acid-sensitive moisture-absorbing mosses on drift-mantled slopes may have enhanced infiltration rates, thus increasing the likelihood of slope failure during extreme rainstorms, though this idea has not been explored further.

The hypothesis that anthropogenic interference with upland vegetation cover is responsible for triggering erosion and reworking of drift slopes was tested using palaeoecological data derived from analyses of pollen assemblages and charcoal fragments from buried palaeosols within debris cones in the Western Red Hills, Glen Docherty and the Pass of Drumochter. Using the methods outlined in section 5.2, four pollen diagrams for buried palaeosols at sections W5, D3, D4 and PD3 (Figures 5.16-5.19 respectively) were produced for buried palaeosols within debris cones. Because of insurmountable problems during palynomorph preparation, only two levels were sampled from buried soils at section W3, and selected data for this site are outlined in Table 5.6. Analysis of soil pollen diagrams requires different techniques to those traditionally adopted for the interpretation of sub-fossil pollen extracted from peat beds or lacustrine gyttjas. Following breakdown of soil aggregates, individual palynomorphs may be displaced down-profile (Dimbleby, 1985). Consequently, high percentage frequencies of older pollen grains resistant to decomposition may accumulate at the lower levels of a soil. Additionally, elevated pollen frequency values for upper levels of a soil may reflect addition of fresh palynomorphs at the surface. A significant implication of this model is that palynomorphs of variable derivation and age may potentially be resident at any given stratigraphic level in the soil.

With the exception of the upper palaeosol (unit 3) in section D4, all the buried palaeosols sampled at the Scottish field sites exhibit relatively higher percentages and concentrations of the extremely resistant Pteropsida (monolete) indet. spores near the base. Enhanced concentrations of this spore in the lower

Figure 5.16. Percentage palynomorph, charcoal and particle-size data obtained from buried palaeosols exposed in section W5, Western Red Hills.

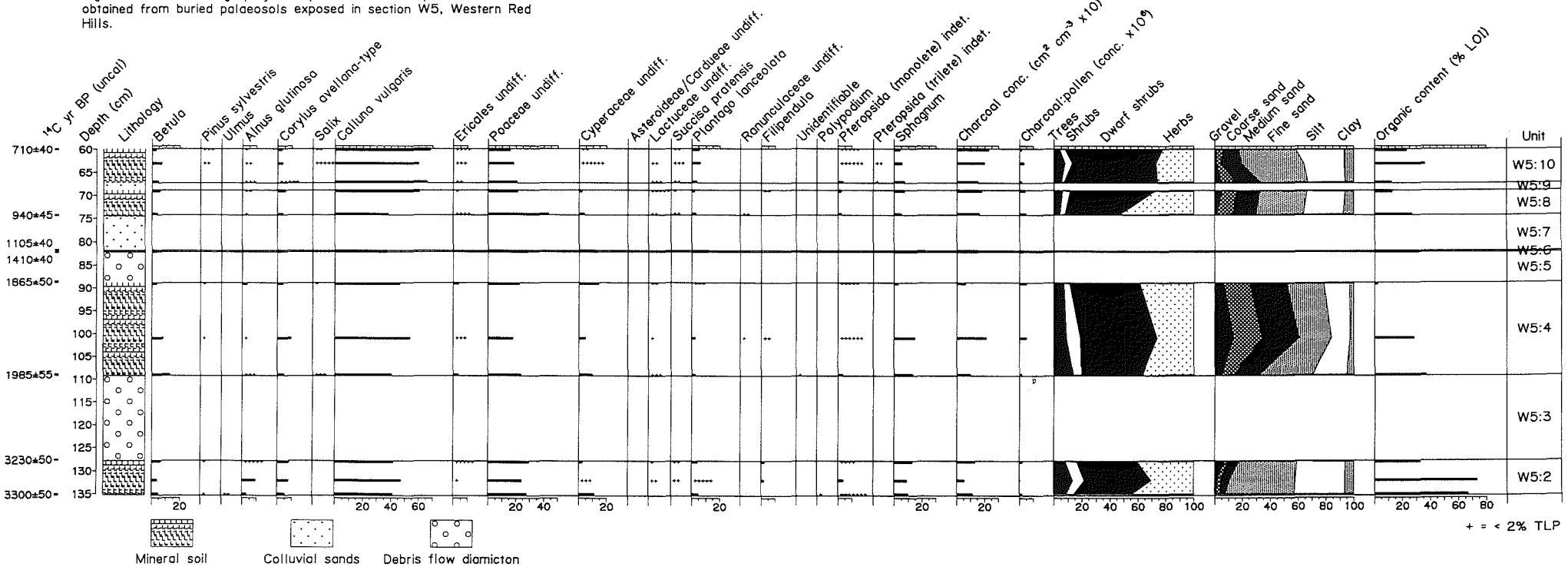


Figure 5.17. Percentage palynomorph and particle-size data obtained from a buried palaeosol exposed in section D3, Glen Docherty.

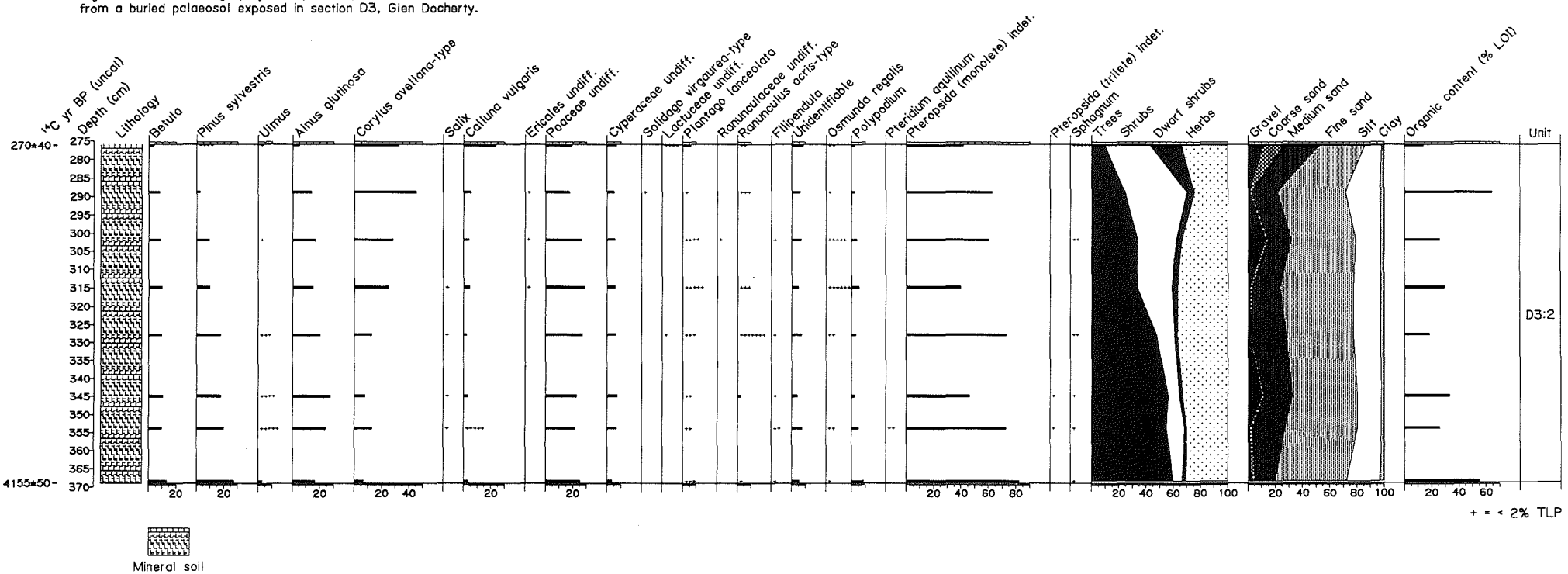


Figure 5.18. Percentage palynomorph and particle-size data obtained from buried palaeosols exposed in section D4, Glen Docherty.

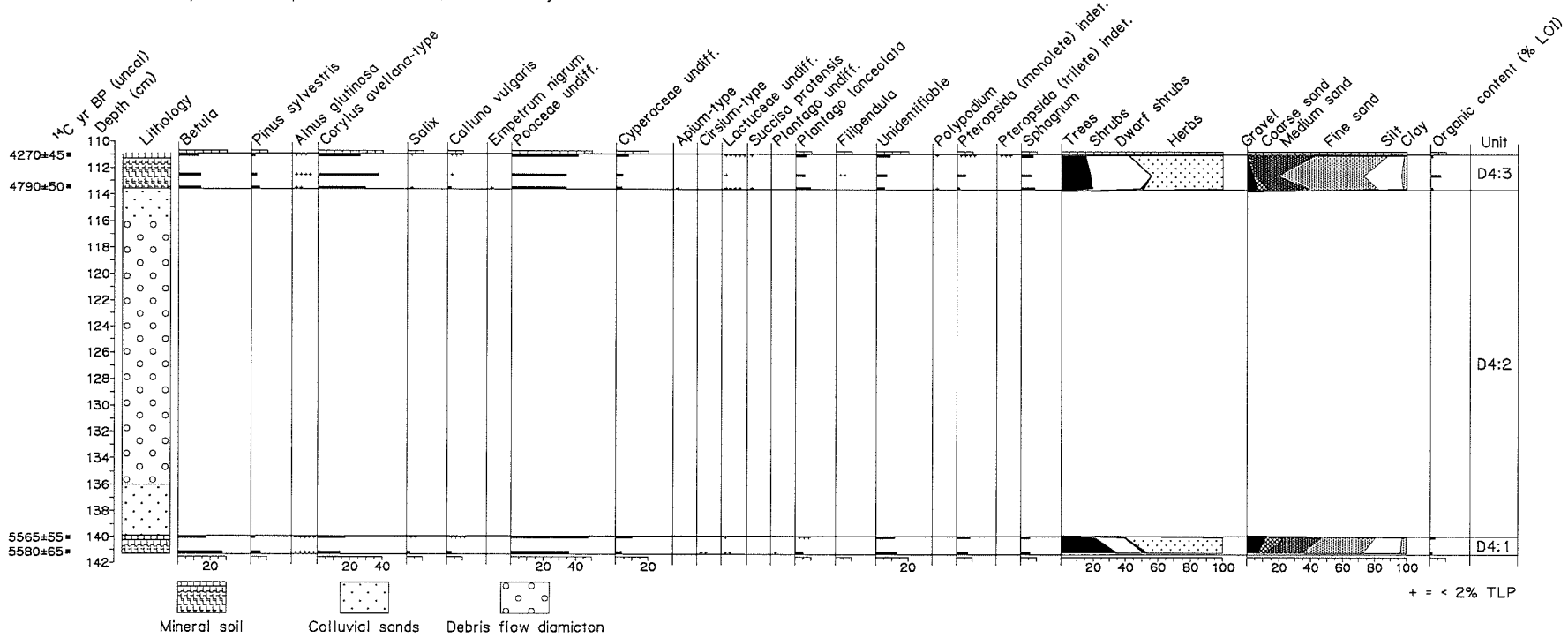
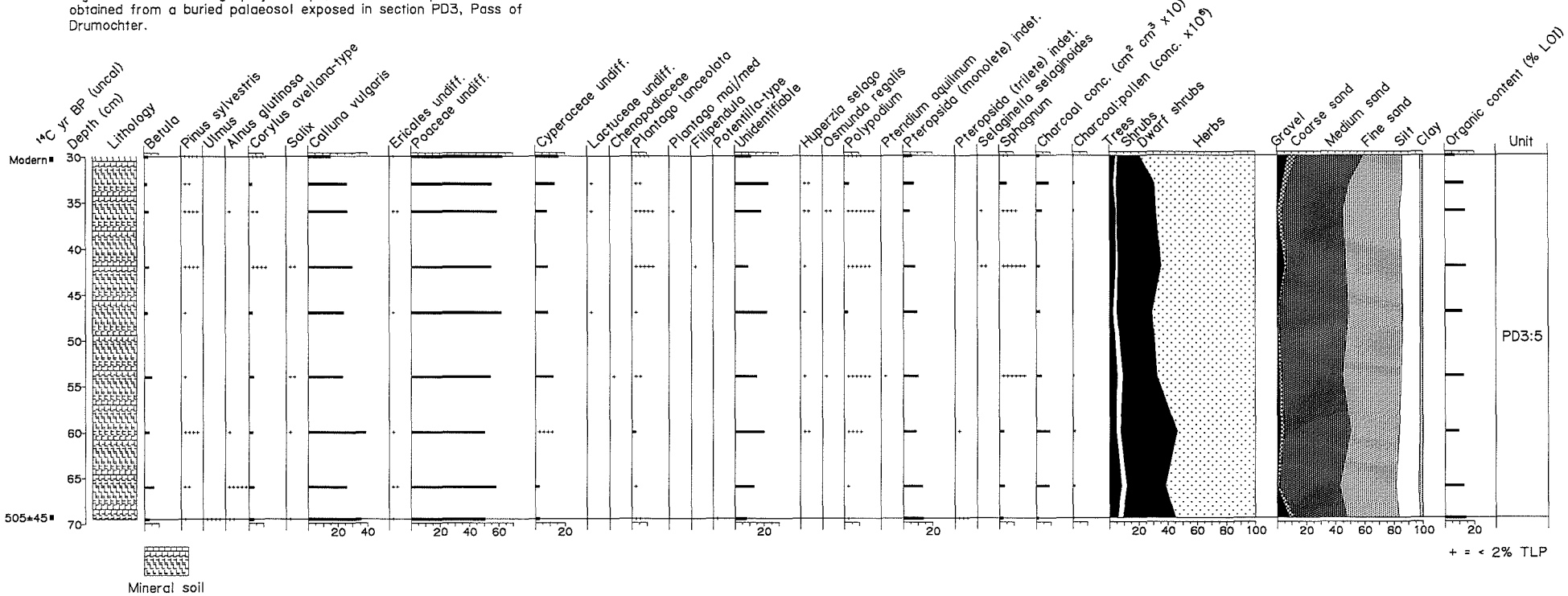


Figure 5.19. Percentage palynomorph, charcoal and particle-size data obtained from a buried palaeosol exposed in section PD3, Pass of Drumochter.



Section W3	Percentage frequency	
Level (cm)	40	113
Betula	2.1	1.1
Alnus glutinosa	5.2	0.8
Corylus avellana-type	21.2	2.4
Calluna vulgaris	40.6	69.4
Poaceae	16.3	16.5
Cyperaceae	6.9	7.7
Plantago lanceolata	5.6	0
Pteropsida (monolete) indet.	4.5	0.3
Sphagnum	8.7	6.9
Total Tree	7.6	1.9
Total Shrub	21.2	2.4
Total Dwarf Shrub	42.0	70.4
Total Herb	29.2	25.3
Charcoal conc. (cm^2 cm^{-3} x10)	0.14	0.92
Charcoal: pollen (conc. x10^6)	0.36	3.97
Organic content (% LOI)	9.1	2.6

Table 5.6. Selected palaeoecological data derived from analyses of palynomorph and charcoal deposits from buried soils exposed in section W3 in the Western Red Hills. Palynomorph data are expressed as percentage of total land pollen.

levels of soils are consistent with the predicted distribution of resistant grains advanced by Dimbleby (1985). Furthermore, within individual palaeosols the highest frequencies of heavily-degraded unidentifiable pollen grains are often recorded near the lower contact (e.g. W5, unit 4, Figure 5.16; D3, unit 2, Figure 5.17; D4, unit 1, Figure 5.18). This result possibly represents mechanical damage of palynomorphs associated with down-washing and translocation, and is thus in general conformity with the Dimbleby model. These findings suggest that organic-rich layers overridden by reworked debris represent undisturbed buried palaeosols for which valid palynological examination may be undertaken, though as noted above (section 5.6.2) there is stratigraphic and radiocarbon evidence to suggest that the upper limits of some buried soil horizons have been truncated by erosion prior to burial.

Pollen data for buried palaeosols at sections W3, W5 and PD3 (Table 5.6, Figure 5.16 and Figure 5.19) exhibit relatively little variation in pollen taxa with depth and exhibit high frequencies of Poaceae (15-63%) and *Calluna vulgaris* (15-69%). The high representation of these taxa is interpreted as suggestive of a similar habitat to that currently evident at these field sites, where low heath and herbaceous swards predominate. In contrast, diagrams for palaeosols at sections D3 and D4 (Figures 5.17 and 5.18) show high frequencies of Poaceae (18-41%) and *Corylus avellana*-type pollen (6-45%), but a relative absence of *Calluna* (< 2-6% below 276 cm at section D3, < 3% at section D4). Appreciable variation in pollen taxa exists throughout unit 2 at section D3. Pollen assemblages for nine of the eleven individual buried palaeosols sampled exhibit a significant component of *Plantago lanceolata*, a taxon traditionally associated with colonisation of disturbed ground, and thus suggestive of localised reworking of drift. Moreover, a penecontemporaneous increase in percentage coarser-grained (> 710 μm) sediment and upwards decrease in organic content is evident near the tops of several buried soil horizons (e.g. W5, units 2, 8, 10, Figure 5.16; D3, unit 2, Figure 5.17; D4, unit 3, Figure 5.18; and PD3, unit 5, Figure 5.19). This characteristic is inferred to represent intermittent inwash of minerogenic sediment by minor reworking processes during peat growth or pedogenesis prior to subsequent burial under debris flow or slopewash deposits.

Arboreal taxa are more poorly represented at sections W3, W5 and PD3 (< 2-14%) than at sections D3 and D4 (10-60%). Recorded frequencies of *Pinus sylvestris* palynomorphs only clearly signify local presence (> 20%; Tipping, 1994; *cf.* Bennett, 1984) at section D3. Levels of *Betula* are generally higher at all sites (< 2-26%), possibly indicating the localised presence of birch stands. Similarly, levels of *Alnus glutinosa* counted for unit 2 at section W5 (< 2-10%) and unit 2 at section D3 (5-27%) may indicate the existence of localised alder carrs after *c.* 4.8 cal ka BP, although this is not necessarily indicative of climatic

wettening (Bennett and Birks, 1990). An upward decline in arboreal taxa is apparent towards the tops of units 2, 4 and 10 at section W5, unit 1 at section D4, and occurs throughout unit 2 at section D3, where the percentage of tree pollen declines from 60% at the base of the palaeosol to 10% at the top. As with the percentage frequencies of arboreal pollen, the recorded frequency of *Corylus* palynomorphs is much higher for buried palaeosols at sections D3 (6-45%) and D4 (13-37%) than at section W5 (< 2-10%) and at section PD3 (< 3%). In general, the paucity of arboreal and shrub pollen recorded for buried palaeosols in sections in the Western Red Hills and the Pass of Drumochter strongly suggests that peat growth or pedogenesis, and therefore emplacement of overlying sediments, postdates the main phase of anthropogenic woodland clearance in these areas (*cf.* Walker, 1975a). Although the possible causes of delayed or renewed paraglacial drift-slope reworking in Glen Einich were not assessed, Birks (1975) found no conclusive evidence for anthropogenic interference in the vegetation of this area from palaeoecological analysis of a blanket peat profile. In contrast, the coeval decline in arboreal pollen frequencies and increase in dwarf shrub frequencies at the top of unit 2 at section D3 may represent woodland coppicing (T.C. Smout, personal communication, 1998), possibly related to iron-working and charcoal-burning which took place in and around Glen Docherty from at least 400 cal yr BP (Dixon, 1886; Durno and McVean, 1959). Removal of a protective woodland canopy at this site may have enhanced the likelihood of slope failure and associated debris flows, and may thus be indirectly related to the emplacement of the reworked deposits overlying unit 2 at section D3 after *c.* 430 cal yr BP.

The possible contribution of fire to erosion of hillslope regoliths has been acknowledged by examination of constituent soil charcoal particles by several researchers (e.g. Durno and McVean, 1959; Innes, 1982, 1983b; Parret, 1987; Brazier *et al.*, 1988; Wohl and Pearthree, 1991). These authors have suggested

that burning of vegetation increases the probability of subsequent erosion of substrates by increasing infiltration rates. Although widespread reworking of drift slopes apparently post-dates woodland clearance at the Western Red Hills and Drumochter field sites, it is possible that subsequent destabilisation of drift may be related to burning of vegetation. Analyses were therefore undertaken to test for the presence of microscopic charcoal particles resident in buried palaeosols. The results indicate very low charcoal concentrations of 0.14-2.28 cm^2 cm^{-3} at sections W3, W5 and PD3 (Table 5.6, Figure 5.16 and Figure 5.19), which appear sufficiently small to exclude local burning of vegetation during peat growth or soil formation (G. Whittington, personal communication, 1997). Minor fluctuations in charcoal concentration with depth, confirmed by the charcoal to pollen ration (*cf.* Swain, 1973), may therefore reflect inblowing of charcoal particles from fires elsewhere. Microscopic charcoal was absent in material sampled from unit 2 at section D3, and only traces were found at section D4.

Climatic causation

Several workers (e.g. Harvey and Renwick, 1987; Brazier and Ballantyne, 1989; Hinchliffe, 1998) have suggested that Late Holocene slope instability in upland Britain might be related to phases of climatic deterioration, in particular those that occurred following the onset of the sub-Atlantic period at *c.* 2.5 ka BP (*c.* 2.7-2.3 cal ka BP), and also during the 'Little Ice Age' of the sixteenth to nineteenth centuries AD (Lamb, 1977, 1979, 1982). Similar conclusions are expressed in a plethora of studies documenting upland erosion within Northwest Europe (e.g. Grove, 1972; Grove and Battagel, 1983; Kotarba and Strömquist, 1984; Starkel, 1984; Innes, 1985a; Matthews *et al.*, 1986; Rapp and Nyberg, 1988; Nesje *et al.*, 1989, 1994; Jonasson, 1991, 1993; Blikra and Nemec, 1993; Blikra, 1994; Alexandrowicz, 1997; Kotarba and Baumgart-Kotarba, 1997). However, the view that climatic deterioration is responsible for triggering episodes of

enhanced debris flow activity rests partly on apparent coincidences in timing, and partly on the untested assumption that 'climatic deterioration' is widespread, synchronous and automatically associated with renewed or enhanced landscape instability in upland areas. Nevertheless, there is an impressive body of evidence for contemporaneous intensification of mass movement activity and climatic deterioration throughout Northwest Europe.

It is possible that the sub-Atlantic climatic deterioration and the 'Little Ice Age' were characterised in Britain by an increase in the frequency of high-magnitude storm events (Lamb, 1979, 1985; Whittington, 1985; Ballantyne, 1991b). In upland Britain, most recent debris flows have been initiated by slope failure in the form of shallow landslides triggered by rapid rises in pore-water pressure associated with intense rainstorms (e.g. Baird and Lewis, 1957; Harvey, 1986, 1996; Addison, 1987; Carling, 1987; Wells and Harvey, 1987). Exceptionally high magnitude rainstorms during the early sub-Atlantic climatic deterioration and the 'Little Ice Age' may have initiated a general intensification of localised debris flow activity by triggering slope failure and lowering the threshold for subsequent events through the removal of vegetation cover (Brazier and Ballantyne, 1989). This idea suggests that the significance of periods of climatic deterioration may have been in increasing the probability of very high-magnitude rainstorm events that rendered slopes vulnerable to further erosion during subsequent storms.

The distribution of approximate ages for reworking of drift deposits at the four Scottish field sites is illustrated in Figure 5.15a. Hypothetically, a general deterioration in regional climate, possibly associated with a marked increase in the frequency of destructive storms, may be represented by a clustering of dates for episodes of sediment reworking at different sites. Infrequent violent rainstorms of random occurrence may be more likely to trigger mass movements on individual

slopes and be represented by apparently unrelated reworking events at different locations. As was noted earlier (section 5.6.2), the timing evidence from the four Scottish field sites displays an apparently random distribution of ages for delayed or renewed paraglacial reworking of valley-side drift, suggesting that these reflect randomly-spaced storm events. However, within the larger dataset for all radiocarbon-dated slope failures in the Scottish Highlands (Figure 5.15b), the calibrated radiocarbon ages of samples W3:1, W3:2, W5:1, W5:6, D3:1, PD3:1 and PD3:2 together with the Glen Einich lichen ages, all fall within the two periods of apparently enhanced reworking of sediment identified above, at *c.* 2.7 to 1.7 cal ka BP and after *c.* 0.7 cal ka BP. The former of these exhibits a correspondence in timing with climatic deterioration at *c.* 2.7-2.3 cal ka BP at the beginning of the sub-Atlantic chronozone, suggesting that at least the episode of reworking responsible for the emplacement of unit 1 at section W3 and unit 3 at section W5 may have been associated with regional climatic deterioration. The second phase of enhanced slope reworking after *c.* 0.7 cal ka BP partly corresponds with the 'Little Ice Age' of the sixteenth to nineteenth centuries, and may represent a link between a phase of climatic deterioration and drift slope reworking in each of the four sites, particularly when the broad resolution of individual radiocarbon ages is taken into account. A climatic interpretation of the most recent phase of enhanced activity requires caution, however, as more debris flows have been recorded (and are more likely to have been recorded) during the last few centuries than for earlier periods (Berrisford and Matthews, 1997). The possibility that the distribution of calendar ages evident in Figure 5.15 simply reflects severe local rainstorms that are unrelated to long-term climatic change cannot be dismissed, given the relatively small number of radiocarbon dates available for a timescale of *c.* 6.5 ka. This alternative interpretation is supported by seven calibrated radiocarbon age ranges obtained for samples from the Scottish field sites (samples W5:2, W5:4, W5:8, D3:2, D4:1, D4:2 and D4:4) that do not

exhibit a correspondence in timing with proposed periods of increased wetness or storminess.

Possible causes of delayed or renewed paraglacial modification of drift: summary

In sum, Late Holocene reworking of drift at the Scottish field sites appears unrelated to burning of vegetation, though debris flow activity since *c.* 0.4 cal ka BP in Glen Docherty may have been associated with the prior removal of woodland cover. Further, the extent to which the most recent phase of activity at all sites reflects overgrazing, degradation of hillslope moss cover or a gradual decline in slope stability due to progressive pedogenesis remains hypothetical and awaits further research. However, periods of particularly severe weather characterised by an increase in the frequency of exceptional storm events during the 'Little Ice Age' and at the onset of the sub-Atlantic chronozone may represent a possible cause of delayed or renewed gully erosion of valley-side drift at the Scottish field sites. Nevertheless, the possibility that such activity was initiated by destructive storms of random occurrence cannot be discounted.

5.6.4 Timing and possible causes of delayed or renewed paraglacial modification of drift: implications

Although caution is necessary in interpreting the wider implications of the findings reported above, three important implications emerge from this study. First, these findings suggest that drift-mantled slopes in the Scottish Highlands are sensitive to climatically-induced instability, possibly resulting from changing climate. Consequently, further dating of buried peats, palaeosols and depositional structures underlying drift slopes may ultimately yield a useful proxy record for reconstructing former climatic changes (*cf.* Matthews *et al.*, 1993; Berrisford and Matthews, 1997). Evidence for erosion and reworking of drift in the four field

sites during the latter half of the Holocene accords with findings from a growing database of episodes of regional reworking of drift and talus slopes in Scotland during recent millennia, and supports the theory of increased rates of landscape change in upland Britain during the Late Holocene, resulting in modification of both fluvial and hillslope landsystems (e.g. Strachan, 1976; Richards, 1981; Harvey *et al.*, 1981, 1984; Ballantyne, 1986a, 1986c; Macklin and Lewin, 1986; Harvey and Renwick, 1987; Brazier *et al.*, 1988; Brazier and Ballantyne, 1989; Tipping, 1995; Hinchliffe *et al.*, 1998).

Secondly, the delayed or renewed paraglacial slope responses observed at the four Scottish sites several thousands of years after deglaciation demonstrate that while the paraglacial period theoretically ends once sediment yields drop to rates typical of unglaciated catchments (Church and Slaymaker, 1989), it is difficult to ascertain whether or not a landscape really fully adjusts following a glacial phase (Benn and Evans, 1998). The legacy of Late Pleistocene glacial and deglacial conditioning represents an extremely powerful control on Late Holocene geomorphic activity in the Scottish Highlands, and exemplifies the persistence of landscape 'memory' in deglaciated upland environments.

A third implication concerns whether Holocene reworking of drift slopes at the Scottish field sites constitutes a delayed paraglacial response, or renewed paraglacial activity. Although there is no radiocarbon dating evidence for a Late Pleistocene pulse of paraglacial activity at these locations, Benn (1990) found that in the Western Red Hills, the maximum slope angle on which Lateglacial moraines are preserved is *c.* 20°, and sedimentological evidence indicates that on steeper slopes in this area, paraglacial reworking occurred during or shortly after deglaciation. At least localised paraglacial reworking of drift is inferred to have accompanied deglaciation elsewhere in the Scottish Highlands (e.g. Peacock, 1986; Brazier *et al.*, 1988; Auton, 1990; Ballantyne and Benn, 1996).

Consequently, it is possible that Holocene drift reworking at some of the field sites examined in this chapter may represent renewed rather than delayed paraglacial activity. If this conjecture is valid, it implies that, following Late Pleistocene paraglacial activity, some slopes remained in a metastable state prior to rejuvenation and renewed incision and drift reworking several millennia later. Alternatively, paraglacial modification of drift may have been delayed at some localities until failure was triggered during the Mid-Late Holocene.

This final consideration leads to the intriguing question of the absence of dating evidence for slope failure and reworking of drift at any site prior to *c.* 6.5 cal ka BP, some six millennia after deglaciation at the end of the Loch Lomond Stade. This temporal hiatus may be a figment of sampling, in that older buried organic horizons are less likely to be exposed in section than younger and stratigraphically higher sections. Alternatively, it may reflect slow reduction in the shearing resistance of sediments mantling steep hillslopes, possibly due to progressive pedogenesis, as suggested by Brooks and her co-workers (Brooks, 1997; Brooks *et al.,* 1993b, 1995; Brooks and Richards, 1993, 1994). A further possibility is that it reflects a climatic signal, possibly indicating that the magnitude and/or frequency of extreme rainstorm events was less in the Early Holocene than in the Mid and Late Holocene. These speculations can only be resolved, however, by future investigation of sites where reworked drift containing intercalated organic horizons directly overlies demonstrably *in situ* till relating to final deglaciation at the end of the Loch Lomond Stade.

5.7 Summary.

The research reported in this chapter provides an insight into the behaviour of drift slopes at four sites in Scotland. The following summary synthesises the

principal findings concerning the extent, nature, constraints and timing of delayed or renewed paraglacial activity at these sites.

1. At the Scottish field sites gully density ranges from 1 gully per kilometre at those sites exhibiting very limited reworking to 25 gullies per kilometre where delayed or renewed paraglacial modification of hillslope drift has been most active. Spatially, the proportion of hillslope drift that has been remobilised by such activity ranges from *c.* 1% in the Pass of Drumochter to *c.* 7% in the Western Red Hills of Skye.

2. Debris flow is the dominant process of paraglacial sediment transfer at the Scottish field sites. Intense rainstorms are the most likely triggers of sediment failure and flow. Snow avalanching is extremely limited at these sites, though slopewash is recognised as a secondary paraglacial process.

3. At the Scottish field sites, gullying is largely present only where drift is steeper than *c.* 30° and thicker than *c.* 3 m, and where the void ratio of unreworked sediment exceeds *c.* 0.4. Both the amount of precipitation and focusing of water delivered onto valley-side drift are seemingly instrumental in conditioning gullying intensity, and may determine the extent of delayed or renewed paraglacial reworking at sites where the intrinsic constraints on drift reworking are satisfied.

4. Analysis of eighteen radiocarbon dates suggests that delayed or renewed paraglacial modification of valley-side drift at the Scottish field sites represents localised, intermittent activity since *c.* 6.5 cal ka BP. When combined with a broader dataset of hillslope reworking throughout the Scottish Highlands, however, broad coincidences of timing are evident at *c.* 2.7 to 1.7 cal ka BP and after *c.* 0.7 cal ka BP. In combination with

analysis of sub-fossil pollen and charcoal, this chronology suggests that recent reworking of ancient drifts may partly reflect the influence of extreme rainfall events during periods of climatic deterioration at *c.* 2.7 to 2.3 cal ka BP and after *c.* 0.3 cal ka BP, and at one site, anthropogenic interference with the vegetation cover in recent centuries. However, the possibility that such delayed or renewed activity was initiated by destructive storms of random occurrence cannot yet be dismissed.

Chapter 6

Paraglacial modification of glacigenic sediments

6.1 Introduction.

The research already reported in this thesis has considered the extent, nature, constraints and timing of paraglacial modification of drift at sites in Norway and Scotland. Attention is now turned to the implications of paraglacial reworking of hillslope drift deposits. As was established in sections 4.4 and 5.4, the dominant agent of reworking of *in situ* glacigenic sediment at most of the field sites investigated has been debris flow activity. This chapter outlines the sedimentological consequences of such activity. Following a brief description of field and laboratory methods (section 6.2), it aims first to establish sedimentological criteria that distinguish glacigenic drift deposits reworked by debris flows from *in situ* glacigenic drift deposits (section 6.3), and secondly to apply such criteria to the interpretation of deposits of uncertain (reworked or *in situ*) origin (section 6.4). The process and stratigraphic implications of these results are then explored in section 6.5, and summary findings outlined in section 6.6.

Several researchers have noted the difficulty of distinguishing *in situ* drift deposits from those reworked by debris flows, largely because the latter tend to retain many of the sedimentological characteristics of the unreworked parent sediments (Eyles *et al.*, 1988; Eyles and Kocsis, 1988; Owen and Derbyshire, 1989; Derbyshire and Owen, 1990; Owen, 1991, 1994). Furthermore, as deposition of *in situ* drift can occur supraglacially by the melt-out and sublimation of debris-rich ice, or subglacially by a combination of lodgement, deposition from

a deforming layer, and melt-out (Benn and Evans, 1998), *in situ* drift deposits are enormously variable in sedimentological character (Boulton and Dent, 1974). Though there is some evidence for granulometric and micromorphologic differences between unmodified drift deposits and paraglacially-reworked drift (e.g. Landim and Frakes, 1968; Owen and Derbyshire, 1989; Owen, 1991, 1994), other studies suggest that paraglacially-reworked deposits are distinguishable from *in situ* deposits only in terms of clast fabric and large-scale structures (e.g. Ballantyne and Benn, 1994). Particle alignment in debris flow deposits is variable, but tends to show a preferred downslope or down-flow orientation in zones of intense shearing (Lindsay, 1968; Lawson, 1979a; Innes, 1983a; Mills, 1984), whereas unmodified clasts in basal tills are often strongly aligned down-valley parallel to ice flow, with *a-b* planes tending to have a gentle upglacier imbrication (Dowdeswell and Sharp, 1986; Benn, 1994b; Hart, 1994; Krüger, 1994; Benn and Evans, 1998). It should be remembered, however, that the direction of shear in a subglacial deforming layer is governed by local stress gradients, which may diverge markedly from the flow of the overlying ice (Benn, 1994b).

Subglacial diamictons may exhibit shear banding, but this is less persistent and less regular than in debris flow deposits (Owen, 1991). Overconsolidation induced by ice overburden is often present in *in situ* tills, but is commonly associated with sub-horizontal dilation joints and associated fissility, features not recognised in debris flow deposits, produced by the unloading of the ice. Whilst deformation tills may contain interbeds and lenses of sorted sediments (e.g. Eyles *et al.*, 1982; Evans *et al.*, 1995; Benn and Evans, 1996) that represent infills of former braided canal systems developed at the ice-till interface (Clark and Walder, 1994; Walder and Fowler, 1994), these commonly possess diagnostic characteristics such as concave-up lower contacts, nearly planar upper contacts and ripple cross stratification. In contrast, whilst debris flow deposits can appear

totally chaotic at first acquaintance, close examination often reveals a surprising amount of internal organisation (Benn and Evans, 1998) that may distinguish them from many unreworked tills. Individual debris flow units often take the form of massive, ungraded diamictons, with clasts of various sizes embedded in a matrix of sand- or silt-rich fines. Sedimentological characteristics may include lenticular structures visible in transverse sections, the alternation of clast-supported or matrix-supported units, and localised inverse grading. Movement may cause shearing or deformation of underlying sediments, particularly in the case of high-density, low-viscosity flows (Lawson, 1988), and individual contacts may be defined by discontinuities or thin beds of silt or sand that reflect surface wash following flow immobilisation (Brazier and Ballantyne, 1989; Nieuwenhuijzen and van Steijn, 1990). The characteristics of debris flow deposits, however, vary according to flow type (Lawson, 1982) and sample-point location (e.g. Nieuwenhuijzen and van Steijn, 1990; Bertran and Texier, 1994; van Steijn *et al.*, 1995), and internal structure can vary greatly with water content during flow and deposition.

6.2 Methods.

Mapping of both the Norwegian (chapter 4) and Scottish field sites (chapter 5) revealed natural sections along the sides of gullies, in debris cones and within *in situ* till on valley floors in which glacigenic drift facies are exposed. Sections were investigated in both deposits reworked by debris flows (fresh debris flow lobes) and within valley-floor *in situ* drift deposits to establish the sedimentological characteristics that distinguish *in situ* and reworked drift. Such criteria were then applied to the analysis of valley-side drift deposits of uncertain post-depositional history. At some of the gully sections investigated, depositional facies were obscured by gully wall collapse, and it was necessary to excavate

vertical sections to permit investigation of intact stratigraphies. All sections were logged in detail on graph paper and photographed.

At each section, the orientation and dip of 50 elongate (*a:b* axial ratio > 1.5) clasts greater than 20 mm in length were measured, and for each clast fabric the three orthogonal eigenvectors (V_1, V_2 and V_3) and their respective eigenvalues (S_1, S_2 and S_3) were calculated according to the method of Mark (1973). Eigenvector V_1 identifies the direction of maximum clustering, and eigenvalue S_1 represents the strength of clustering around V_1. In addition, samples of 50 randomly-selected clasts removed from each section were investigated in terms of clast shape, angularity and texture. Long (*a*), intermediate (*b*) and short (*c*) axes were measured, *c:a* flatness indices calculated (Sneed and Folk, 1958) and shape summarised in terms of C_{40} and C_{50} indices, which respectively measure the percentage of clasts with *c:a* ratios ≤ 0.4 and ≤ 0.5 (Ballantyne, 1982). Angularity of clasts was assessed in terms of a six-point scale (very angular, angular, sub-angular, sub-rounded, rounded and well-rounded) and expressed as the RA index, which expresses the proportion of very angular plus angular clasts as a percentage of the total sample (Benn and Ballantyne, 1994). Clast texture was assessed in terms of the presence or absence of chips and facets on sampled clasts. The granulometry of samples weighing *c.* 500 g of both *in situ* and reworked fine (< 2 mm) sediments was established by dry-sieving material through the range 2000 μm to 710 μm and analysing the fraction < 710 μm using a Coulter LS100 laser granulometer. Intact samples of matrix material were also removed in steel cylinders and were dried, weighed, saturated and re-weighed to calculate void ratio as a measure of sediment packing (Attewell and Farmer, 1976).

6.3 Comparative sedimentological characteristics of *in situ* and reworked drift.

6.3.1 Clast fabric

Measurements of the orientation of 50 clasts were made at 8 localities on the forelands of Fåbergstølsbreen and Lodalsbreen and 14 sites in the Western Red Hills and Glen Einich to determine possible contrasts between the macrofabric characteristics of *in situ* till and those of glacigenic sediments reworked by debris flow activity. All fabrics reveal preferred clast orientation in the form of a girdle around the azimuth of the principal eigenvector V_1 (θ in Figures 6.1 and 6.2). However, the preferred orientation of clasts in all of the *in situ* drift samples (IS 1-8; Figure 6.1) tends to be aligned parallel or sub-parallel to the valley axis, whereas that of all debris flow-reworked samples (DF 1-14; Figure 6.2) exhibits a preferred down-flow alignment. This difference reflects contrasting modes of sediment deposition, with elongate clasts in basal till adopting an alignment of least resistance parallel to glacier movement (Lindsay, 1970; Benn, 1994a, 1995; Hart, 1994; Benn and Evans, 1996), whereas clasts emplaced by debris flows often parallel the direction of flow (Lindsay, 1968; Boulton, 1971), and hence show no relationship to regional ice flow directions (Lawson, 1979a, 1979b). This result therefore supports previous research (Ballantyne and Benn, 1994) which suggests that clast orientation measurements provide a valid means of distinguishing paraglacial deposits reworked by debris flows from *in situ* drift deposits. The dip values identified by the principal eigenvector (V_1), however, appear to be less useful in discriminating the two types of deposit, as those for *in situ* till samples (1° - 33°) overlap those for debris flow samples (5° - 32°) (*cf.* Eyles and Kocsis, 1988; Ballantyne and Benn, 1994).

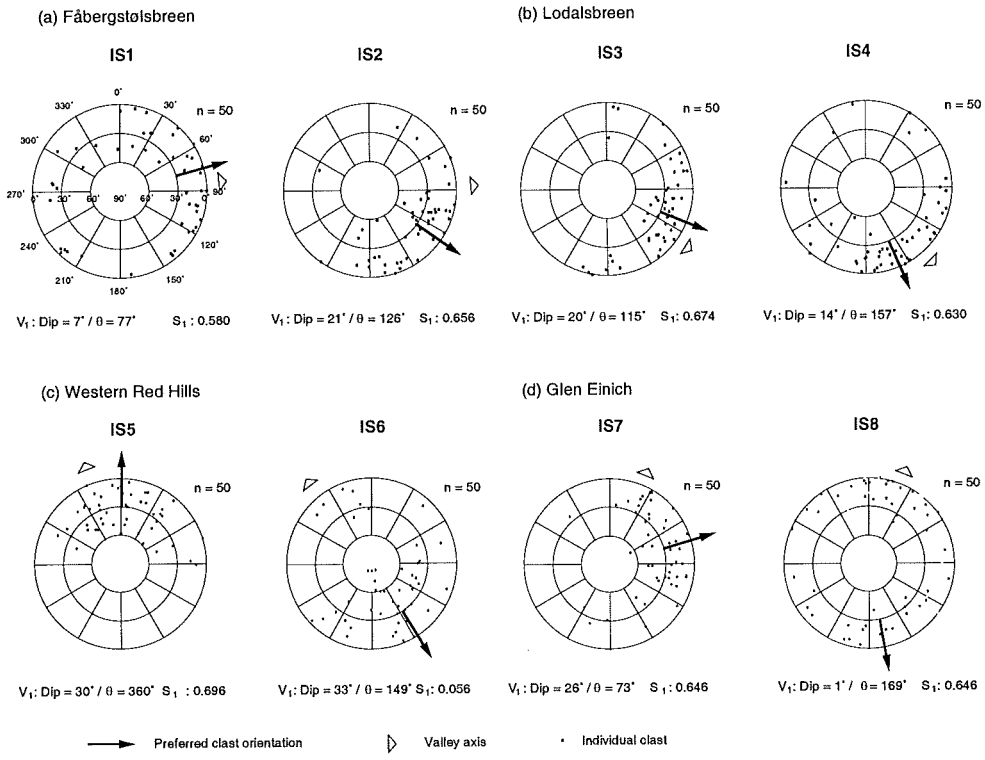


Figure 6.1. Stereographic fabric plots for samples of 50 elongate clasts removed from *in situ* drift deposits (IS 1-8) exposed at Fåbergstølsbreen and Lodalsbreen, and in the Western Red Hills and Glen Einich. θ represents the preferred orientation of the long axes of clasts as identified by the principal eigenvector, V_1.

Possible contrasts in fabric shape or 'type' were assessed by calculating for each fabric an isotropy index ($I = S_3/S_1$) and elongation index ($E = 1 - S_2/S_3$) and plotting the results on a fabric shape triangle (Benn, 1994b; Figure 6.3). The results show that though the *in situ* till samples tend to plot farther from the 'cluster' apex of the diagram than the reworked drift samples, there is overlap between the two sets of plots, which share low isotropy and moderate to high elongation. Whilst the basal till samples and debris flow sediments differ in mode of deposition and clast orientation, both experience rotating of clasts by their respective deforming matrices, and consequently share a common fabric type. This research indicates that fabric type as determined by eigenvector data has

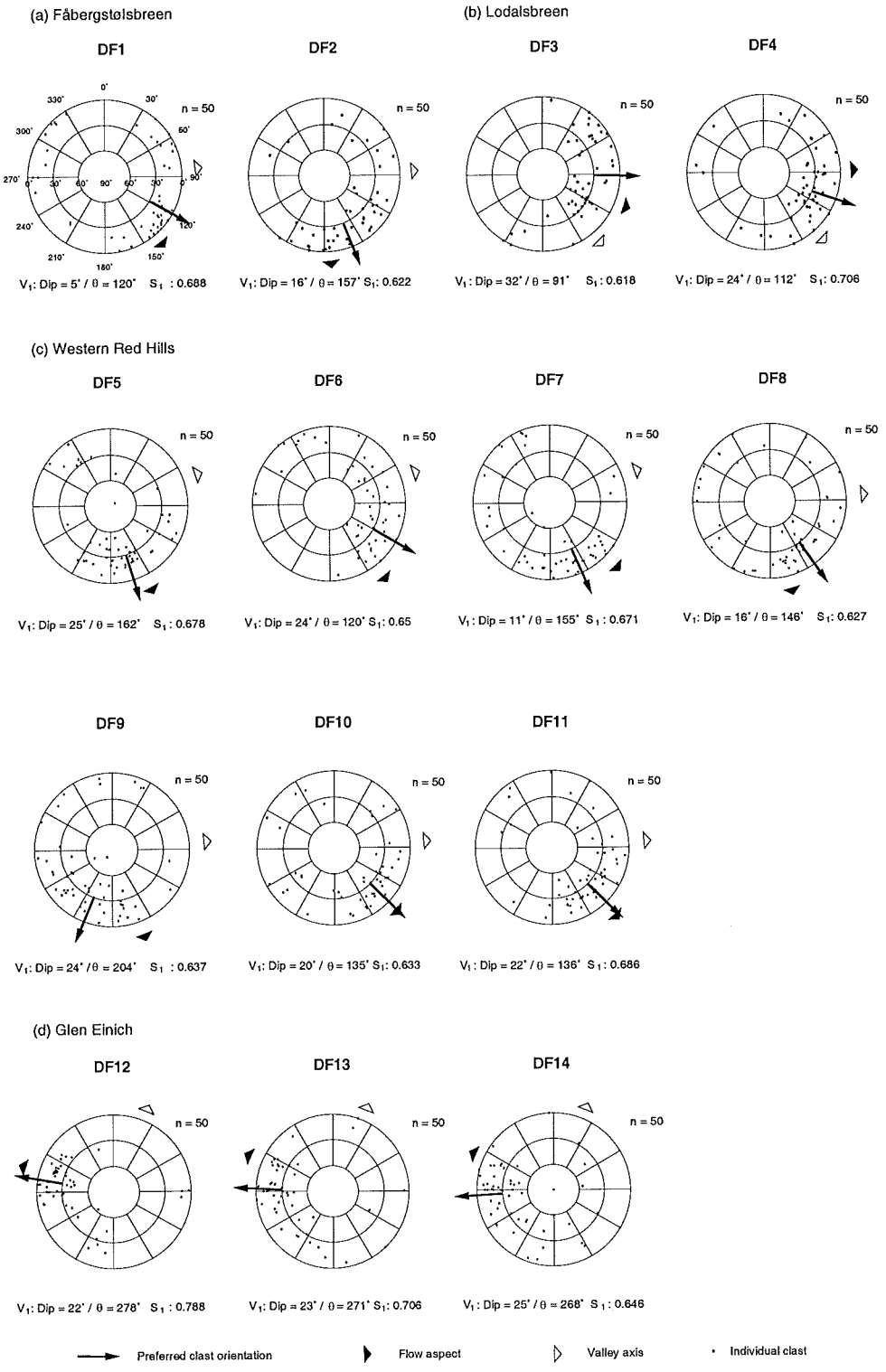


Figure 6.2. Stereographic fabric plots for samples of 50 elongate clasts removed from debris flow deposits (DF 1-14) exposed at Fåbergstølsbreen and Lodalsbreen, and in the Western Red Hills and Glen Einich. θ represents the preferred orientation of the long axes of clasts as identified by the principal eigenvector, V_1.

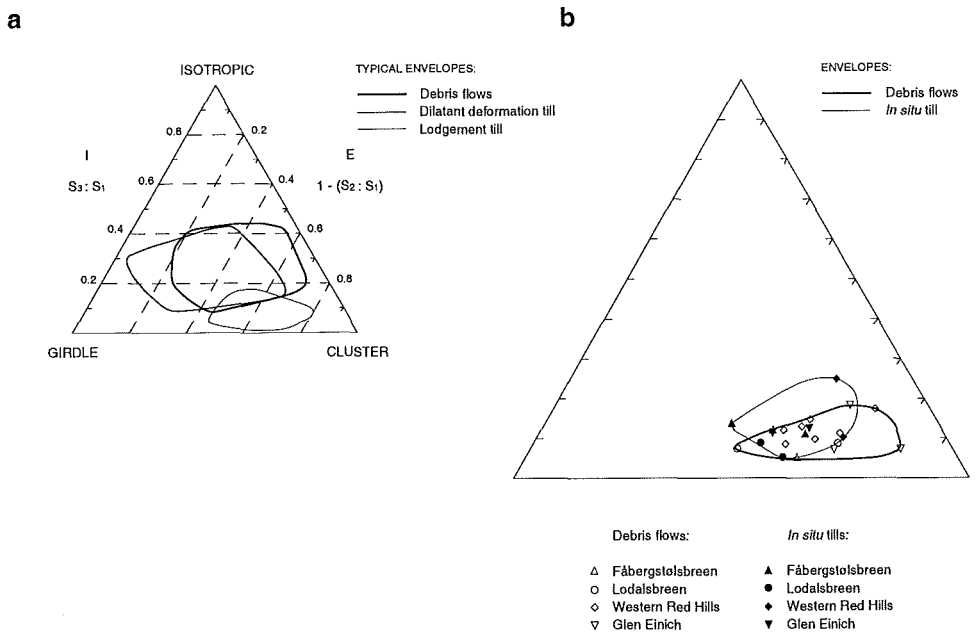


Figure 6.3. Fabric shape represented on triangular diagrams using ratios between the eigenvalues S_3, S_2 and S_1. (a) The continuum of fabric types scaled using the isotropy ($I = S_3/S_1$) and elongation ($E = 1 - (S_1/S_2)$) indices. Type envelopes based on Ballantyne and Benn, 1994; Benn, 1994a. (b) Fabric type for the samples removed from debris flow deposits (DF 1-14) and *in situ* drift deposits (IS 1-8) from Norway and Scotland.

limited value in differentiating reworked from *in situ* glacigenic sediment (*cf.* Dowdeswell *et al.*, 1985; Ballantyne and Benn, 1994).

6.3.2 Clast form

The aggregated clast shape data for the Fåbergstølsbre and Lodalsbre field sites (Figure 6.4a) reveal only limited differences between samples from *in situ* and reworked drift. However, although the C_{40} and C_{50} indices for reworked sediments overlap those for unreworked till, the distribution of C_{50} values for reworked samples exceeds that for unreworked samples at $p < 0.01$ when tested using the Mann-Whitney two sample test. This suggests that aggregate clast shape may offer a means of distinguishing reworked from *in situ* drifts, though previous studies have detected no significant clast shape differences between the

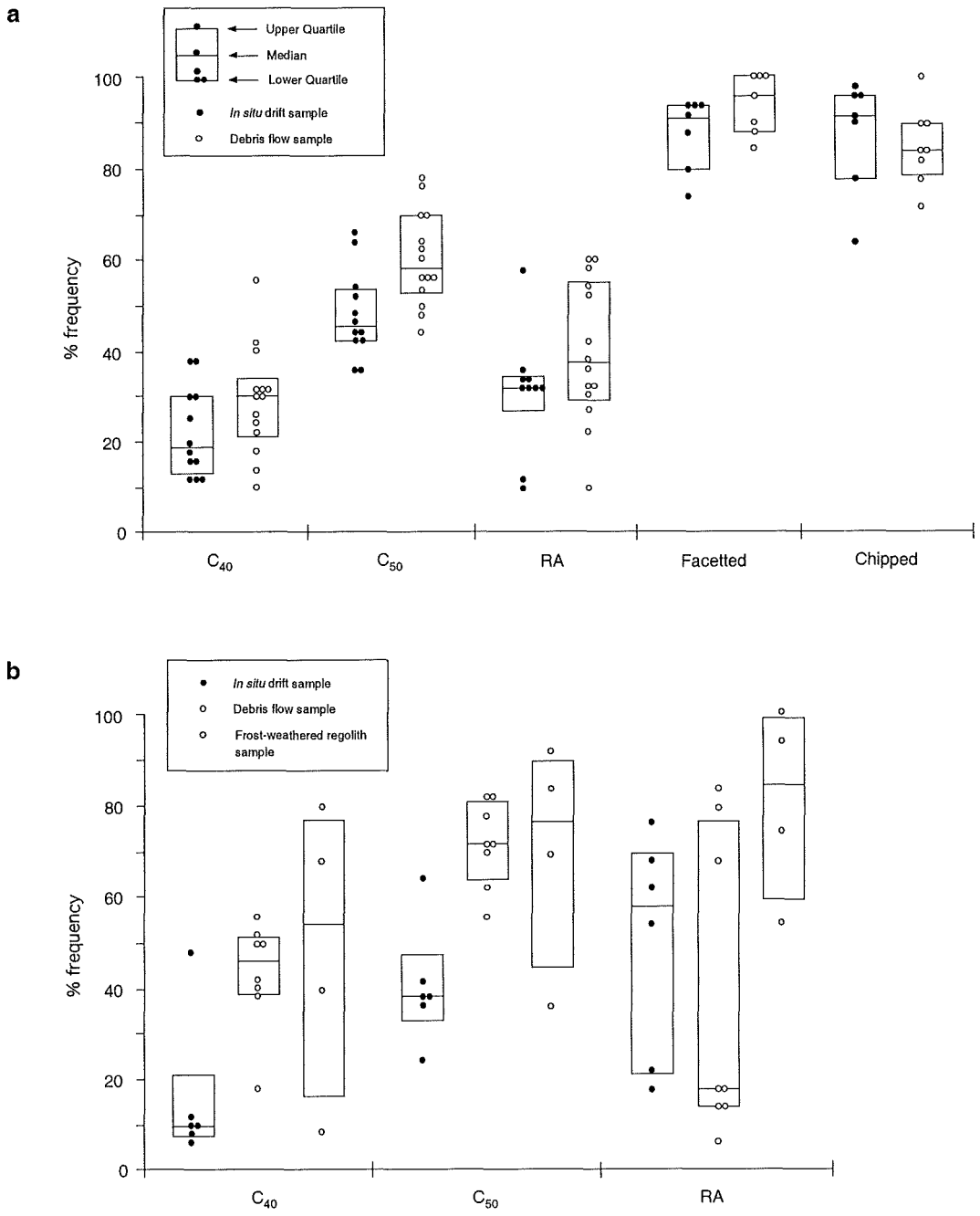


Figure 6.4. Dispersion diagrams summarising clast shape and angularity analyses on samples of 50 clasts from (a) *in situ* drift and debris flow-reworked drift deposits at Fåbergstølsbreen and Lodalsbreen, and (b) *in situ* drift, debris flow-reworked drift deposits and frost-weathered regolith in the Western Red Hills and Glen Einich. See text for definition of indices.

two (Ballantyne and Benn, 1994). The difference detected here suggests that clasts reworked by debris flows tend to be less equidimensional than those in

basal till. This difference could reflect selective comminution or fracture during flow, though this process might be expected to produce lower rather than higher C_{40} and C_{50} indices, through fracture across the principal axes of reworked clasts. An alternative explanation is that the difference reflects a higher proportion of supraglacially- or englacially-transported clasts in the valley-side drifts that form the source of the reworked sediments than in the basal till on the valley floor. Such clasts travel passively within the ice and escape modification by abrasion at the glacier sole, and hence often retain a slabbier or more elongate shape than subglacially-modified clasts (Ballantyne, 1982; Benn and Ballantyne, 1994).

This interpretation is reinforced by the aggregate angularity data, as the mean RA index (% very angular + angular clasts) is greater for the reworked sediments (39.4) than for the basal till samples (30.4), though in this case the difference is not statistically significant. Angular and very angular clasts are characteristic of passive transport within glacier ice, and thus the greater aggregate angularity of the reworked sediments may imply a higher proportion of supraglacially- or englacially-transported clasts in the parent valley-side drift than in the till on the valley floor (*cf.* Benn and Ballantyne, 1994). Alternatively, the greater aggregate angularity of the reworked clasts could indicate fracture during downslope flow. The textural data, however, suggest that this is unlikely, as the median percentage of chipped clasts in the samples of reworked sediment (85%) is actually lower than in the *in situ* till (88%), and the percentage of facetted clasts is only slightly (but not significantly) higher (median values 95% and 88% respectively). In sum, then, though significant differences in aggregate clast shape and weaker differences in aggregate angularity are detectable between *in situ* basal till on the valley floor and sediments reworked by paraglacial debris-flow activity at Fåbergstølsbreen and Lodalsbreen, there is little difference in textural characteristics. Moreover, the differences in shape and angularity probably reflect initial contrasts in the proportion of passively-transported clasts

in the valley-side and valley-floor drifts, rather than modification of clast shape during reworking.

The data from the two Scottish sites (Figure 6.4b) indicate more pronounced differences in aggregate clast shape between reworked and *in situ* samples than are evident in the data from the two Norwegian sites. Results of the Mann-Whitney two sample test demonstrate that the C_{40} and C_{50} values for reworked samples from the Western Red Hills and Glen Einich significantly exceed those for unreworked samples at $p < 0.05$ and $p < 0.01$ respectively. This difference may again indicate a higher proportion of passively-transported clasts in samples reworked by debris flows, though in this case the RA values for reworked samples completely overlap those for *in situ* till. At both Scottish sites an alternative explanation of the observed differences in aggregate clast shape is possible, as the sources of debris flows in these areas frequently extend upslope beyond the upper limit of glacigenic drift on to a cover of frost-weathered regolith. Clasts sampled from this regolith cover tend to have high C_{40} and C_{50} values (Figure 6.4b), and hence the aggregate clast shape differences between *in situ* and reworked glacigenic sediment in the Western Red Hills and Glen Einich could reflect incorporation within the latter of clasts derived from this frost-weathered regolith. However, this explanation is not supported by the aggregate angularity data, as the RA index for some samples of reworked sediments is not only much lower than that for samples of frost-weathered debris, but also lower than that on samples of *in situ* drift.

In sum, then, though significant differences in aggregate clast shape and weaker differences in aggregate angularity are detectable between *in situ* drift on the valley floor and sediments reworked by debris flow activity, there is little difference in textural characteristics. Moreover, the differences in shape and angularity probably reflect initial contrasts in the proportion of passively-

transported clasts in the valley-side and valley-floor drifts, rather than modification of clast shape during reworking, particularly given the generally low inter-particle contact forces operating during flowage.

6.3.3 Fine-fraction granulometry

A total of 82 samples of fine sediment were collected from *in situ* glacigenic deposits and sediments reworked by recent debris flows at Fåbergstølsbreen, Lodalsbreen, Leirdalen and Visdalen in Norway, and from the Western Red Hills, Glen Docherty, Glen Einich and the Pass of Drumochter in Scotland. The purpose of this work was to evaluate the suggestion that reworked deposits emplaced by debris flow may have experienced eluviation and removal of the finest grains (Ballantyne and Benn, 1994). This proposition was tested by comparing proportion by weight of clay (< 2 μm) and clay plus silt (< 63 μm) particles present in samples of fine (< 2 mm) sediment from both *in situ* and reworked glacigenic deposits. The results from the Norwegian sites (Figure 6.5) suggest no systematic pattern. At Fåbergstølsbreen there is little difference between the two groups of samples in terms of either of the above criteria, and at Lodalsbreen and Visdalen there is only slight evidence for depletion of clay and silt in reworked sediment. Moreover, the results from material sampled at Leirdalen demonstrate that percentage clay and percentage clay plus silt content in the debris flow samples significantly exceeds that for unreworked samples at $p < 0.05$ when tested using the Mann-Whitney two sample test. The results from the Scottish sites (Figure 6.6) reveal only a marginally clearer pattern. Although there is no apparent difference in fine-fraction composition between the two types of sediment sampled in the Pass of Drumochter, samples from the Western Red Hills indicate that percentage clay plus silt in the debris flow samples is less than that for unreworked samples at $p < 0.005$, and values of percentage clay are also less for reworked samples than for *in situ* drift at $p < 0.0001$. Similarly,

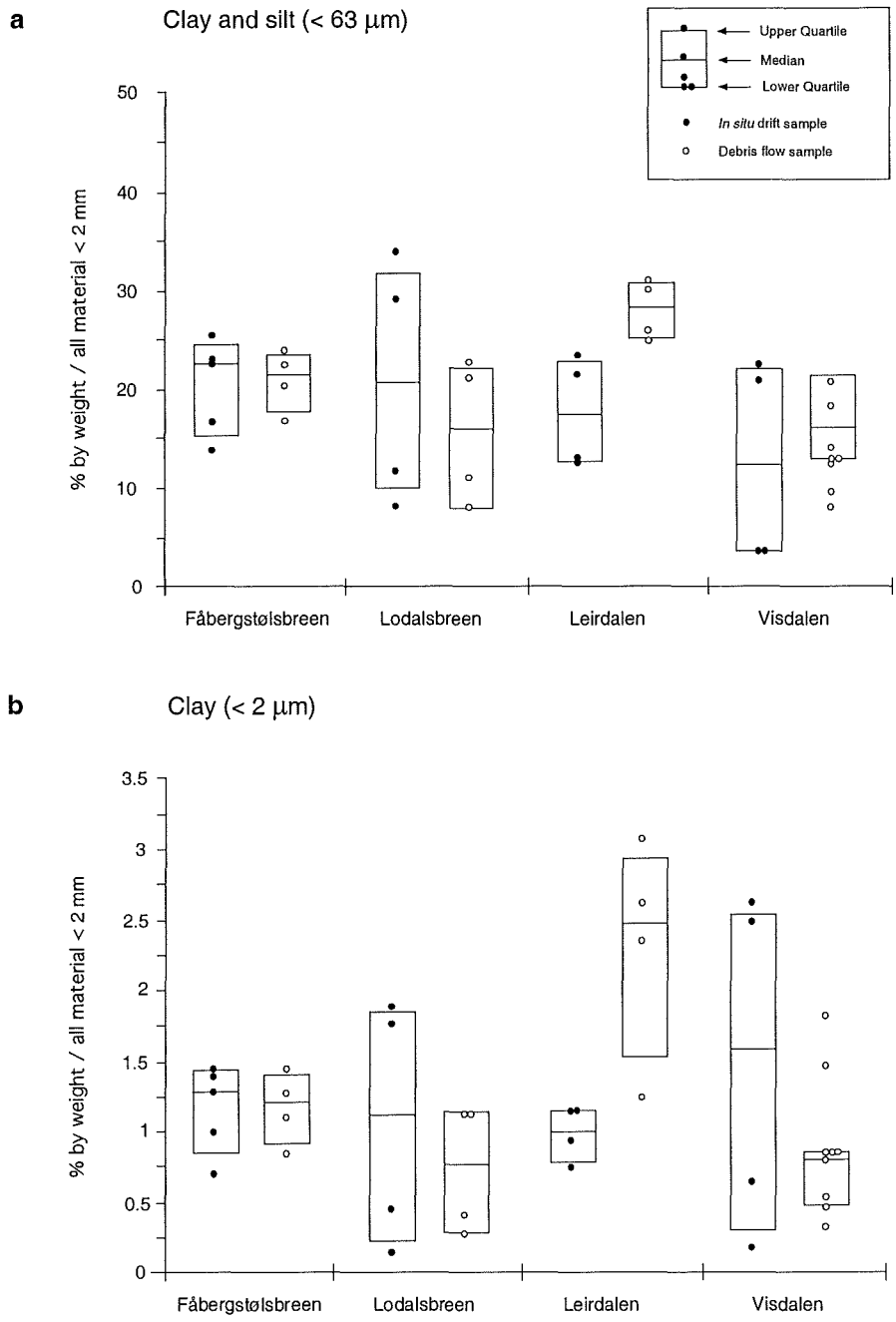


Figure 6.5. Dispersion diagrams illustrating (a) percentage clay plus silt (< 63 µm) and (b) percentage clay (< 2 µm) particles within fine-fraction samples (< 2 mm) from *in situ* drift and debris flow-reworked drift at Fåbergstølsbreen, Lodalsbreen, Leirdalen and Visdalen.

percentage clay plus silt content in reworked samples from Glen Docherty is less than that for unreworked drift at $p < 0.1$, and percentage clay in *in situ* drift

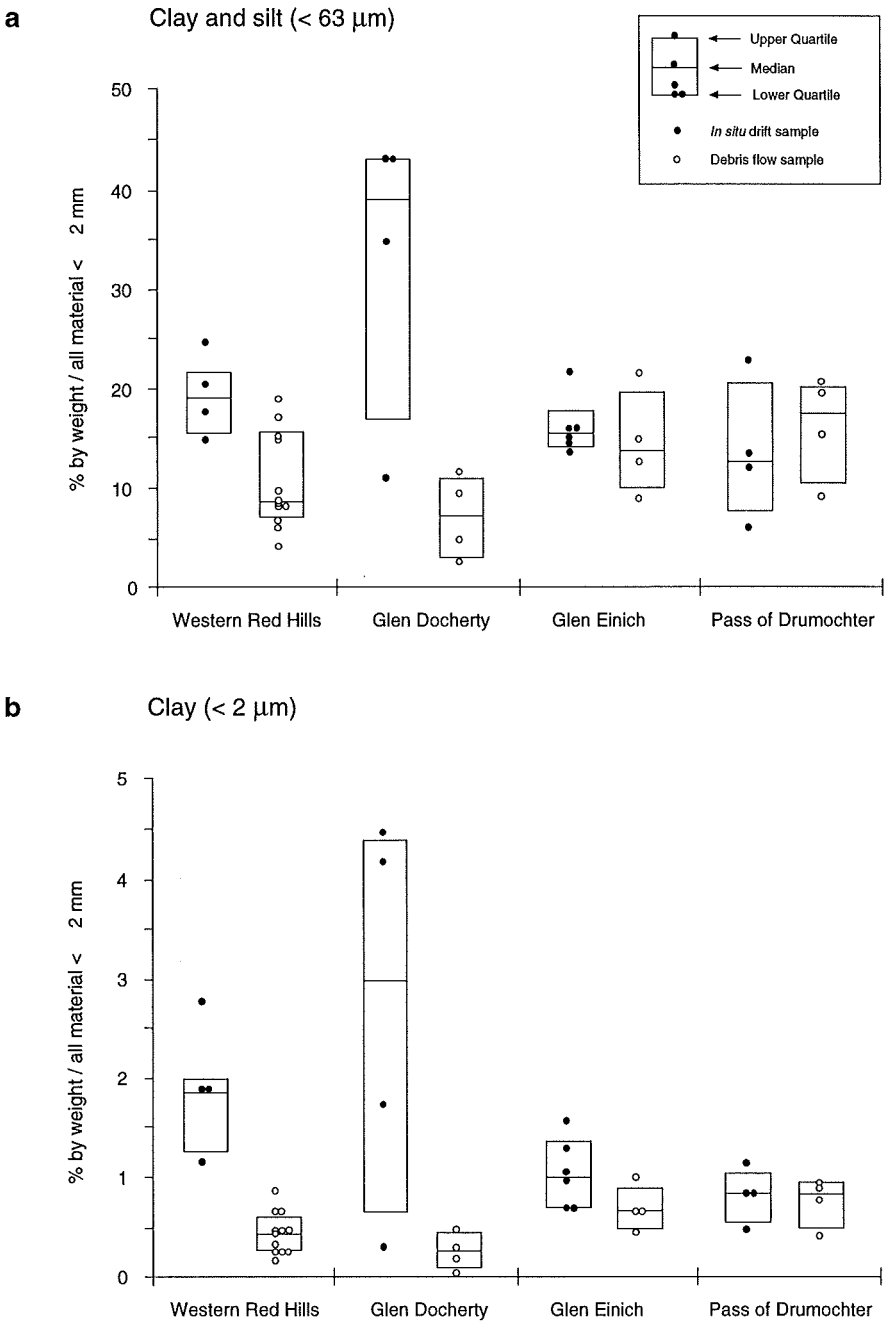


Figure 6.6. Dispersion diagrams illustrating (a) percentage clay plus silt (< 63 μm) and (b) percentage clay (< 2 μm) particles within fine-fraction samples (< 2 mm) from *in situ* drift and debris flow-reworked drift in the Western Red Hills, Glen Docherty, Glen Einich and the Pass of Drumochter.

deposits from Glen Einich exceeds that for debris flow deposits at $p < 0.1$. Taken together, however, the combined results suggest that eluviation of fines from

debris-flow deposits is only locally evident, and that comparative grain-size analyses therefore offer an uncertain means of differentiating *in situ* glacigenic sediments from those reworked by debris flows. The between-site variations evident in Figures 6.5 and 6.6 suggest that fine-fraction composition of both *in situ* and reworked glacigenic sediments may be determined more by lithology, mineralogy and granulometry of the parent materials, initial water content, deformation strain history and the mode of sediment deposition (*cf.* Lawson, 1979b; Haldorsen, 1981; Hooke and Iversen, 1995) than post-depositional history.

6.3.4 Sediment packing (void ratio)

Samples of intact sediment were removed from 68 *in situ* and debris flow-reworked glacigenic deposits at Fåbergstølsbreen, Lodalsbreen, Leirdalen, Visdalen, the Western Red Hills, Glen Docherty and the Pass of Drumochter, and void ratios calculated as a measure of sediment compaction or packing. It was hypothesised that debris flow facies may be more loosely packed (higher void ratio) than *in situ* basal till sediments. This suggestion was based on the observation that lodgement till deposits are commonly overconsolidated due to dewatering under ice overburden pressure, and consequently often have high bulk density and low void ratio, as do subglacial melt-out tills and comminution tills (Elson, 1989; Owen, 1991; Benn and Evans, 1998). In contrast, rapid downslope flowage of sediment tends to disaggregate and mix water and/or air with the constituent material, such that recent debris flow facies might be expected to be more loosely packed (higher void ratio) than *in situ* basal till sediments. The results (Figure 6.7), however, show that matrix compaction is an uncertain discriminant of *in situ* and reworked drift deposits. At three sites (Visdalen, Western Red Hills and Glen Docherty), void ratio is, as predicted, significantly lower in sediments reworked by debris flows than in samples of *in situ* till. However, in the samples from Lodalsbreen and the Pass of Drumochter no

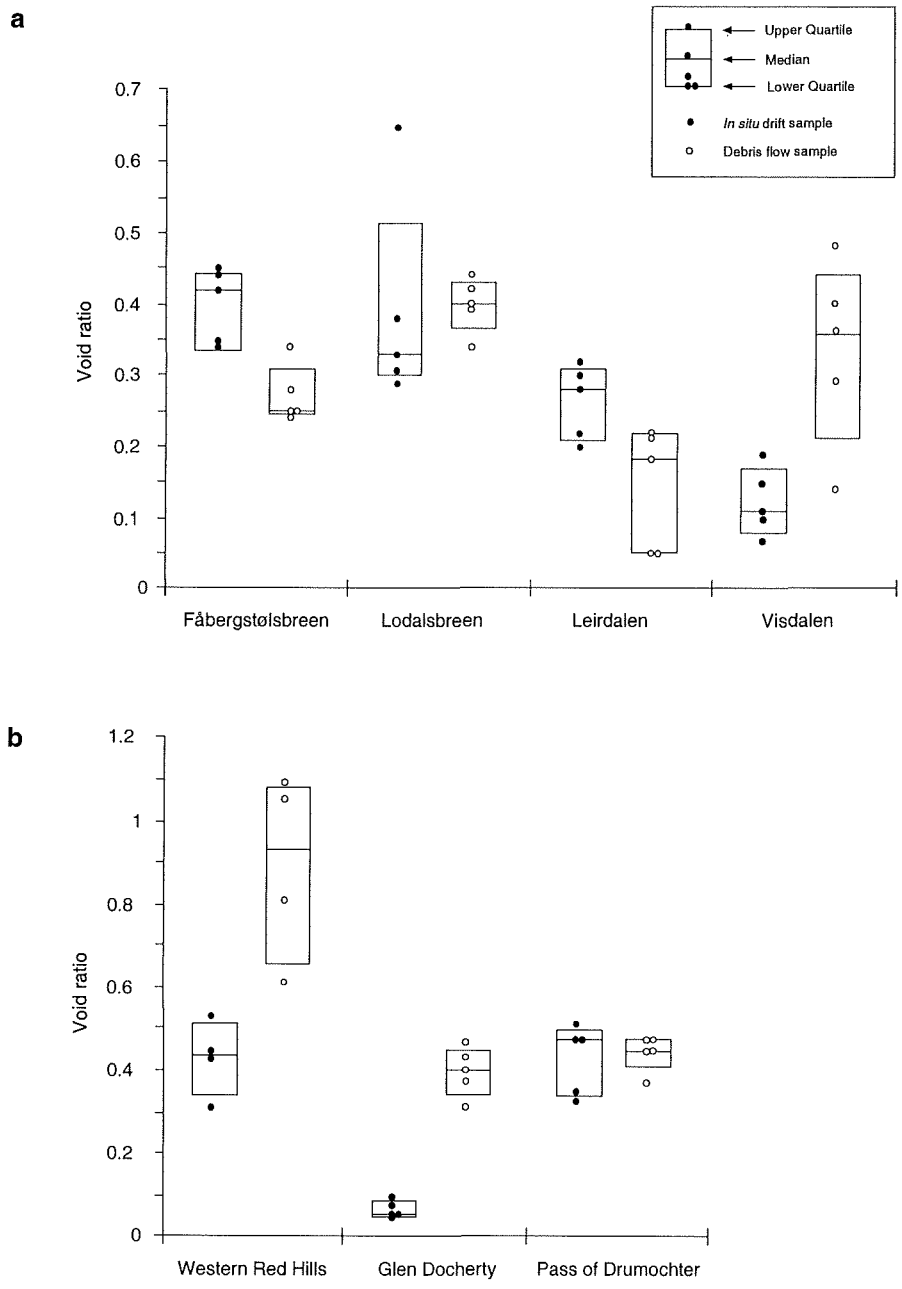


Figure 6.7. Dispersion diagram outlining void ratios of matrix material removed from *in situ* drift and debris flow-reworked drift deposits (a) at Fåbergstølsbreen, Lodalsbreen, Leirdalen and Visdalen, and (b) in the Western Red Hills, Glen Docherty and the Pass of Drumochter.

significant difference was detected in void ratio between the two sets of samples, and at Fåbergstølsbreen and Leirdalen void ratio is actually significantly higher in

the unreworked drift. These results indicate that only locally has reworking by debris flows produced a systematic decrease in sediment packing. The high void ratio in unreworked drift deposits at Fåbergstølsbreen and Leirdalen may be explained in part by deformation of tills by ductile flow (shearing) in a dilatant state. High void ratios in *in situ* drift may also reflect the melt-out of debris-rich ice, resulting in a loosely-packed sediment structure (Ronnert and Mickelson, 1992).

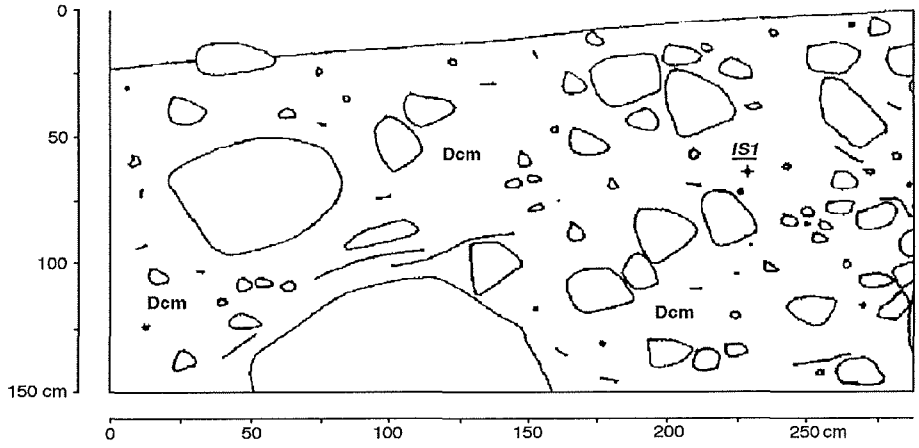
6.3.5 Structure and lithofacies

In situ drift facies

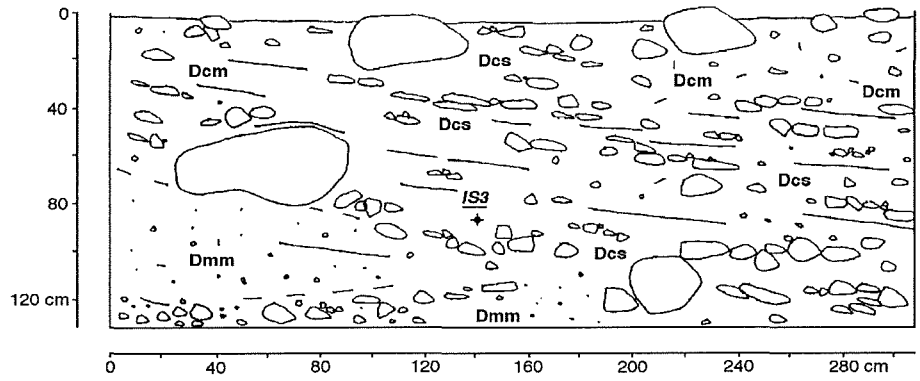
In Norway, clear exposures of *in situ* glacigenic sediment were present only in shallow sections eroded through basal till at Fåbergstølsbreen and Lodalsbreen, but sections through unreworked till were more extensive in the Scottish Highlands where valley-floor drift has been exposed by fluvial incision. A key to all lithofacies logs is provided in Figure 5.13.

The exposure of *in situ* drift at Fåbergstølsbreen displayed a very poorly-sorted massive clast-supported diamict structure (F1, Figure 6.8) and a notable lack of stratification. There is some evidence of folding around boulders, but slope-parallel bedding is absent. The presence of striations on the upvalley sides of large clasts embedded in the surface of the till bench indicates glacial over-riding following lodgement and lack of postdepositional reworking (*cf.* Boulton, 1978; Sharp, 1982; Benn, 1994b). An *in situ* drift section exposed in the Lodalsbre foreland (L1, Figure 6.8) also comprises an unsorted clast-supported diamicton, but unlike F1 is traversed by numerous sub-horizontal joints and folds, giving the diamict a fissile structure. These joints are locally striated, suggesting they are former shear planes (Boulton *et al.*, 1974), and are characteristic of

F1



L1



W1

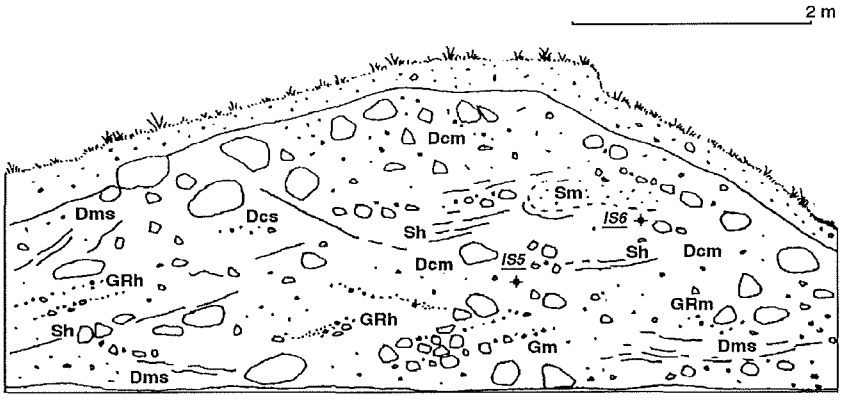
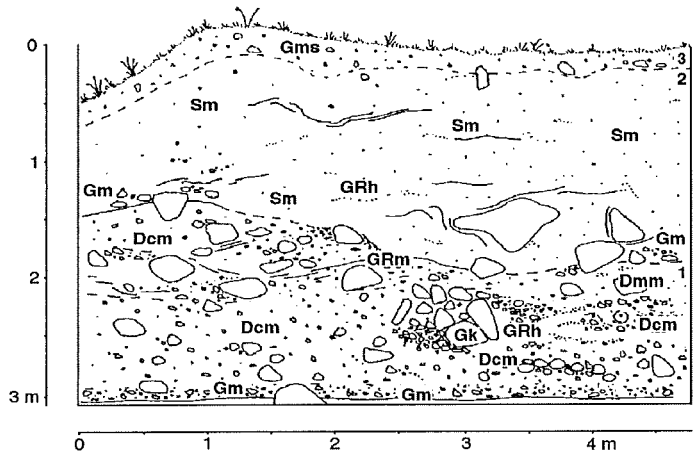


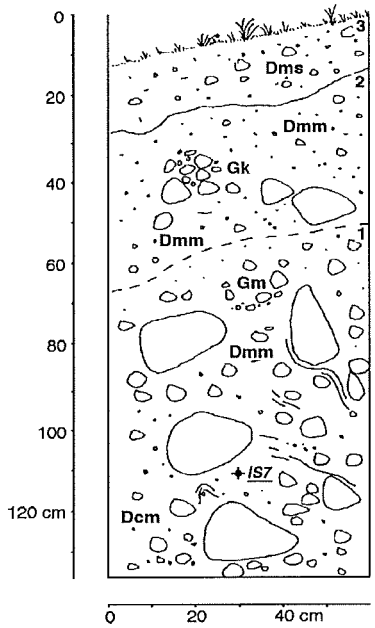
Figure 6.8. Sections F1, L1 and W1 showing structure and lithofacies exposed in valley-floor *in situ* drift at Fåbergstølsbreen (F1), Lodalsbreen (L1) and in the Western Red Hills (W1), and location of clast fabric samples. No vertical exaggeration. Key in Figure 5.13.

lodgement till deposited by plastering of glacial debris from a sliding glacier sole, possibly by pressure melting (Dreimanis, 1989). Section W1 (Figure 6.8) is representative of unreworked valley-floor drift in the Western Red Hills, and highlights the variability of the internal characteristics of *in situ* drift, particularly where the formation of a drift assemblage involved multiple cycles of sediment redeposition (Benn and Evans, 1998). The lithofacies revealed consist of clast-supported diamictons, lenses of fissile matrix-supported diamictons, massive and bedded gravels, and bedded sands. Whilst the lateral facies parallel the hummock surface, the lower central area is largely structureless and in turn overlain by folded sands and a massive faulted block of diamicton. These interbedded structures are interpreted as recording a complex history of syndepositional deformation and reworking involving the melt of buried ice (Benn, 1990). Section D1 (Figure 6.9) is representative of many sections exposed in *in situ* hummocky drift which lies on the floor of lower Glen Docherty. Two sediment associations are revealed, the lower (unit 1) characterised by massive, clast-supported diamictons, gravels and granules, overlain by massive sands (unit 2). The unit 1 diamictons contain predominantly sub-rounded and striated clasts, and folding has imparted a fissile appearance to the upper left part of the unit, highly suggestive of shearing of *in situ* drift during lodgement. A zone of folded interbedded granules at the right-hand margin of this section may indicate deformation during melt-out deposition. An exposure of *in situ* drift in a ridge located on the floor of Glen Einich (E1, Figure 6.9) is dominated by very poorly-sorted matrix-supported diamictons which exhibit a notable lack of stratification. Although there is some evidence of folding around boulders in unit 1, indistinct downslope parallel bedding is only present near the surface (unit 3). Two sections exposed in the floor of the Pass of Drumochter (PD1 and PD2) reveal the internal structure of *in situ* drift and are illustrated in Figures 6.9 and 6.10. As with sections F1, L1, and E1, these exposures reveal predominantly structureless diamictons, with evidence of only local reworking, probably as a result of

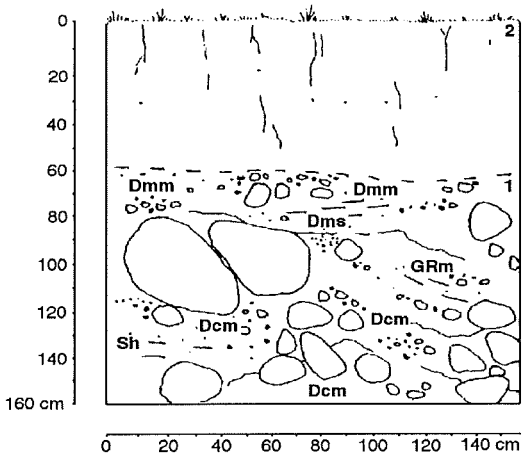
D1



E1



PD1



PD2

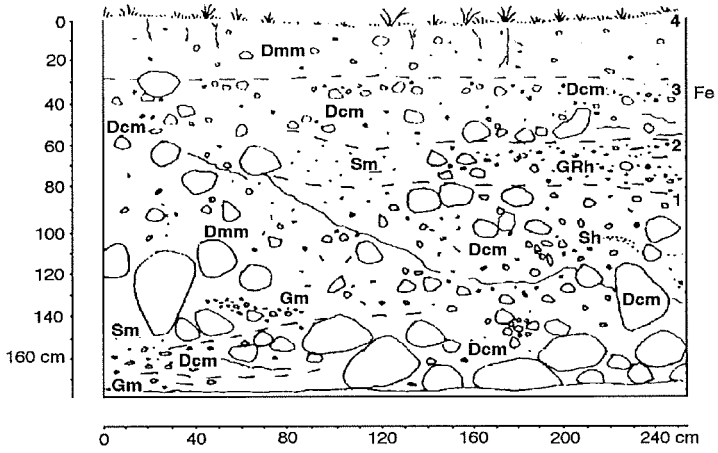


Figure 6.9. Sections D1, E1, PD1 and PD2, showing structure and lithofacies exposed in *in situ* drift in Glen Docherty (D1), Glen Einich (E1) and the Pass of Drumochter (PD1 and PD2). Location of clast fabric samples is also shown. Key in Figure 5.13.

Figure 6.10. Section PD2 exposed in the floor of the Pass of Drumochter, revealing the predominantly structureless internal architecture of an *in situ* drift deposit.

pressure melting. Evidence for this includes occasional sand-rich layers (e.g. PD2, unit 2, Figure 6.9) and zones of local shearing and folding, (e.g. PD1, unit 1 and PD2, unit 1, Figure 6.9).

Debris flow facies

Lithofacies logs illustrating the sedimentological characteristics of drift deposits reworked by debris flows are illustrated in Figures 6.11 to 6.15. These sections were exposed in slope-foot debris cones or levée deposits beside debris flow tracks, and generally reveal drift accumulations composed of stacked sediment units aligned sub-parallel to the slope surface (Figures 6.11 to 6.18). Individual units vary in thickness from a few centimetres to two metres, though units more than one metre thick are rare. In many of the Scottish examples these units are separated by thin organic-rich horizons which parallel the slope surface

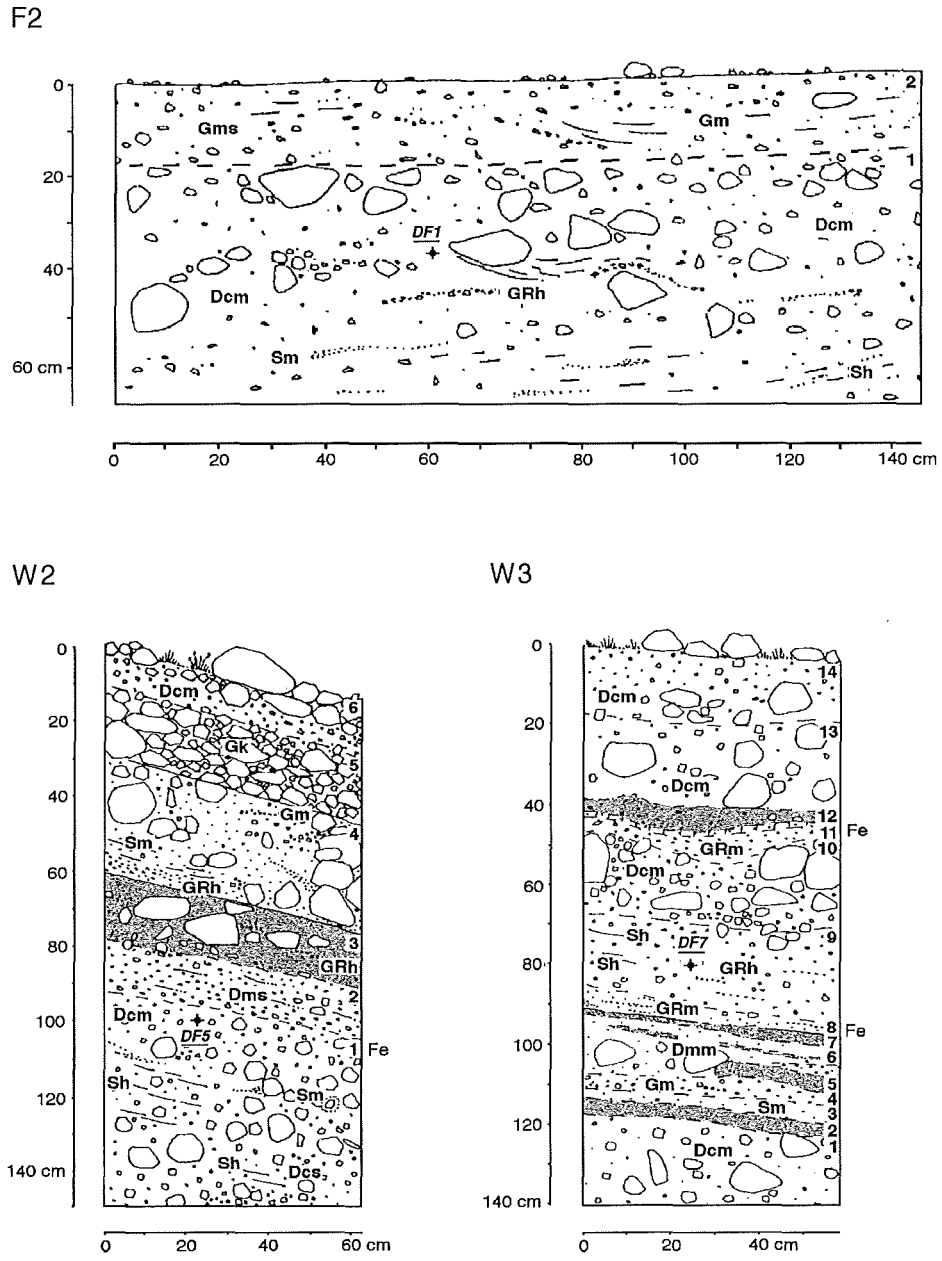


Figure 6.11. Sections F2, W2 and W3, showing structure and lithofacies exposed in a debris cone at Fåbergstølsbreen (F2) and in levée deposits in the Western Red Hills (W2 and W3). Location of clast fabric samples is also shown. Key in Figure 5.13.

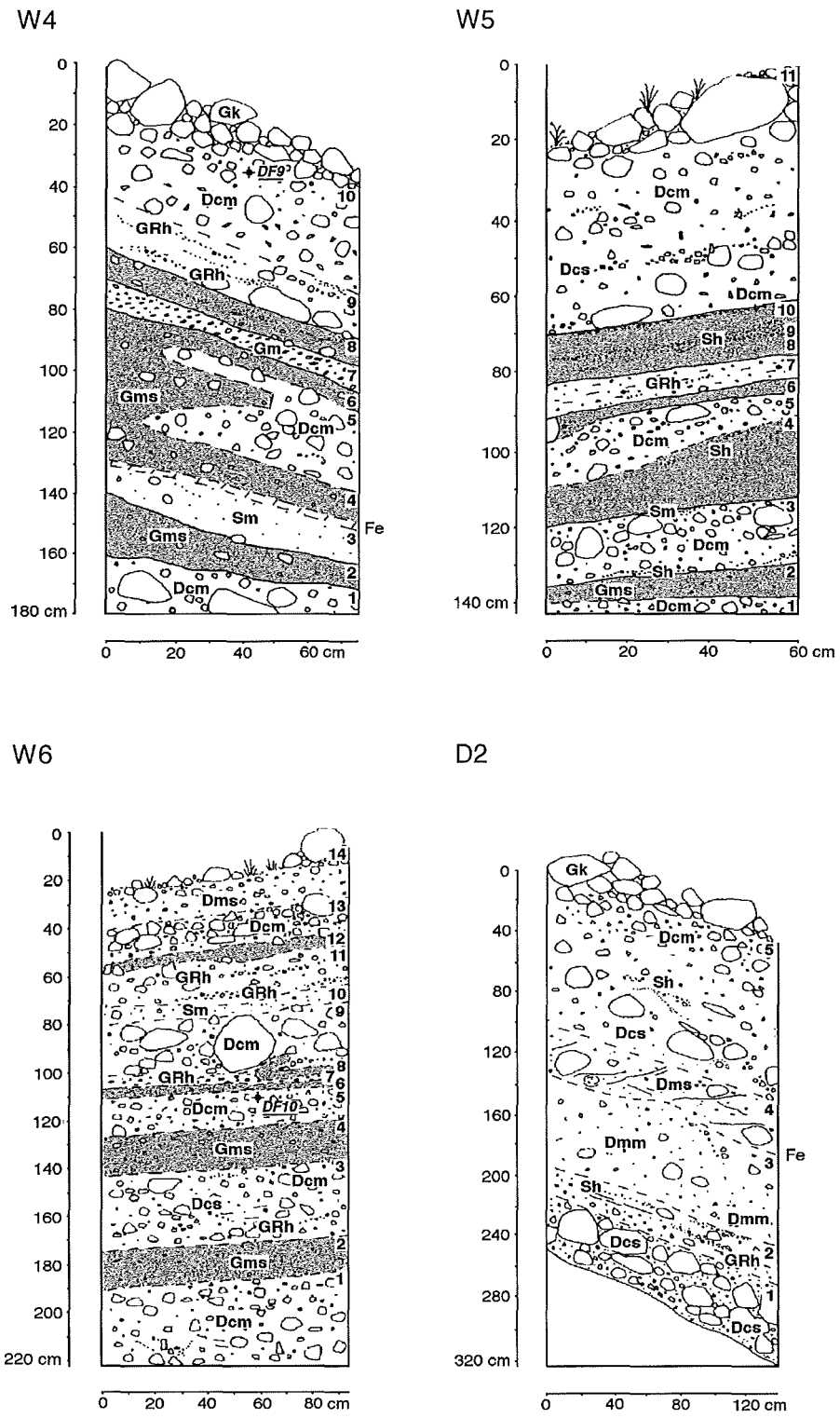


Figure 6.12. Sections W4-W6 and D2, showing structure and lithofacies exposed in a debris cone in the Western Red Hills (W5) and in levée deposits in the Western Red Hills (W4 and W6) and Glen Docherty (D2). Location of clast fabric samples is also shown. Key in Figure 5.13.

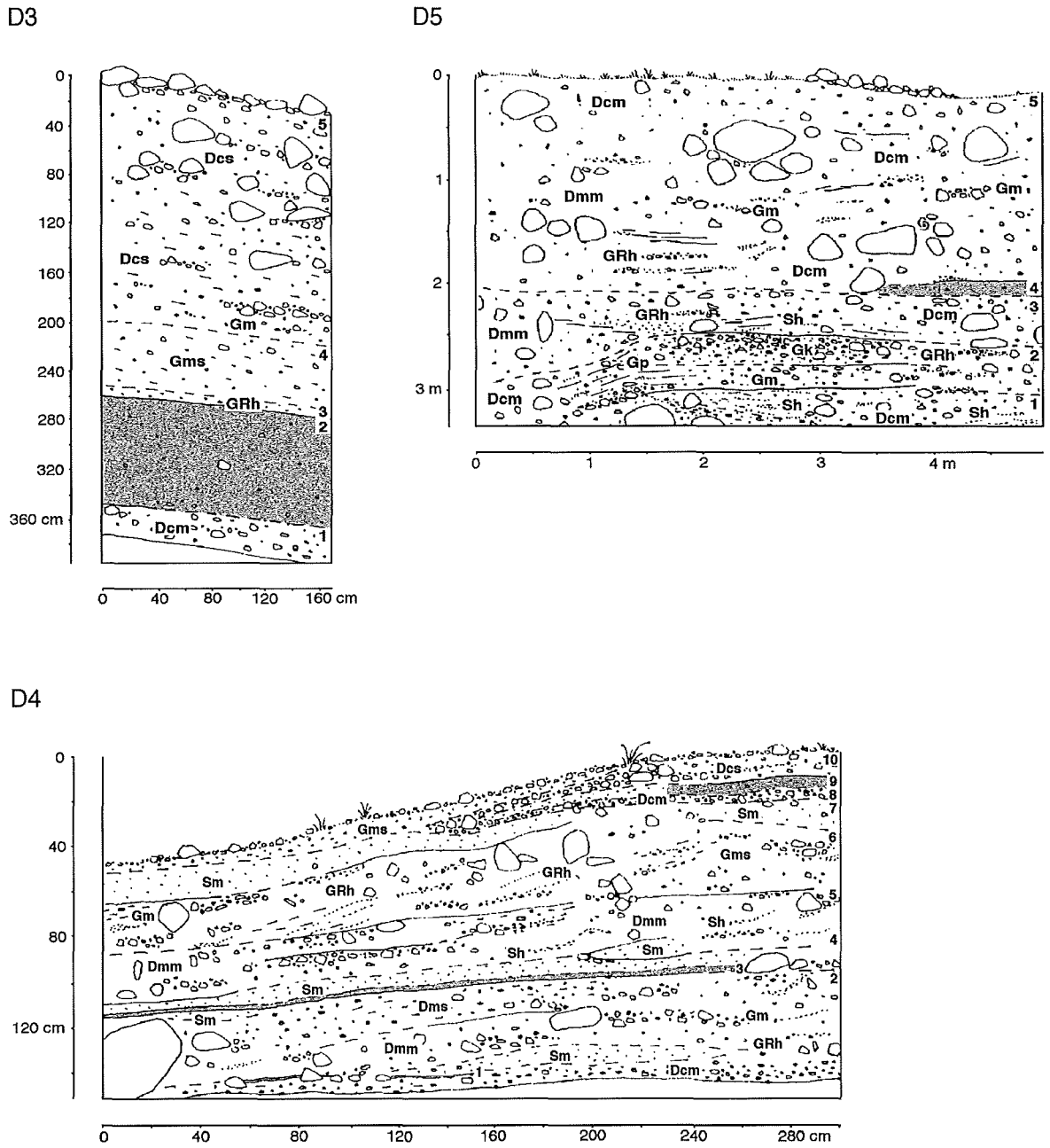


Figure 6.13. Sections D3-D5, showing structure and lithofacies exposed in a debris cone in Glen Docherty. Key in Figure 5.13.

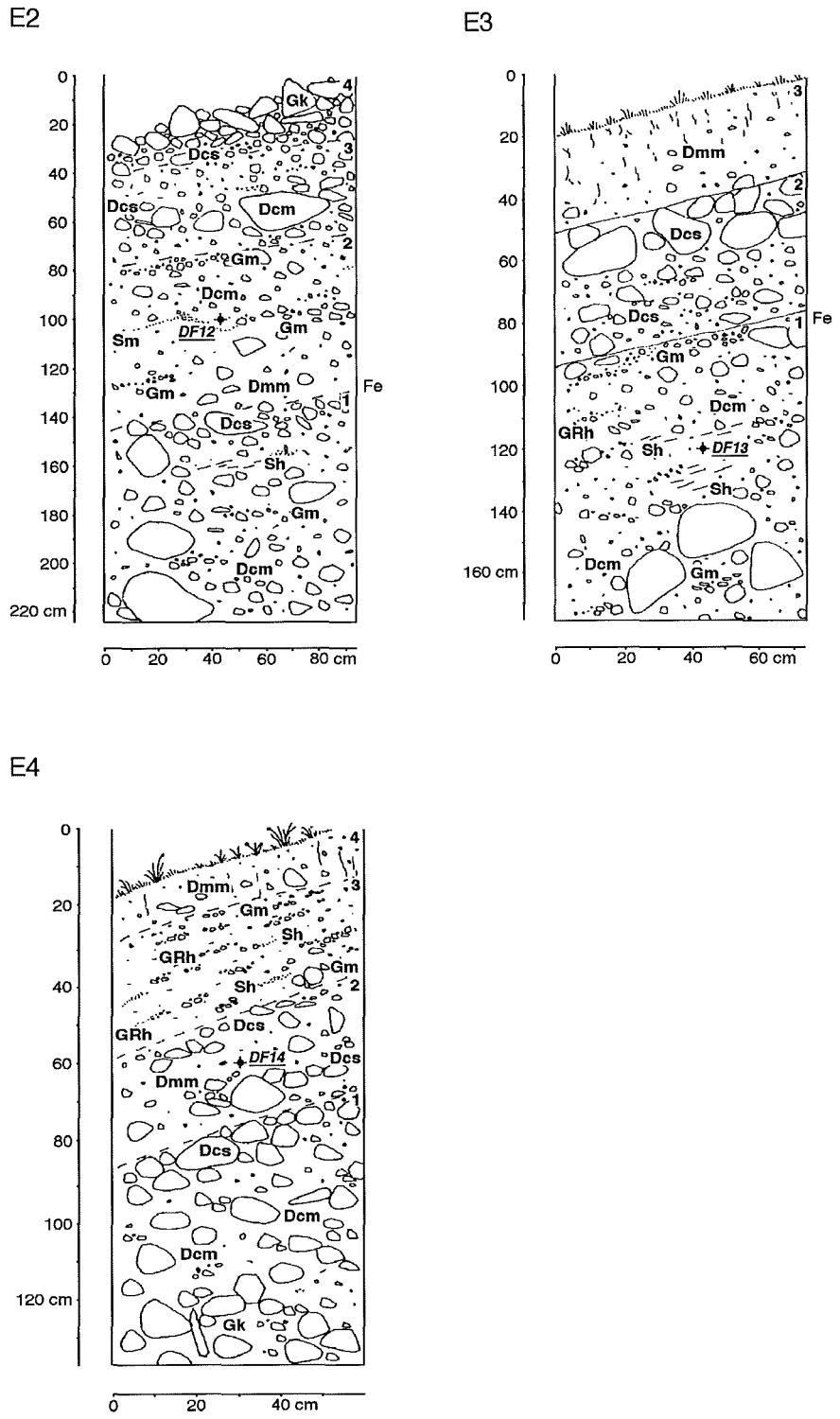
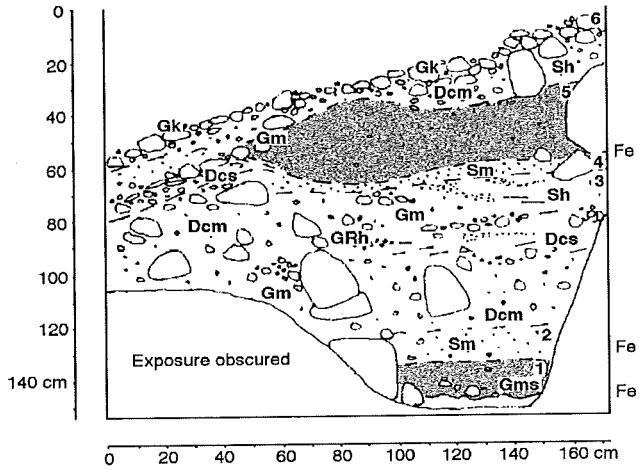
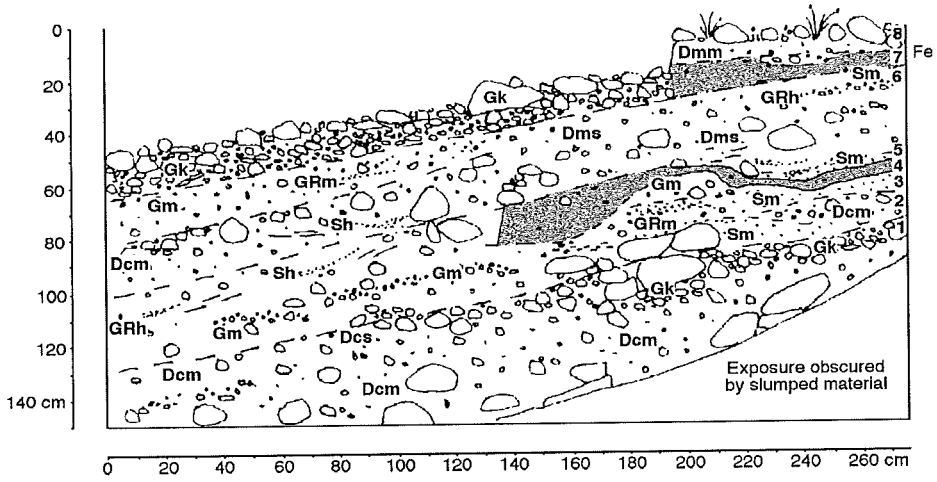


Figure 6.14. Sections E2-E4, showing structure and lithofacies exposed in levée deposits in Glen Einich, and location of clast fabric samples. Key in Figure 5.13.

PD3



PD4



PD5

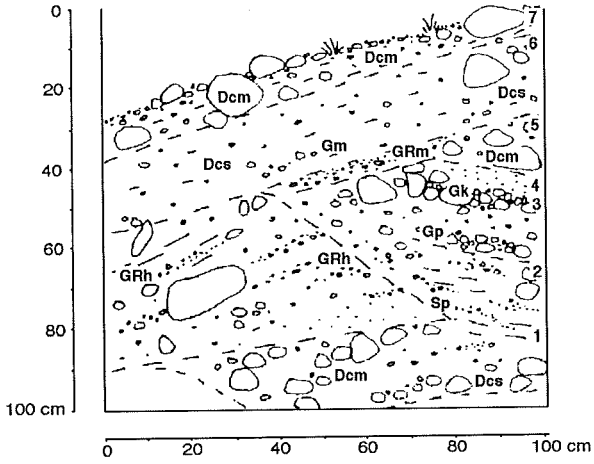


Figure 6.15. Sections PD3-PD5, showing structure and lithofacies exposed in a debris cone in the Pass of Drumochter. Key in Figure 5.13.

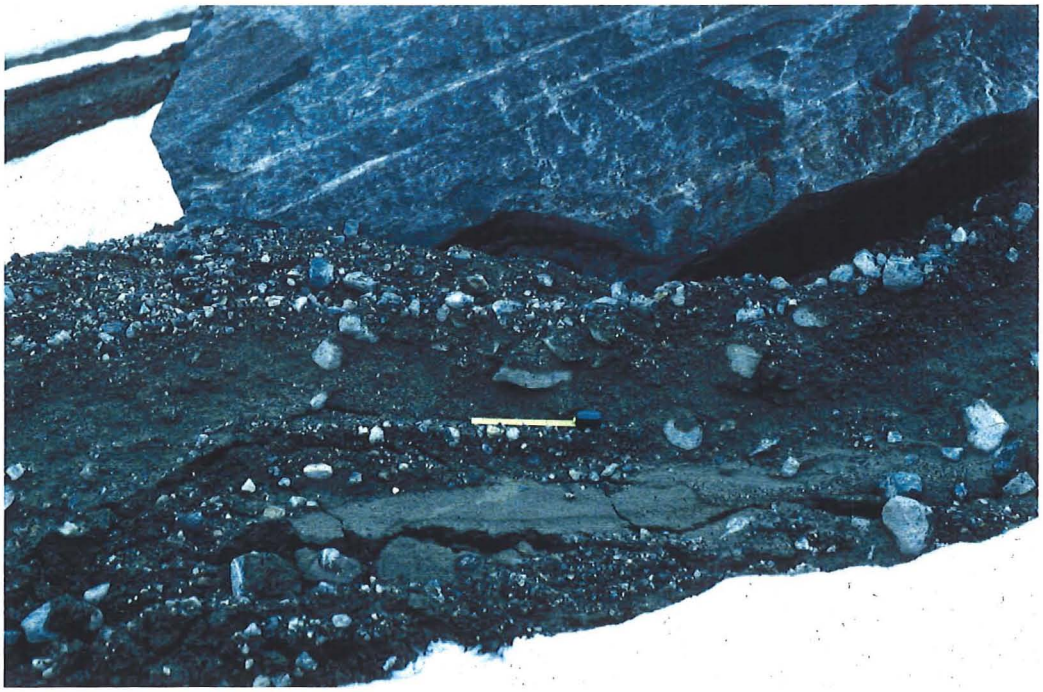


Figure 6.16. Recent paraglacial debris flow deposit at the snout of Fåbergstølsbreen. Notice the crude stratification, and the clast-rich horizon that separates distinctive units at this site.

Figure 6.17. Slope-parallel stratification of coarse- and fine-grained reworked deposits in section D4, Glen Docherty. The coarse-clastic sediment units represent resedimentation of drift by debris flow activity, and the sand- and gravel-rich facies are regarded as slopewash deposits.

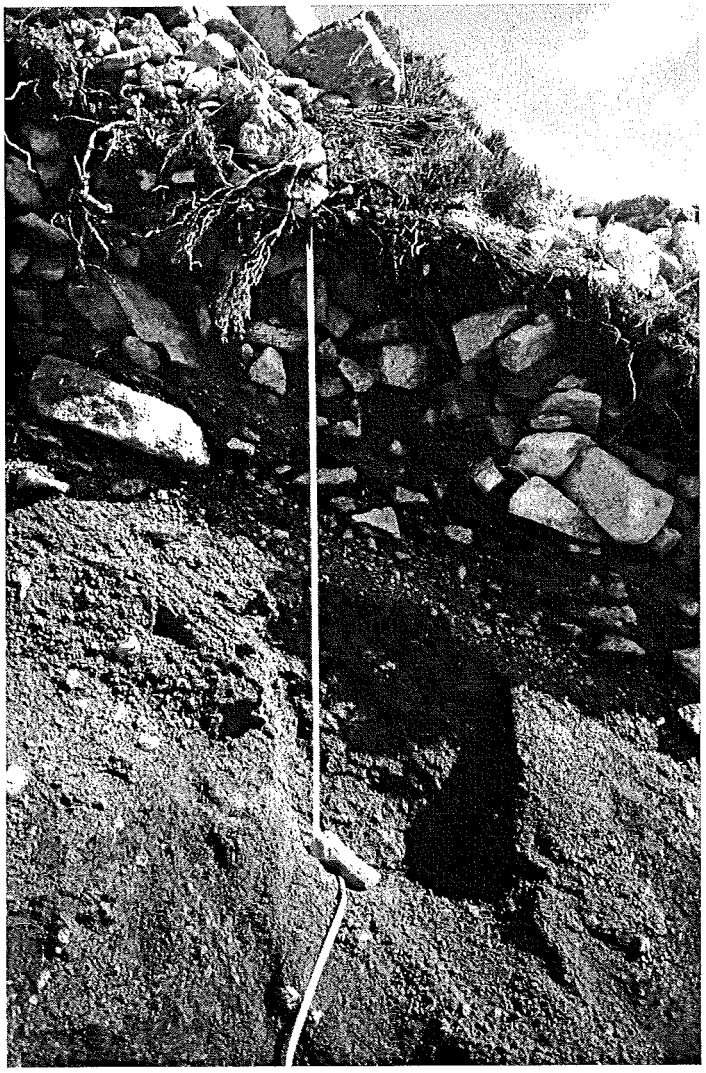


Figure 6.18. Section W2, exposed in levée deposits in the Western Red Hills, revealing a stacked stratigraphy of reworked drift deposits. Notice the alignment of sediment units crudely-parallel to the slope surface, consisting of openwork and matrix-supported diamictons, gravels and sands.

and are interpreted as buried soils that developed on former ground surfaces during periods of slope stability. Contacts between individual units often appear to be conformable over short distances, but truncation of both inorganic sediment units and organic-rich layers indicate erosion of the upper surfaces of some beds. Structural elements not present in the *in situ* drift facies were commonly identified in both coarse-clastic and fine-grained facies in the logged debris flow sections, and these are described in more detail below.

Coarse-clastic units occur at various depths below the surface and may be openwork (e.g. W2, unit 5, Figure 6.11), clast-supported diamictons (e.g. W3, unit 10, Figure 6.11; W6 unit 9, Figure 6.12) or matrix-supported diamictons (e.g. D2, unit 3, Figure 6.12). These units exhibit a wide range in matrix texture, clast concentration and clast size. Many coarse-clastic units exhibit crude stratification aligned approximately parallel to the surface (e.g. W2, unit 2, Figures 6.11 and 6.18; W5, unit 11, Figure 6.12; unit 6, PD5, Figure 6.15), and many beds, for example unit 10 in W4 (Figure 6.12), unit 2 in E3 (Figure 6.14) and unit 6 in PD3 (Figure 6.15), exhibit an increase in the size and concentration of clasts towards the upper contact. Within individual diamicton units, upward coarsening (inverse grading) is occasionally accompanied by a rise in the concentration of clasts downslope, and some diamictons grade into tightly-packed terminal 'lobes' over relatively short distances (e.g. PD4, unit 1, Figure 6.15). Stratified matrix-supported diamictons possibly reflect internal shear during flow or superimposition of multiple sediment pulses, and are thus indicative of mass transport (Lindsay, 1968; Pierson and Costa, 1987; Wells and Harvey, 1987). The absence of miniature shear structures in some matrix-supported diamict facies may reflect poor preservation of such features due to variations in matrix characteristics, or lower levels of internal deformation within some diamictons. Inverse grading and bouldery terminal lobes are also indicative of sediment reworking by debris flow (e.g. Ryder, 1971a; Takahashi, 1981; Schultz, 1984; McArthur, 1987; Nieuwenhuijzen and van Steijn, 1990; Bertran and Texier, 1994), and have been explained in terms of dispersive forces generated through particle collisions within a flowing mass (Bagnold, 1954; Takahashi 1981, 1991). Although variable in character, these coarse-clastic facies have strong affinities with those in debris flow facies reported for other environments (e.g. Suwa and Okuda, 1980; Rapp and Nyberg, 1981; Wells and Harvey, 1987; Eyles *et al.*, 1988; Eyles and Kocsis, 1988; Nieuwenhuijzen and van Steijn, 1990; Derbyshire and Owen, 1990; Owen, 1991; Ballantyne and Benn, 1994; van Steijn *et al.*, 1995;

Coussot and Meunier, 1996; Salt and Ballantyne, 1997) and contrast sharply with largely unstratified, in *situ* drift facies.

In addition to coarse-clastic facies, almost all debris flow sections contain discrete fine-grained interbeds of silt, sand and fine gravel in which large clasts are generally absent (e.g. W3, unit 9, Figure 6.11; W5, unit 7, Figure 6.12; E4, unit 3, Figure 6.14). Although some gravel-rich units, for example unit 4 in D3 (Figure 6.13), attain thicknesses in excess of 50 cm, most fine-grained interbeds are thinner than the diamict units. They are frequently several metres long but tend to pinch out both upslope and downslope to form elongated lenticular horizons. Their composition is variable, from silty fines to coarse gravels (Figure 6.19). Many beds, for example those within unit 1 in F2 (Figure 6.11), unit 11 in W6 (Figure 6.12) and unit 1 in section D5 (Figure 6.13), exhibit a crude stratification parallel to the surface, whilst others, including unit 4 in D4 (Figure 6.13) and unit 5 in PD5 (Figure 6.15) are massive. These sand- and fine gravel-rich facies are regarded as slopewash deposits (*cf.* Wells and Harvey, 1987; Carling, 1987; Eyles and Kocsis, 1988; Eyles *et al.*, 1988; Lawson, 1988; Brazier and Ballantyne, 1989; Derbyshire and Owen, 1990) and may represent eluviation of fines from recently immobilised debris flows upslope (Takahashi, 1991; Hinchliffe *et al.,* 1998), or reworking of the surfaces of debris flow deposits during rainstorms (Figure 6.20). The presence of bedding within such units has previously been considered representative of fluctuations in discharge or variations in sediment concentration (Wells and Harvey, 1987; Carling, 1987), though exceptionally it may also represent several discrete slopewash events. In particular, the frequent intercalation of sand and fine gravel units with diamict and organic-rich layers in D4 (Figure 6.13) suggests successive episodes of slopewash activity over a prolonged time period.

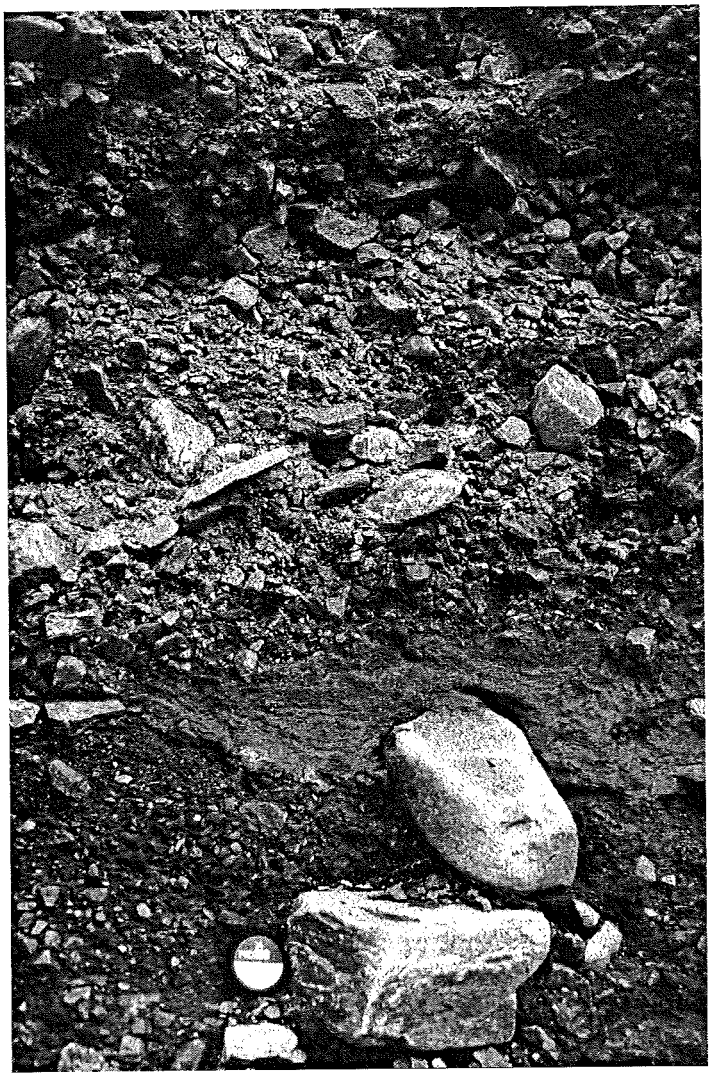


Figure 6.19. Part of section D5 exposed in a debris cone in Glen Docherty. Notice the intercalation of fine-grained sediment units comprising silt, sand and gravel with coarse-clastic units.

6.3.6 Comparative sedimentological characteristics of *in situ* and reworked drift: summary

The above analyses confirm previous suggestions (e.g. Lawson, 1988; Ballantyne and Benn, 1994) that preferred clast orientation and structural and lithofacies relationships are the most valuable criteria for distinguishing *in situ* drift deposits from those reworked by debris flow activity. Conversely, clast angularity, shape and texture, matrix granulometry and void ratio appear to be

Figure 6.20. Deposition of fine-grained sediment over recent debris flow deposits in Leirdalen, representing reworking of the surfaces of debris flow deposits during rainstorms.

poor or equivocal discriminants of *in situ* and glacigenic sediments reworked by debris flows. In particular, the preferred orientation of clasts in all of the *in situ* drift samples tends to be aligned parallel or sub-parallel to the valley axis (*cf.* Krüger, 1994; Hart, 1994; Benn and Evans, 1998), whereas that of all the debris flow-reworked samples exhibits a preferred downflow alignment, as identified by others (e.g. Ballantyne and Benn, 1994). As observed by Eyles and Kocsis (1988), structural characteristics such as crude slope-parallel stratification and lenses and interbeds of better-sorted sediment offer a means of distinguishing

glacigenic deposits reworked primarily by debris flows from most unreworked, *in situ* drift deposits. Diagnostic structural elements most frequently exposed in section in debris flow deposits include stacked sediment units whose alignment and internal stratification crudely parallels that of the slope surface (*cf.* Lawson, 1988; Nieuwenhuijzen and van Steijn, 1990), elongated lenticular fine-grained units, and bedded sand- and fine gravel-rich facies representing mobilisation of sediment by slopewash (*cf.* Brazier and Ballantyne, 1989; Hinchliffe *et al.*, 1998). Less common structural unconformities identified in reworked sediments that are diagnostic of subaerial remobilisation include inverse grading, bouldery terminal lobes (*cf.* Takahashi, 1981; Bertran and Texier, 1994), openwork clastic units and miniature shear structures within matrix-supported diamict facies, as observed in resedimented soliflucted tills by Wells and Harvey (1987).

6.4 Interpreting hillslope glacigenic drift deposits.

The use of structural characteristics and preferred clast orientation as discriminants of *in situ* and reworked glacigenic drift was evaluated in three contrasting topographic situations: in steep valley-side drifts flanking two of the Jostedalsbre foreland areas; in valley-side slope locations at the two sites in Jotunheimen; and finally at slope-foot locations in the Western Red Hills on Skye.

6.4.1 Case study 1: Valley-side drifts around Jostedalsbreen

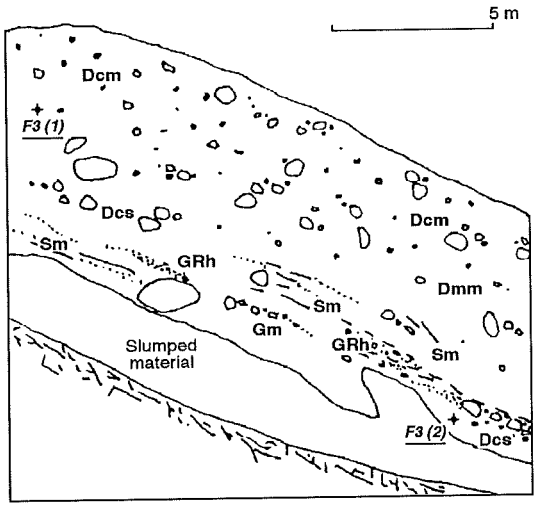
The characteristics of the steep valley-side drifts that have been exposed by recent glacier retreat at Fåbergstølsbreen and Lodalsbreen were investigated at exposures in the sides of deep gullies that feed the paraglacial debris-flow accumulations on the valley floors at these sites (Figure 6.21). All of the sections examined contain two sediment associations, and five representative sections are illustrated in Figures 6.22 and 6.23. The upper association consists of a massive

Figure 6.21. Fresh section through glacigenic drift in a gully sidewall above Fåbergstølsbreen, illustrating the characteristics of steep valley-side drifts exposed by recent glacier retreat.

diamicton 2-7 m thick. Clasts are up to 3 m long, predominantly angular or subangular, blocky in shape, and preferentially aligned parallel or subparallel to the valley axis (Figure 6.24). These characteristics strongly suggest that the upper sediment association at both sites represents an *in situ* glacigenic deposit, an interpretation confirmed at Fåbergstølsbreen where this deposit terminates upslope in the outermost 'Little Ice Age' lateral moraine.

In contrast, at all exposures the lower association exhibits crude stratification parallel to the slope. Clast- and matrix-supported diamictons 0.5-3.0 m thick predominate, reaching a cumulative thickness of 13 m at one site. Individual diamicton units are separated by thin bands of bedded sands, gravels and granules, and by occasional clast-rich beds such as those in section L3 (Figure 6.23). In sections L2, L3 and L4 the contact between the two associations exhibits marked truncation and is clearly erosional. Clast fabric measurements for

F3



F4

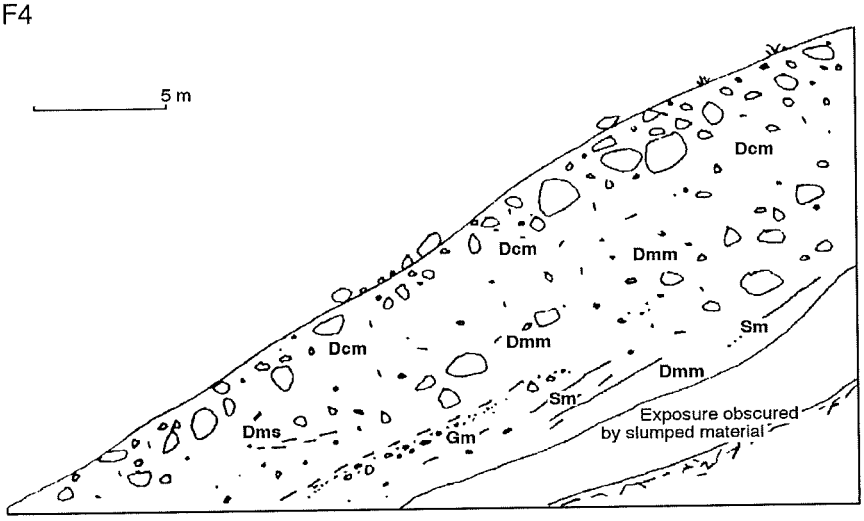


Figure 6.22. Sections F3 and F4 showing structure and lithofacies exposed in gully sidewalls at Fåbergstølsbreen and location of clast fabric samples. No vertical exaggeration. Key in Figure 5.13.

the lower association at sections F3 and L4 display a preferential downslope orientation (Figure 6.24). The lower association thus exhibits structural characteristics very similar to those of recent debris flow deposits at Fåbergstølsbreen and at sites in the Scottish Highlands (see section 6.3.5), and indeed to those of debris flow deposits derived from the remobilisation of glacial

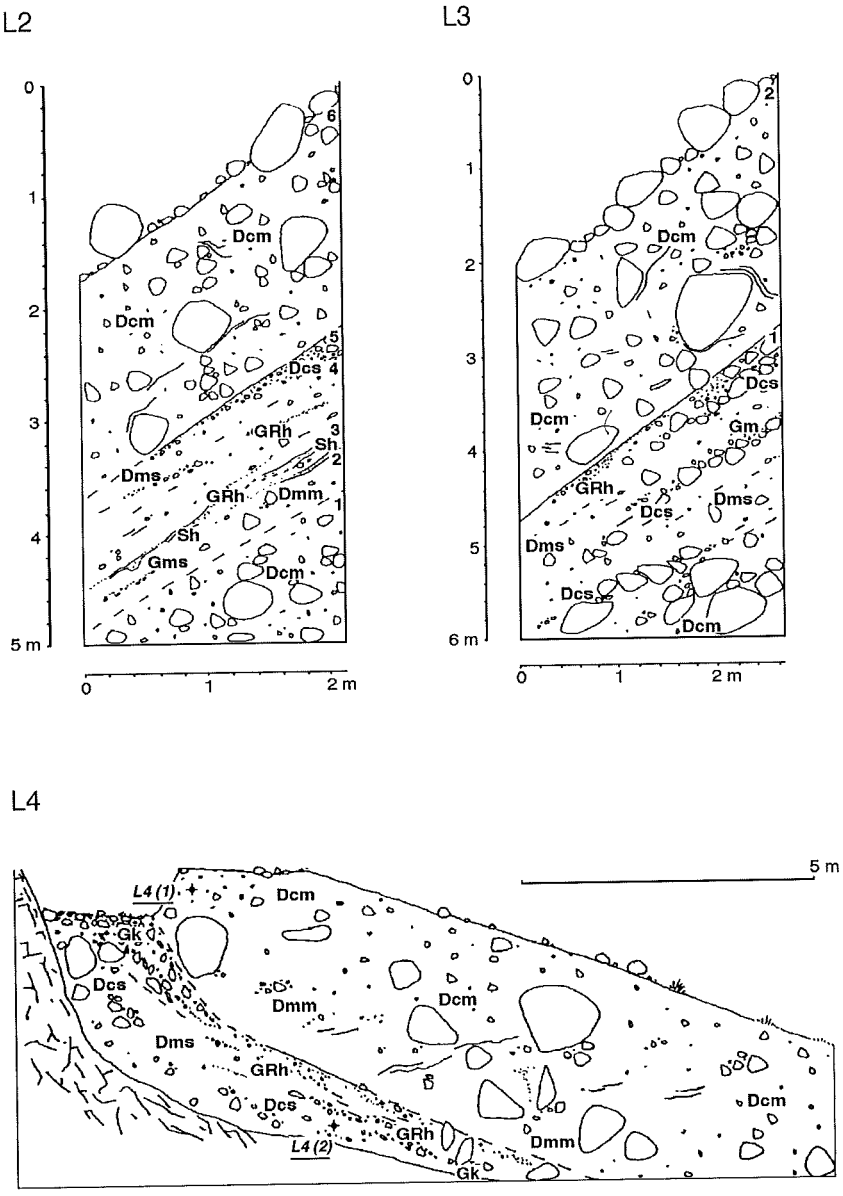


Figure 6.23. Sections L2-L4 showing structure and lithofacies exposed in gully sidewalls at Lodalsbreen and location of clast fabric samples. No vertical exaggeration. Key in Figure 5.13.

drift in other environments (e.g. Eyles *et al.*, 1988; Eyles and Kocsis, 1988; Brazier and Ballantyne, 1989; Owen, 1991). This association is therefore interpreted in terms of accumulated debris-flow facies representing ancient (probably Preboreal) remobilisation of glacigenic drift prior to the emplacement of the overlying ('Little Ice Age') unreworked glacigenic sediments (*cf.* Ballantyne

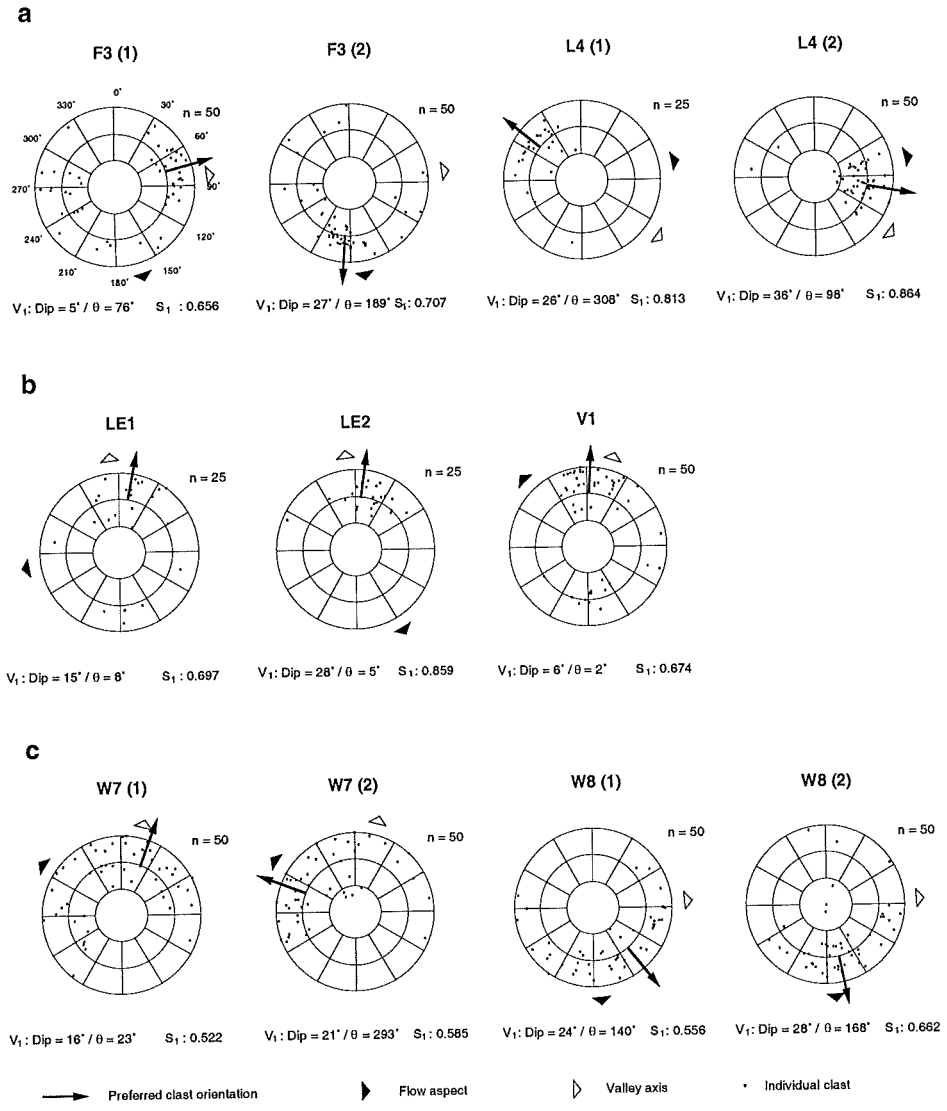


Figure 6.24. Stereographic fabric plots for samples of elongate clasts from valley-side glacigenic sediments exposed in section (a) at Fåbergstølsbreen (F) and Lodalsbreen (L) in the Jostedalsbre area, (b) Leirdalen (LE) and Visdalen (V) in Jotunheim, and (c) in the Western Red Hills (W) on the Isle of Skye. θ represents the preferred orientation of the long axes of clasts as identified by the principal eigenvector, V_1.

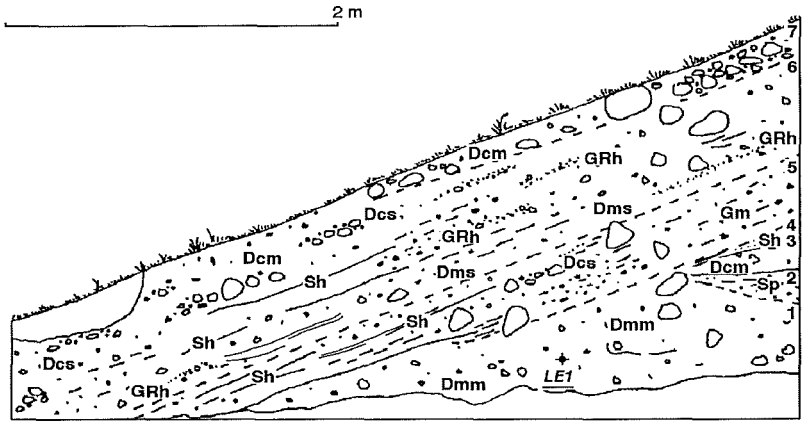
and Benn, 1994, 1996). The discontinuities between individual diamicton units are inferred to represent the boundaries between successive flows, and the sand and gravel beds are interpreted as the products of the reworking of the surfaces of individual flows by wash.

The two sediment associations represented in the valley-side drift deposits at Fåbergstølsbreen and Lodalsbreen are therefore inferred to have a common, two-stage history. The lower association apparently reflects widespread paraglacial reworking of valley-side glacigenic deposits following regional deglaciation at the end of the Preboreal chronozone. The upper unit represents predominantly unreworked glacigenic sediment deposited during recent ('Little Ice Age') advance and retreat of Fåbergstølsbreen and Lodalsbreen. The predominantly erosional contact between the two associations implies reworking of the Preboreal paraglacial deposits during the advance of these glaciers to their 'Little Ice Age' limits.

6.4.2 Case study 2: Valley-side drifts in Jotunheimen

Numerous gullies are cut through vegetated valley-side drift exposed by Preboreal ice retreat at Leirdalen and Visdalen. Three sidewall exposures in the upper parts of active gullies were excavated and logged and their sedimentary characteristics investigated using the diagnostic criteria established above. Two representative exposures in Leirdalen (LE1) and Visdalen (V1) are illustrated in Figure 6.25. Section LE1 comprises two sediment associations that are similar to those exposed in gully walls at Fåbergstølsbreen and Lodalsbreen. At this site, however, the stratigraphic relationship between the two associations is reversed. The upper association exhibits pronounced slope-parallel stratification and comprises stacked diamictons *c.* 0.3 m thick, separated by thin interbeds of sands and fine gravel (Figure 6.26). The upper association in section LE1 therefore possesses all the structural hallmarks of drift reworked by successive debris flows, and has strong affinities with the debris flow-reworked drift exposures in Figures 6.11-6.15 and the lower sediment association exposed in gully sidewalls at Fåbergstølsbreen and Lodalsbreen (Figures 6.22 and 6.23). The lower association in LE1, however, consists of a massive, structureless matrix-supported diamicton

LE1



V1

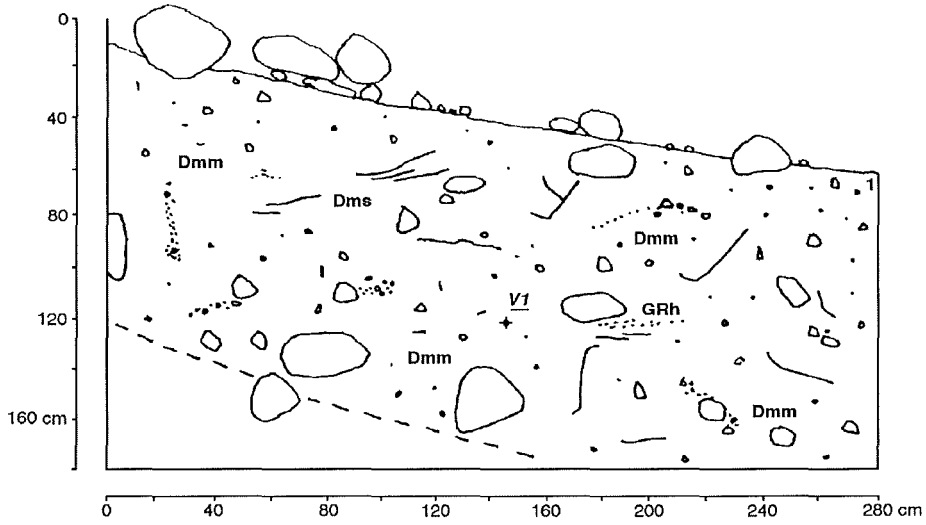


Figure 6.25. Sections LE1 and V1 showing structure and lithofacies exposed in gully sidewalls in Leirdalen (LE1) and Visdalen (V1), and location of clast fabric samples. No vertical exaggeration; key in Figure 5.13.

c. 1 m thick, and clasts in this association display a preferred down-valley orientation (Figure 6.24). These features suggest that the lower sediment association at section LE1 is an unreworked till, and the sediment sequence as a whole records paraglacial reworking to a depth of about 1.4 m of drift exposed by glacial retreat in Preboreal times. Section V1 resembles the lower unit of section

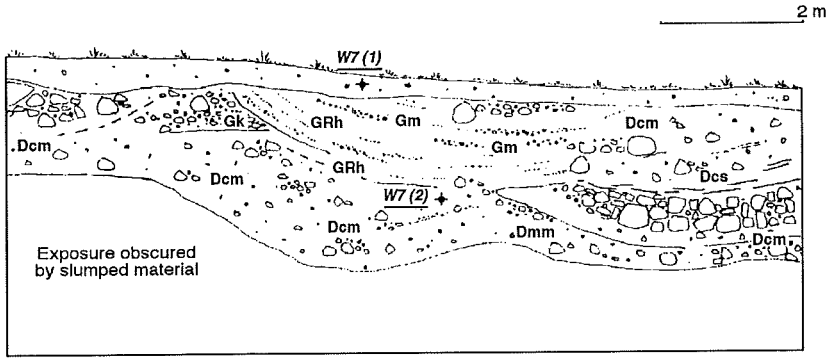
Figure 6.26. Slope-parallel stratification revealed in section LE1, Leirdalen. Stacked and stratified diamictons are separated by thin interbeds of sands and fine gravel, and are interpreted as ancient paraglacial deposits.

LE1, and is similarly dominated by a massive diamicton with only localised structural features, particularly small clusters of clasts and stringers of fine gravel with variable dip (Figure 6.25). Clasts in this section again exhibit a preferred orientation parallel to the valley axis (Figure 6.24), confirming that the sediments in this section are *in situ* glacigenic deposits that have not been reworked by debris flow activity.

6.4.3 Case study 3: Slope-foot deposits on the Isle of Skye

In the Western Red Hills of Skye, glacial drift of Loch Lomond Stadial age mantles lower slopes and valley floors. Two particularly clear slope-foot sections (W7 and W8) are shown in Figure 6.27. Section W7 reveals interstratified diamictons, gravels and boulder beds that exhibit the crude, subparallel stratification typical of sediments reworked by flow, but with complex

W7



W8

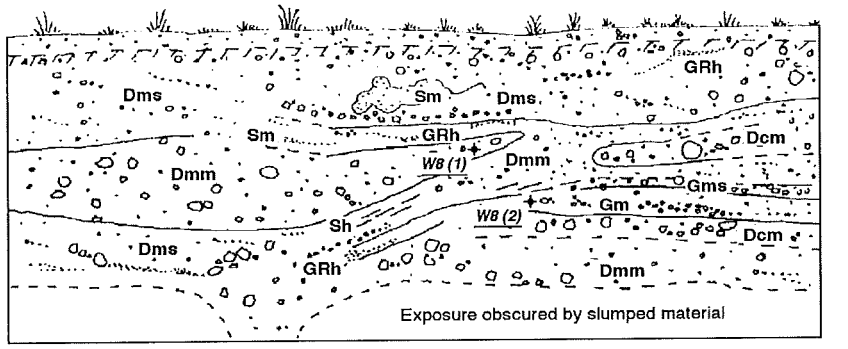


Figure 6.27. Sections W7 and W8 showing structure and lithofacies exposed in slope-foot sections in the Western Red Hills. Location of clast fabric samples is also shown. No vertical exaggeration; key in Figure 5.13.

cross-cutting and truncation of units that is absent from the reworked deposits described above. The diamicton units are clast- or matrix-supported, and are conformably interbedded with boulder-rich beds and poorly-sorted openwork gravels, and less frequently with well-sorted beds of fine granules and coarse sands. Clast fabrics measured at two levels record weak to moderate preferred orientations of clasts in both down-valley and downslope directions (Figure 6.24). Taken together, the complex structure of this deposit and the contrasting clast fabrics appear consistent with an interpretation involving penecontemporaneous release of debris from the deglaciated valley side and from glacier ice, as

proposed for this area by Benn (1991, 1992). Section W8 comprises interbedded stratified and massive matrix-supported diamictons, massive gravels, bedded gravels and sands. The contacts between units are mostly conformable but exhibit truncation in the centre of the section. Fabric measurements revealed a weak to moderate preferred downslope orientation of clasts in both the diamicton and imbricate gravel facies (Figure 6.24). Although the structural characteristics and fabric measurements are consistent with interpretation of this deposit as the product of reworking by slope processes, the evidence is insufficiently strong to dismiss a more complicated interpretation involving short-distance remobilisation of drift during initial emplacement.

6.5 Paraglacial modification of glacigenic sediments: implications.

A number of wider implications emerge from this study. The first concerns the sedimentological similarity between glacial and paraglacially-reworked drift deposits. Through comparison of *in situ* drift deposits with recent paraglacial debris flow deposits it has been shown that many features of paraglacial debris flow sediments are difficult to distinguish from those of *in situ* drift deposits, including clast shape, angularity and texture, matrix granulometry, and packing. It therefore seems likely that some hillslope or slope-foot deposits previously mapped as tills (especially flow tills) may actually be paraglacial debris flow deposits (e.g. Wright, 1983, 1991; Eyles *et al.*, 1988; Lawson, 1988; Fernlund, 1994; Owen, 1994), and that the extent of paraglacial resedimentation of till may have been greatly underestimated (*cf.* Dardis *et al.*, 1994). As the above examples demonstrate, however, whilst some exposures display ambivalent sedimentary features or highly-complex sedimentologies, detailed logging of sediment units in combination with clast fabric measurements appears to offer a reasonably robust means for distinguishing glacigenic drift that has been

reworked by debris flows and other slope processes from that which has remained intact since deposition.

A related implication is that detailed sedimentological analyses of hillslope or slope-foot drifts have potential for the reconstruction of ancient paraglacial remobilisation and resedimentation of drift following glacier retreat in formerly-glaciated mountain environments. The sedimentological similarities between the recently-reworked deposits investigated in section 6.3 and ancient deposits underlying valley-side drift slopes in Jotunheimen, for example, suggest that the latter formerly experienced reworking processes similar to those currently operating on the forelands of Fåbergstølsbreen and Lodalsbreen. Moreover, the rapidity of recent (post-'Little Ice Age') paraglacial landscape modification by debris flow activity at the Jostedalsbre sites suggests that deglaciation in Late Pleistocene or Early Holocene times may also have been followed by similar rapid paraglacial landscape changes of a timescale of a few decades (*cf.* Ballantyne and Benn, 1994, 1996; Harrison and Winchester, 1997). The evidence for apparent interdigitation of drift derived from glacier ice and adjacent hillslopes in the Western Red Hills (section W7, Figure 6.27) provides direct evidence for extremely rapid reworking of hillslope drift during or immediately after deglaciation in Late Pleistocene times.

A third implication of this study concerns cycling of glacial (i.e. *in situ*) and paraglacial (i.e. reworked) deposits. At the Jostedalsbre sites the preservation of thick debris flow deposits under unreworked 'Little Ice Age' drift (Figures 6.22 and 6.23) indicates extensive paraglacial resedimentation of older glacigenic sediments prior to the 'Little Ice Age' advance. As the latter was the most pronounced Neoglacial event in the area, this earlier phase of extensive paraglacial resedimentation must have followed glacier retreat in late Preboreal times. (A similar pattern of Preboreal paraglacial resedimentation is inferred from

section LE1 in Leirdalen, Jotunheimen, where debris flow sediments up to 1.4 m thick (Figure 6.25) have been deposited *c.* 120 m above the valley floor, well outside the limits of Neoglacial advances). The sediment sources for current paraglacial reworking of drift in the forelands of Fåbergstølsbreen and Lodalsbreen therefore include not only glacial deposits emplaced by the 'Little Ice Age' advance, but also ancient (late Preboreal) paraglacial sediments. Furthermore, the truncation of paraglacial sedimentary units of inferred Preboreal age by glacial erosion during the 'Little Ice Age' advance implies that these ancient paraglacial deposits contributed sediment to the glacier transport system. It appears, therefore, that glacial and paraglacial reworking of sediment can be seen as alternating modes of sediment transfer, the former dominant during glacier advance and the latter following glacier retreat (*cf.* Ballantyne and Benn, 1994; Fitzsimons, 1996).

6.6 Summary.

This chapter aimed to establish the sedimentological consequences of paraglacial modification of drift. Sedimentological criteria that distinguish glacigenic drift deposits reworked by debris flows from unreworked, *in situ* glacigenic drift deposits have been identified and applied to the interpretation of deposits of uncertain (reworked or *in situ*) origin. The principal findings are summarised below:

1. Glacigenic drift reworked by paraglacial slope processes (primarily debris flow) retains many of the sedimentological characteristics of parent material, and cannot be readily distinguished from *in situ* till in terms of systematic differences in clast shape, angularity or texture, fine-fraction granulometry or packing (void ratio). The combined use of clast fabric analysis (preferred orientation) and large-scale structural and lithofacies

relationships, however, appears to offer a reasonably robust method for differentiating reworked from *in situ* glacigenic deposits.

2. Application of the above criteria to the investigation of steep valley-side drift deposits at two recently-deglaciated sites in the Jostedalen area has revealed that valley-side deposits comprise *in situ* glacial drift emplaced during the 'Little Ice Age' overlying a crudely stratified diamicton that represents paraglacial reworking of glacigenic deposits by debris flows following deglaciation in late Preboreal times. Both sediment associations are currently subject to paraglacial reworking. Cyclic alternation of glacial and paraglacial sediment transfer is implied. Use of the same criteria to differentiate sediment associations in Leirdalen, Jotunheimen, has demonstrated that unreworked till is overlain by at least 1.4 m of paraglacial debris flow deposits, implying widespread reworking of sediment and associated drift slope modification following deglaciation in Preboreal times. In the Western Red Hills on the Isle of Skye, evidence for juxtaposition of melt-out deposits and reworked valley-side sediment indicates penecontemporaneity of paraglacial remobilisation of drift and deglaciation at the end of the Loch Lomond Stade.

3. The above findings suggest that paraglacial reworking of steep, drift-mantled slopes during and immediately after deglaciation is a widespread phenomenon. By analogy with the rapid paraglacial resedimentation evident on recently deglaciated terrain, it is likely that most reworking of glacigenic drift by hillslope processes at the end of the last glacial stage occurred within decades or at most centuries of deglaciation.

Chapter 7

Paraglacial modification of slope form

7.1 Introduction.

The research reported in the previous chapter of this thesis considered the sedimentological implications of paraglacial reworking of hillslope drift deposits. In this chapter the morphological consequences of paraglacial activity are examined. Following an outline of field and analytical methods (section 7.2), this chapter describes how drift slope long profiles are modified by gullying (section 7.3), explains how gullies evolve through time (section 7.4), and assesses the rates of slope adjustment involved (section 7.5). Summary findings are outlined in section 7.6.

Paraglacial slope adjustment is a stabilising response to disequilibrium generated within drift slopes in a dynamic environment. However, it is notable that most studies of paraglacial resedimentation of glacigenic drift have paid little attention to the effects of paraglacial activity in modifying slope form. Owen *et al.* (1995) observed that paraglacial erosion of the sidewalls of drift gorges in the Lahul Himalaya was by parallel retreat of gully heads, but provided no information concerning the precise form and rate of such slope modification. Several workers have recognised that snow and slush avalanches and debris flow activity can cause substantial modification of slope form by translocating debris downslope and thereby creating slopes of pronounced overall concavity with gradients lower than those of unaffected slopes. However, many of these studies concern reworking of rockfall talus (e.g. Luckman, 1971, 1972, 1977, 1978; Gray, 1973; Kotarba, 1976; Church *et al.*, 1979; Ballantyne and Eckford, 1984;

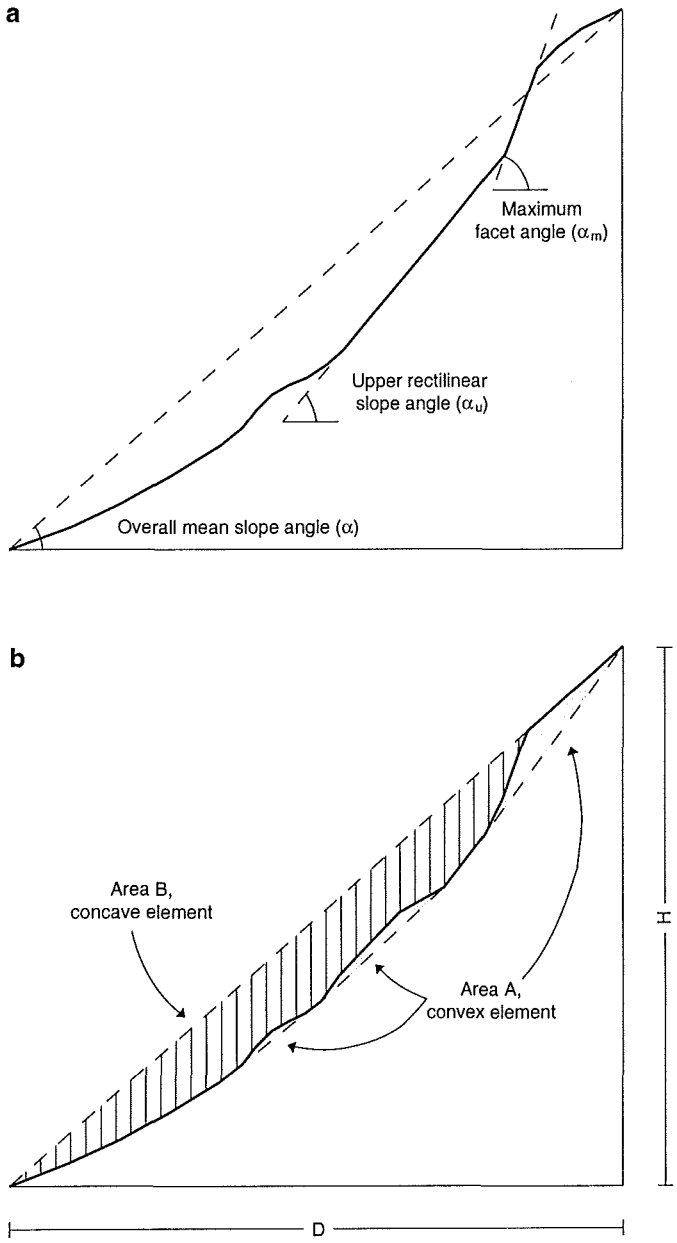
Hinchliffe, 1998) or pyroclastic deposits (e.g. Okuda *et al.*, 1980; Suwa and Okuda, 1980) rather than glacigenic drift, and hence cannot be regarded as investigations of 'paraglacial' slope adjustment. Harvey (1982; 1986; 1987; 1992) has described detailed investigations of gully development on soliflucted drift deposits in northwest England, and in badland environments in southern Spain, but these features are frequently influenced by basal incision and differ quite markedly from the much larger valley-side features at the Norwegian and Scottish field sites. Research which specifically considers paraglacial modification of the form and behaviour of steep drift slopes is largely confined to earlier work carried out in Fåbergstølsdalen and Bergsetdalen by Ballantyne and Benn (1994, 1996; Ballantyne, 1995a). These authors found that paraglacial drift modification in Fåbergstølsdalen since AD 1943 had led to localised reduction of overall slope gradients by *c.* 5°, and that slope-foot debris cone accumulation had contributed to a general reduction in overall slope concavity. They also calculated minimum average rates of slope surface lowering due to gullying at particular sites of 50-100 mm yr^{-1}, and estimated a maximum rate of 200 mm yr^{-1} at one location. The wider representativeness of their findings is unknown, however, and in particular their predictions as to how paraglacial gully systems evolve remain untested. The research reported below addresses these issues by investigating the nature and rate of paraglacial slope adjustment under a wide range of field situations in Norway and Scotland.

7.2 Methods.

As demonstrated in sections 4.4 and 5.4, the dominant process of paraglacial drift slope adjustment at the field sites investigated is debris flow activity, which has commonly contributed to marked gully incision. Assessment of paraglacial modification of slope form at these sites involved both instrumental survey of slope profiles and measurement of the dimensions of gullies incised into

hillslope drift, as outlined in section 4.2. Slope profile measurements and gully survey data were obtained using an EDM at Fåbergstølsbreen, Leirdalen, the Western Red Hills, Glen Einich and the Pass of Drumochter, and by Abney level at Lodalsbreen, søre Illåbreen, Visdalen, Heillstugubreen and Glen Docherty. At each field site, parallel slope profiles were surveyed up gully floors and adjacent unmodified slopes to allow changes in slope form associated with gully development to be established. To allow comparative quantitative analysis of the profiles of paired gullied and ungullied drift slopes, three measures were calculated to describe slope gradient (Figure 7.1a) and two indices used to assess overall slope geometry (Figure 7.1b). The overall mean slope angle (α) measures the average gradient of the entire drift slope profile from foot to crest. The form of the upper straight slope is summarised by the upper rectilinear slope angle (α_u), defined as the average slope angle excluding any upper slope convexity and the basal concavity (the lower part of the overall slope, characterised by a consistent downslope decrease in gradient). Finally, the maximum facet angle (α_m) represents the steepest gradient recorded for any section of slope along each profile. The indices of concavity (c) and linearity (l) employed here were proposed by Church *et al.* (1979) to describe overall debris slope morphology. The concavity index c is calculated in terms of the ratio of convex (A) and concave (B) elements of the slope (Figure 7.1b), and effectively reflects the depth of overall slope concavity. The linearity index l describes the departure from linearity of the slope, in terms of the sum of concave and convex elements divided by HD/2, where H and D respectively represent the vertical height and planimetric length of the slope (Figure 7.1b). Calculations of the areas A and B for each profile were obtained by digitising the slope profile data.

Transects were also surveyed across the gullies to allow calculation of gully volume and hence implied sediment loss, and to allow trends in gully form to be identified. Data describing gully depth, length, width and volume have been



Standardising factor F = HD/2 (the area of the right-angled triangle defined by the limits of the slope)

l: A+B/F indexes the departure from linearity — straight: sum < 0.03

c: A/B indexes the convexity/concavity of the slope — concave: A/B ≤ 0.125

concave, minor convexity: 0.125 < A/B ≤ 0.75

convex-concave: 0.75 < A/B ≤ 1.25

convex, minor concavity: 1.25 < A/B ≤ 8.75

Figure 7.1. Definition of slope morphology parameters referred to in the text: (a) slope gradient, and (b) slope geometry (after Church *et al.*, 1979).

referred to in chapters 4 and 5, and are summarised in Tables 4.1 and 4.2 for the Norwegian sites, and in Table 5.1 for the Scottish field sites. Calculations of sediment removed since gully initiation (gully volume) were divided by the gully area to estimate average surface lowering since deglaciation. Because the timing of deglaciation represents a maximum age for gully initiation, rates of surface lowering implied from these figures are expressed as minima. To estimate short-term rates of sediment removal and accumulation in active paraglacial gully systems, 51 steel erosion pins measuring 1m in length were inserted into gully floors and sidewalls and across cone surfaces at Fåbergstølsbreen in June 1996 and 24 were re-measured in July 1997. The use and limitations of this technique in the study of slope evolution are summarised by Haigh (1977).

Finally, to permit comparison of the characteristics and dimensions of gullies at different stages of evolution, gully cross-sectional data from field sites in Norway were placed into three age categories: T_1, T_2 and T_3. Category T_1 includes all the surveyed gullies which lie inside 'Little Ice Age' glacier limits and are close to the present-day ice margins (gullies Fa-Fc, La-Lc, Sa-Sb and Ha-Hd in Figures 7.2-7.7); category T_2 incorporates gullies furthest from the present glacier margin but still within 'Little Ice Age' glacier limits (Fd, Fe and Ld); and category T_3 comprises gullies surveyed outside the maximum limit of 'Little Ice Age' glacier cover (LEa-LEe and Va-Ve). Although there may locally have been some delay in gully initiation following exposure of valley-side drift by glacier retreat (as occurred in Fåbergstølsdalen; see section 4.6.1 above), the maximum potential age of the onset of formation of the T_1 gullies is less than that of the T_2 gullies, which is in turn markedly less than that of the T_3 gullies, given the different ages of drift exposure following deglaciation. Comparison of the dimensions and morphological characteristics of the three categories of gullies therefore provides the opportunity to investigate the nature of gully evolution through time.

7.3 Slope profile adjustment.

A total of 45 pairs of ungullied and gullied slope profiles are depicted in Figures 7.2 to 7.11 and illustrate the modification in profile characteristics due to gullying at each of the field sites investigated. With the exception of slopes in Leirdalen and Visdalen, the general form of most of the ungullied drift slopes is that of an approximately straight upper slope resting at gradients between *c.* 26° and *c.* 41° and a basal concavity, although the relative length of these two slope units varies from profile to profile. Many of these unmodified drift slopes possess an overall form resembling that of rockfall talus slopes (Statham, 1976b; Francou and Manté, 1990), though the profiles investigated here result predominantly from the collapse of lateral moraines rather than the accumulation of rockfall debris. Drift slopes in Leirdalen and Visdalen typically have much shallower gradients than those surveyed at other sites. At sites where gully erosion is still active, the effects of paraglacial erosion and redeposition on slope form record only an intermediate stage in paraglacial slope modification; nevertheless, modification of drift slope gradient and geometry is evident to some extent at all sites, and is discussed below.

7.3.1 Modification of drift slope gradient

Changes in slope gradient that result from paraglacial gully erosion and redeposition of eroded drift are summarised in Tables 7.1 and 7.2. Overall mean slope gradients (α) have been reduced slightly from 13.3-38.5° to 12.2-35.6°, showing a reduction in overall gradient for each set of slope profiles. This trend results from incision of the mid-upper drift slope, and accords with the observations of Ballantyne and Benn (1994), who calculated a reduction in overall mean slope gradients from 27-29° to 24-26° on three sets of slope profiles in upper Fåbergstølsdalen. The reduction in overall slope gradients observed

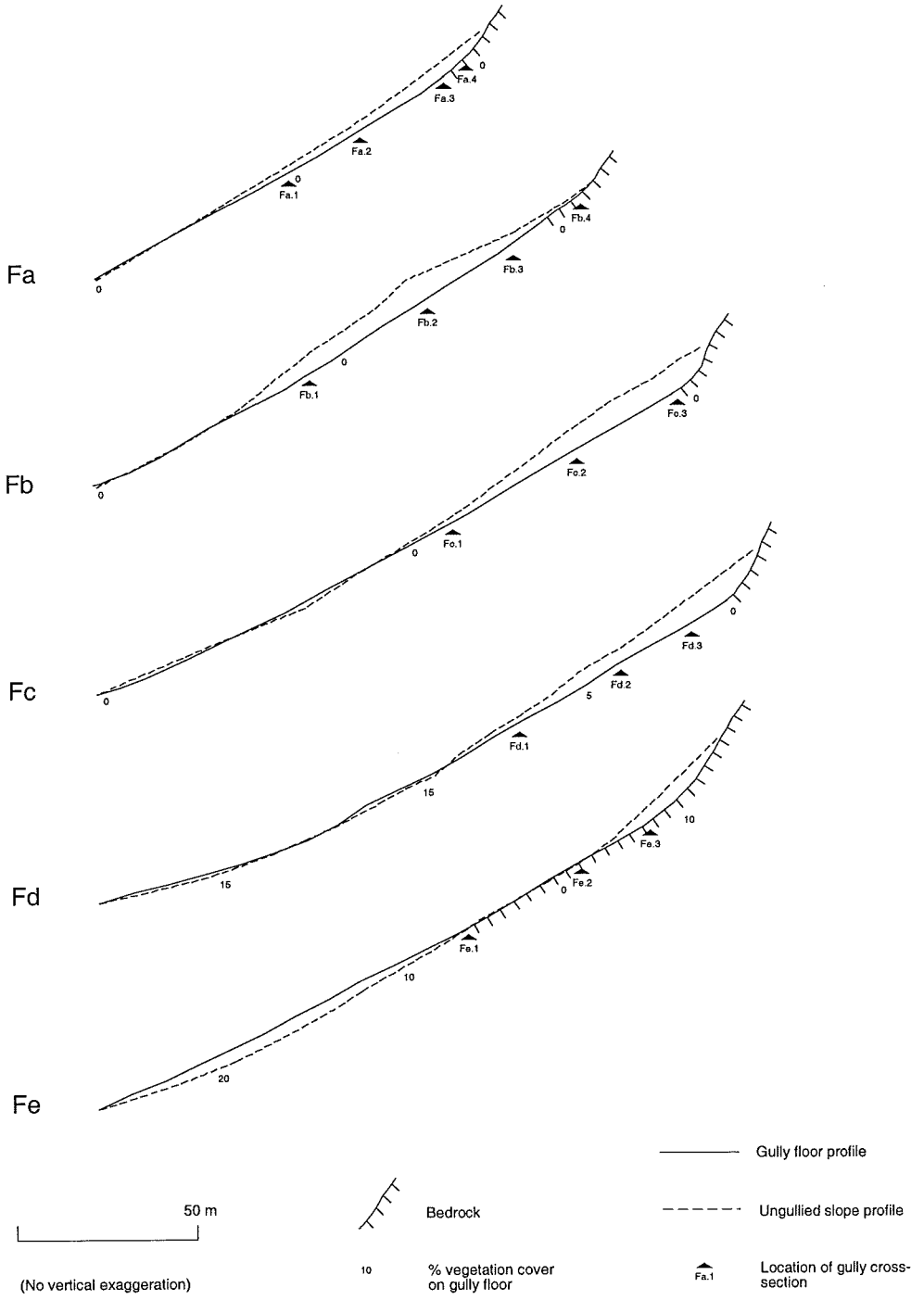


Figure 7.2. Slope profiles Fa-Fe on the foreland of Fåbergstølsbreen, Norway.

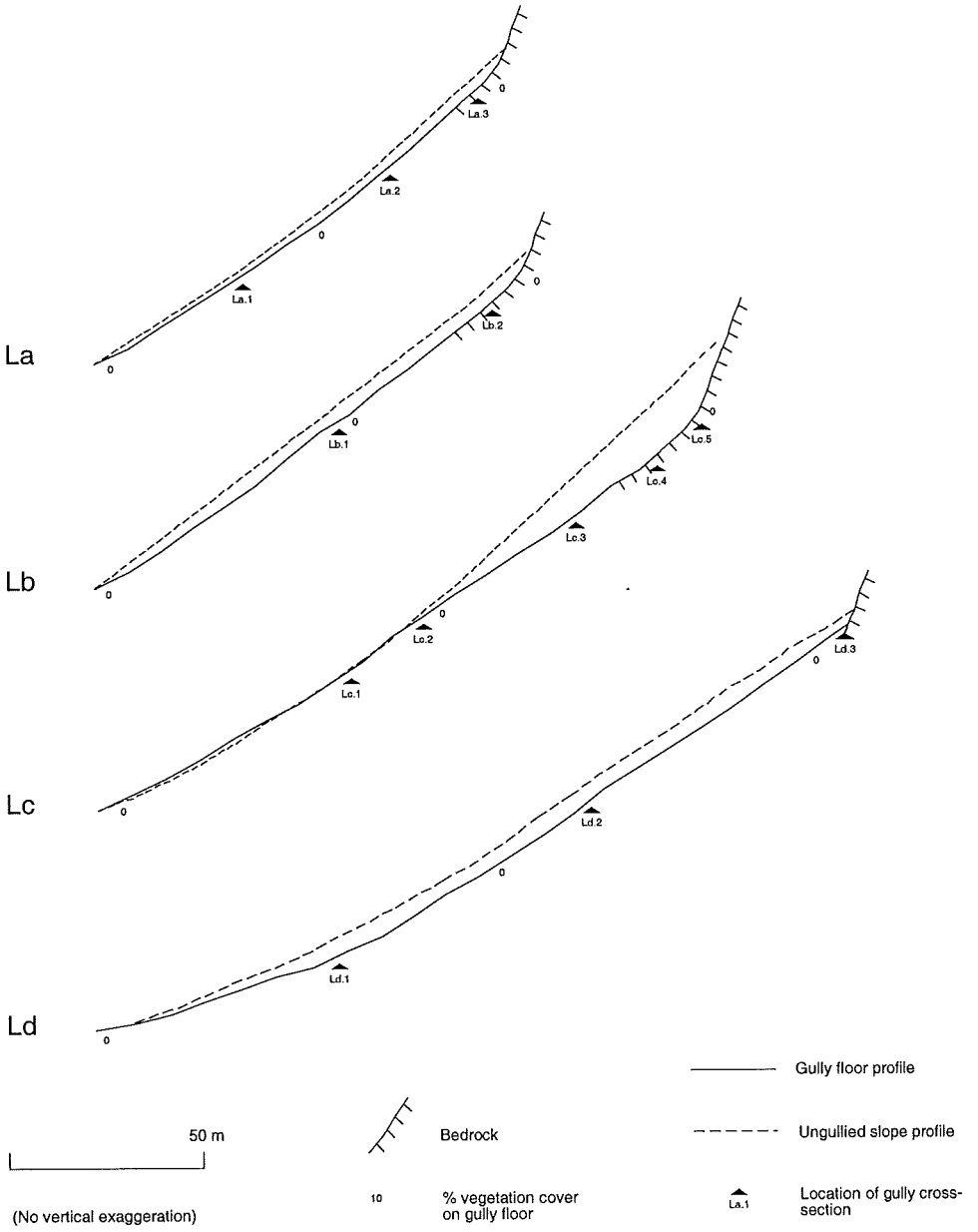


Figure 7.3. Slope profiles La-Ld on the foreland of Lodalsbreen, Norway.

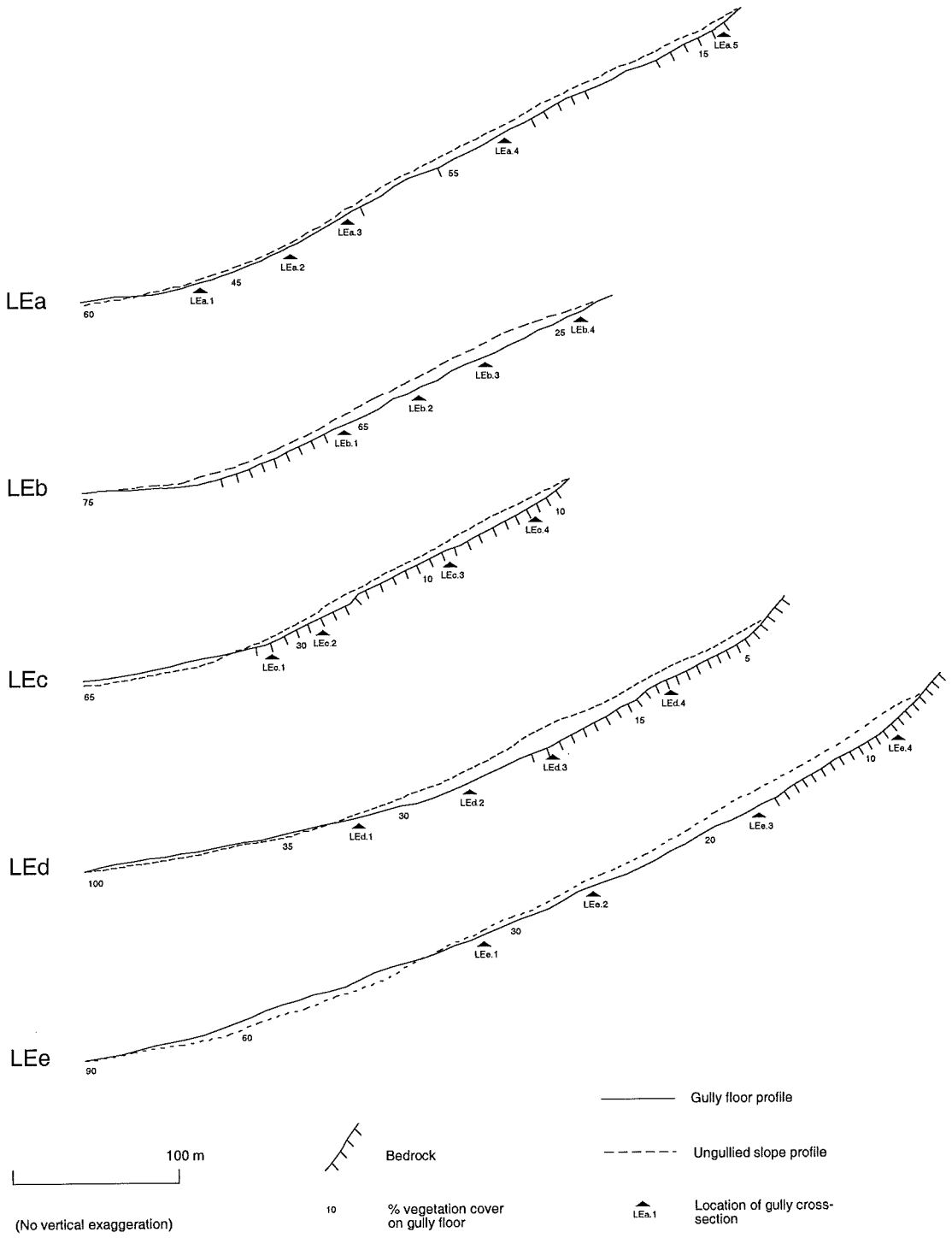


Figure 7.4. Slope profiles LEa-LEe in Leirdalen, Norway.

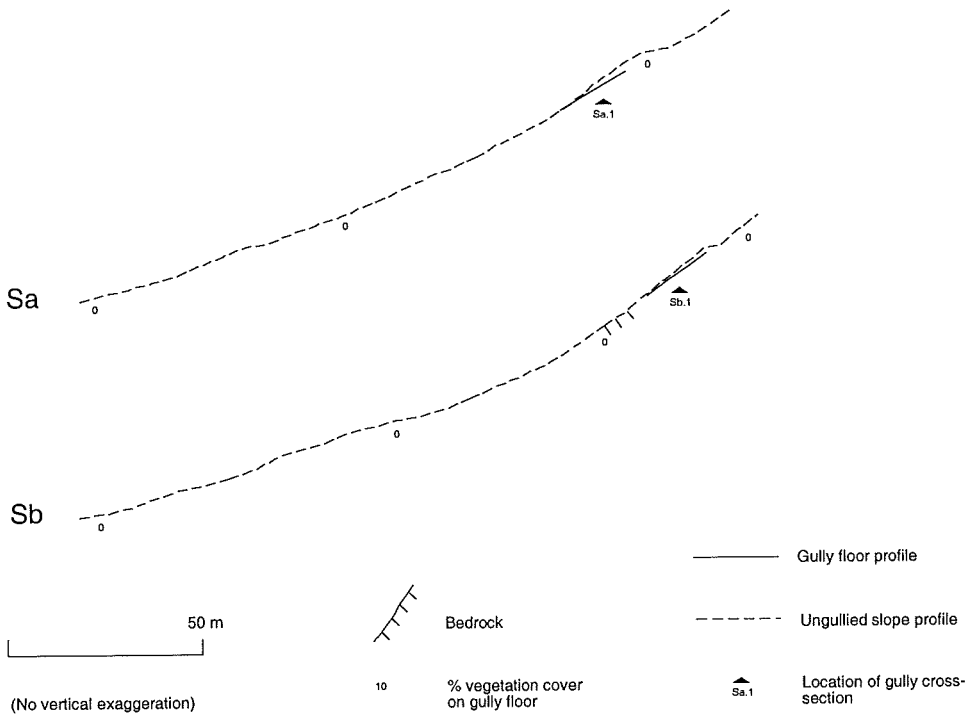


Figure 7.5. Slope profiles Sa-Sb on the foreland of søre Illåbreen, Norway.

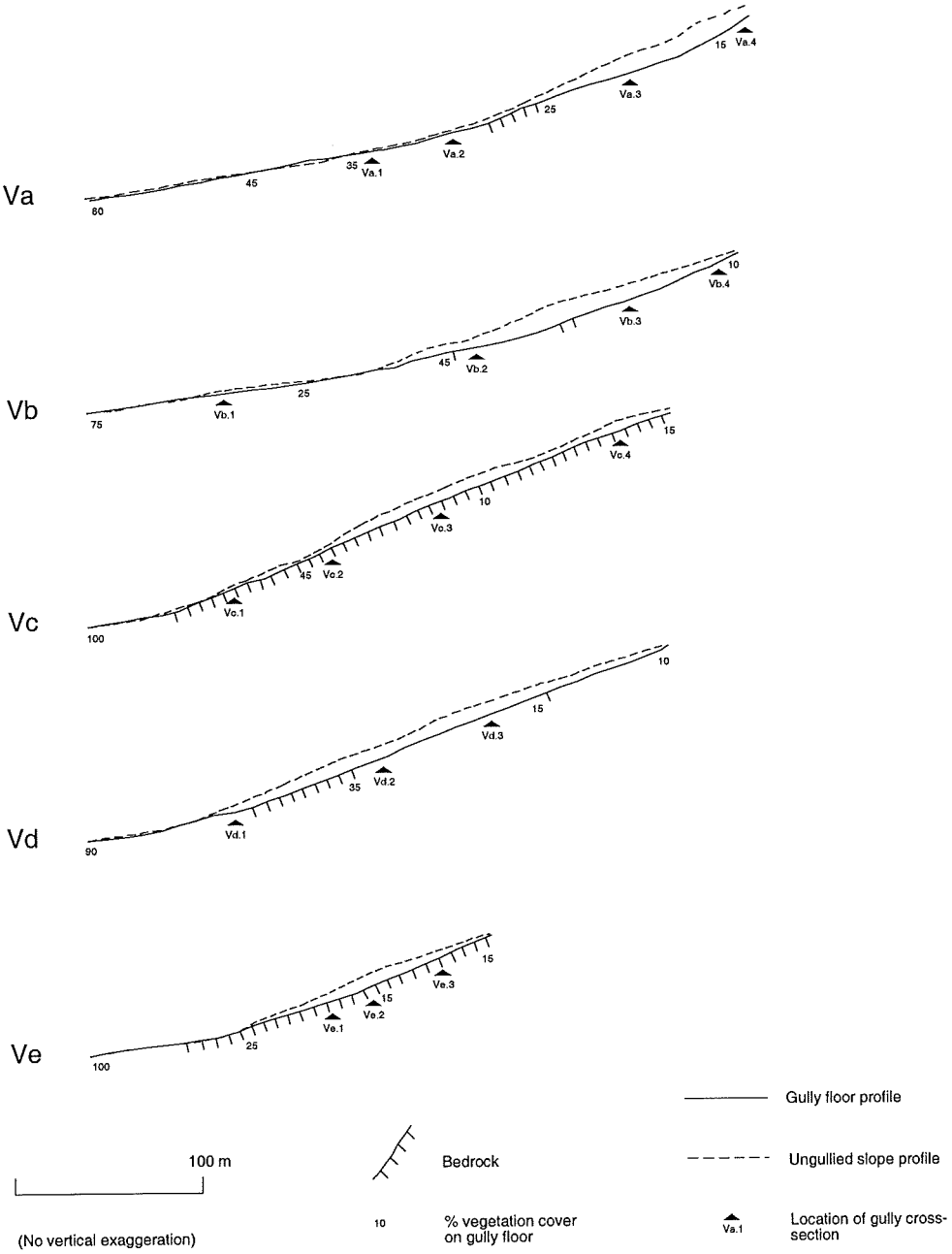


Figure 7.6. Slope profiles Va-Ve in Visdalen, Norway.

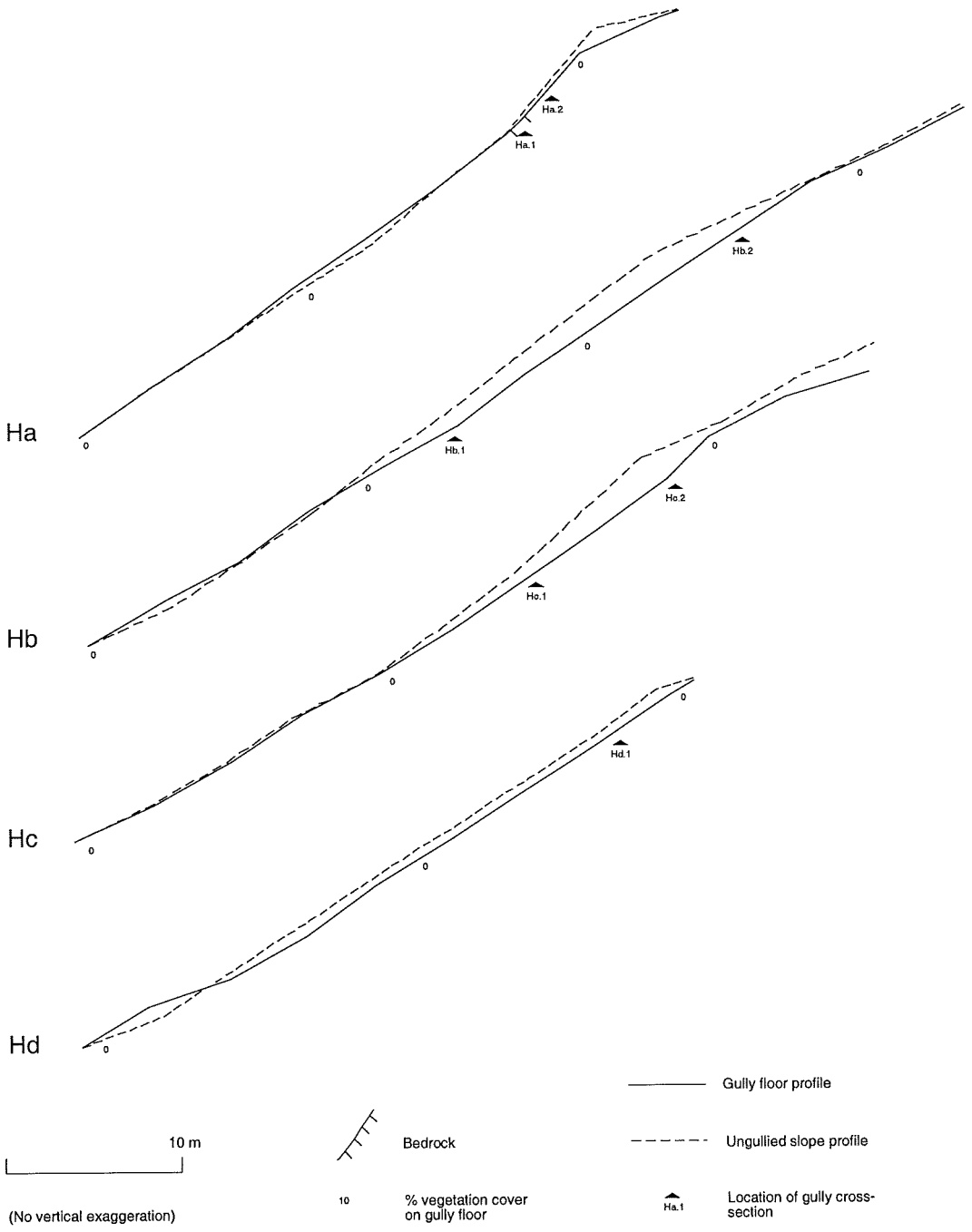


Figure 7.7. Slope profiles Ha-Hd on the foreland of Heillstugubreen, Norway.

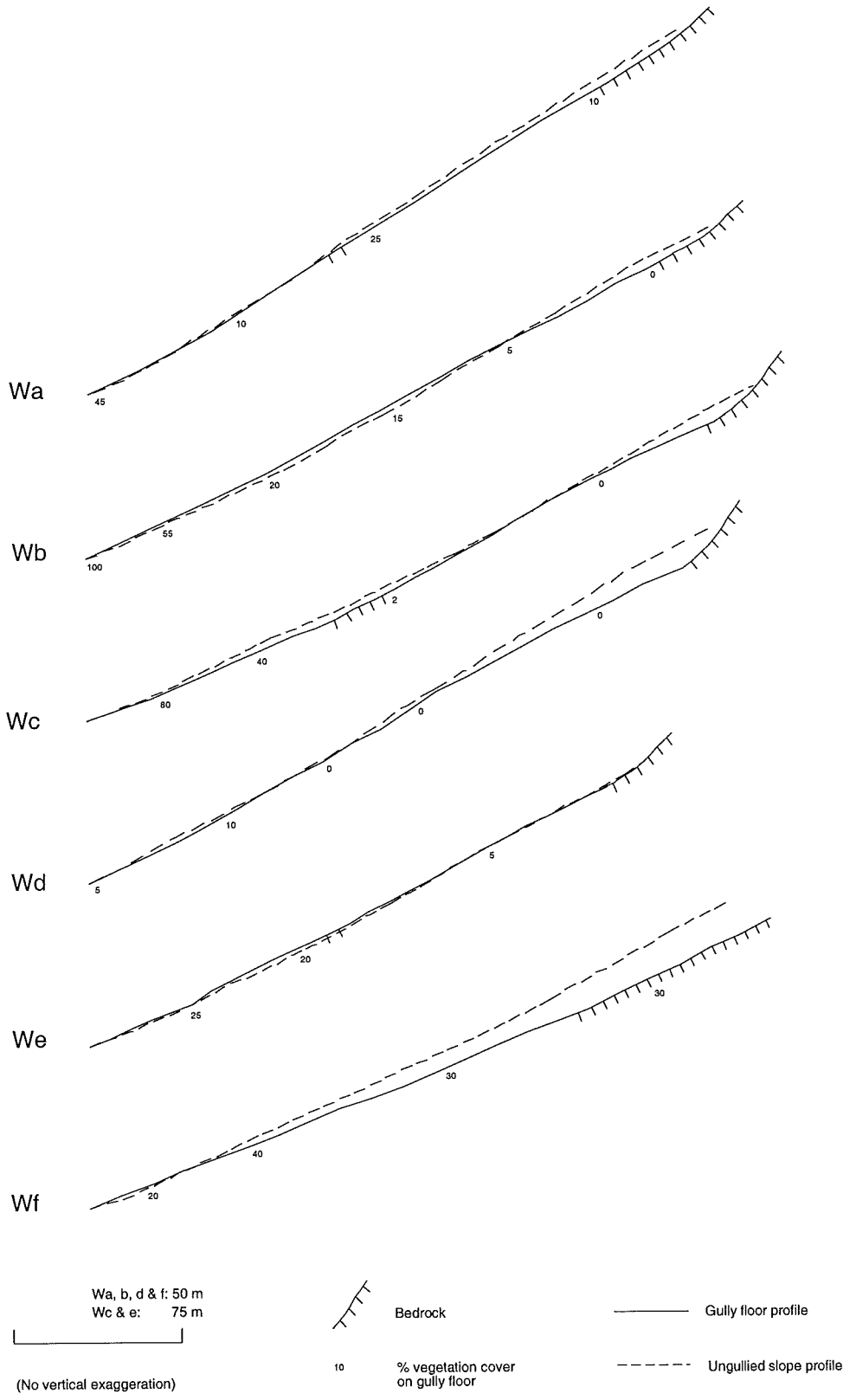


Figure 7.8. Slope profiles Wa-Wf in the Western Red Hills, Scotland.

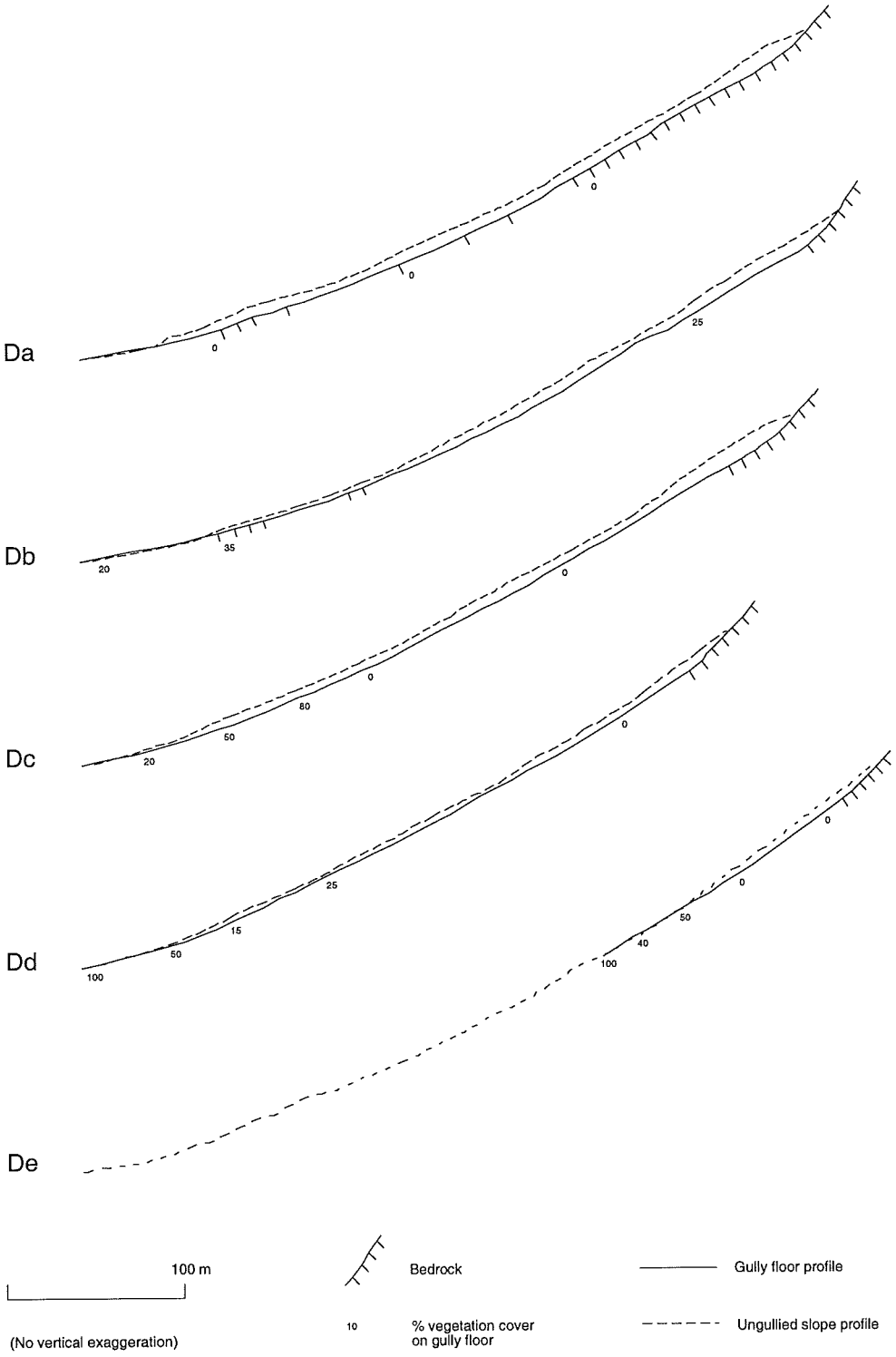


Figure 7.9. Slope profiles Da-Df in Glen Docherty, Scotland.

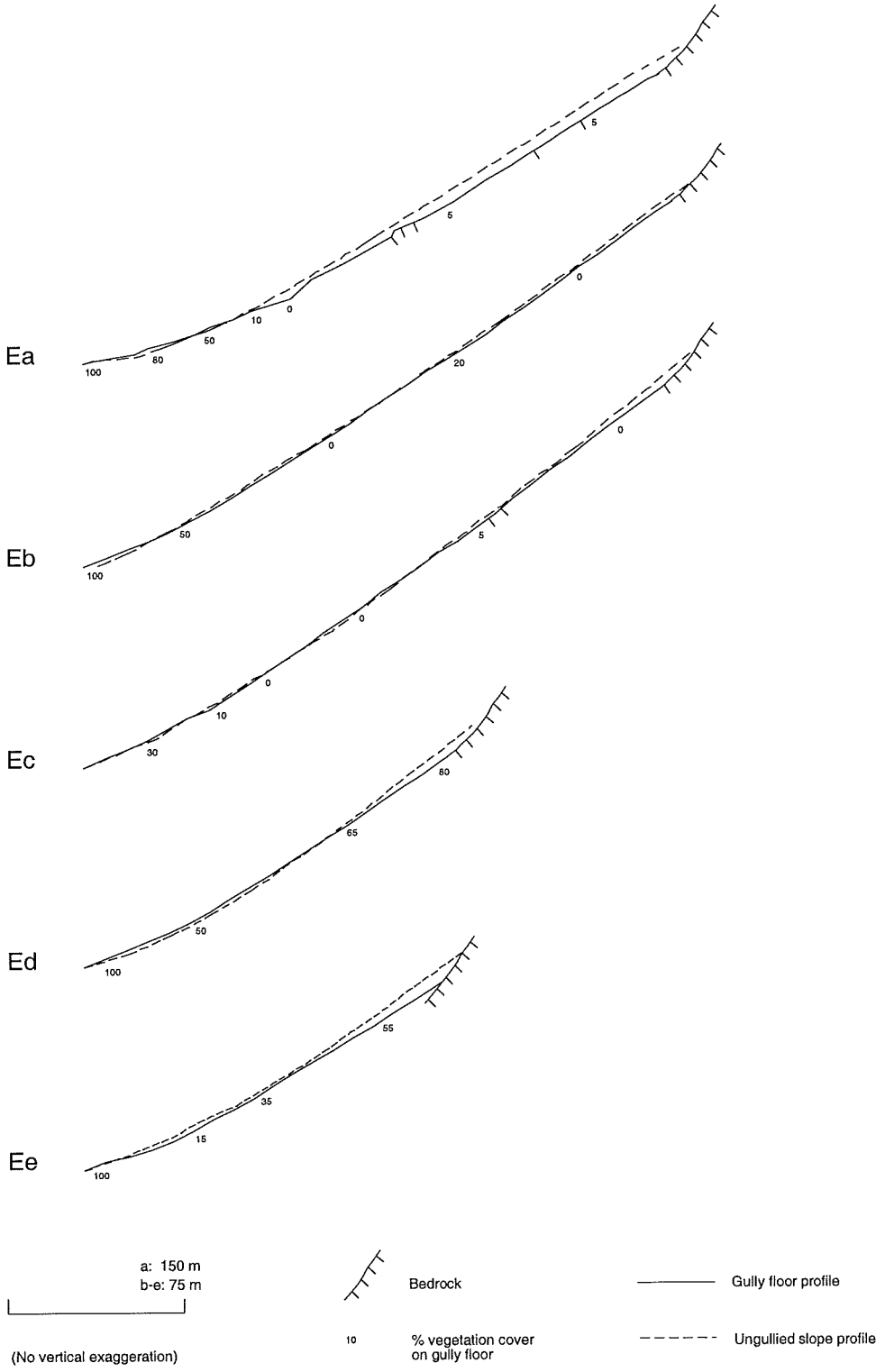


Figure 7.10. Slope profiles Ea-Ee in Glen Einich, Scotland.

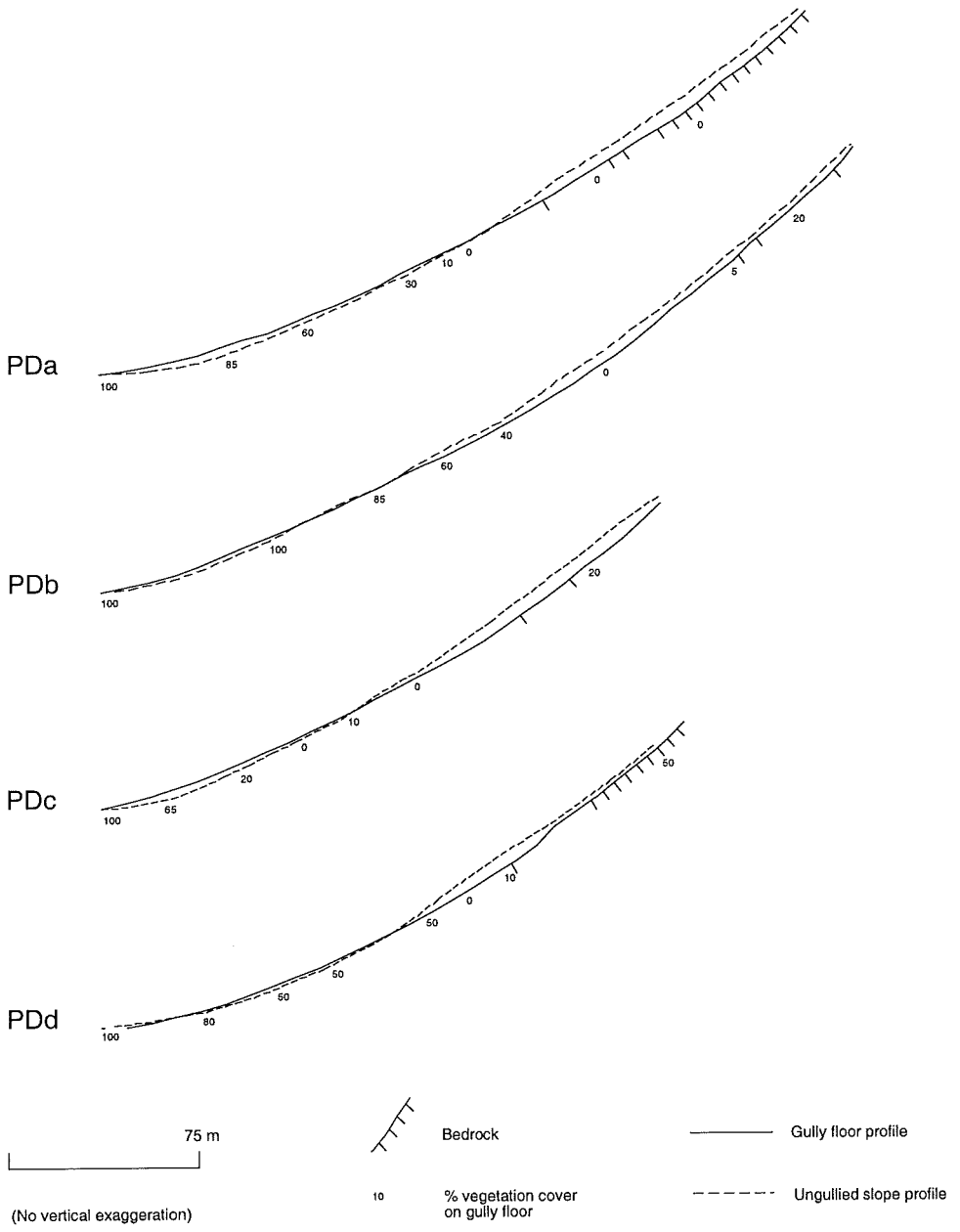


Figure 7.11. Slope profiles PDa-PDd in the Pass of Drumochter, Scotland.

Profile	Overall mean slope angle (α)			Upper rectilinear slope angle (α_u)			Maximum facet angle (α_m)		
	UD	GF	Δ	UD	GF	Δ	UD	GF	Δ
Fåbergstølsbreen									
Fa	32.4	31.3	-1.1	33.6	32.3	-1.3	36.4	39.2	2.8
Fb	31.2	31.0	-0.2	32.0	33.0	1.0	42.8	35.9	-6.9
Fc	29.9	27.9	-2.0	33.5	29.7	-3.8	37.3	32.6	-4.7
Fd	28.2	25.6	-2.6	34.4	29.9	-4.5	40.0	35.0	-5.0
Fe	29.9	28.3	-1.6	34.8	30.2	-4.6	34.0	34.0	0
Mean	*30.3*	*28.8*	*-1.5*	*33.7*	*31.0*	*-2.6*	*38.1*	*35.3*	*-2.8*
Lodalsbreen									
La	36.7	35.0	-1.7	41.2	38.9	-2.3	41.2	41.5	0.3
Lb	36.9	34.6	-2.3	38.3	36.9	-1.4	42.0	39.0	-3.0
Lc	36.1	32.2	-3.9	41.5	34.3	-7.2	41.5	40.0	-1.5
Ld	27.7	27.1	-0.6	32.3	33.0	0.7	35.0	37.0	2.0
Mean	*34.4*	*32.2*	*-2.1*	*38.3*	*35.8*	*-2.6*	*39.9*	*39.4*	*-0.6*
Leirdalen									
LEa	23.2	23.1	-0.1	25.8	25.9	0.1	31.0	32.0	1.0
LEb	19.9	19.0	-0.9	19.3	24.5	5.2	32.0	33.5	1.5
LEc	22.3	21.6	-0.7	26.2	26.6	0.4	31.5	29.0	-2.5
LEd	17.7	16.4	-1.3	24.8	24.5	-0.3	32.0	32.0	0
LEe	21.9	20.8	-1.1	28.2	28.1	-0.1	32.0	32.0	0
Mean	*21.0*	*20.2*	*-0.8*	*24.9*	*25.9*	*1.1*	*31.7*	*31.7*	*0*
søre Illåbreen									
Sa	35.9	31.4	-4.5	35.9	31.4	-4.5	39.5	32.0	-7.5
Sb	38.5	35.6	-2.9	38.5	35.6	-2.9	40.0	36.0	-4.0
Mean	*37.2*	*33.5*	*-3.7*	*37.2*	*33.5*	*- 3.7*	*39.8*	*34.0*	*-5.8*
Visdalen									
Va	14.4	12.2	-2.2	18.3	21.2	2.9	26.0	30.0	4.0
Vb	13.3	12.4	-0.9	15.8	16.9	1.1	25.0	24.0	-1.0
Vc	19.9	19.1	-0.8	19.5	21.2	1.7	33.0	26.0	-7.0
Vd	18.3	17.8	-0.5	18.1	20.1	2.0	24.0	29.0	5.0
Ve	17.7	17.0	-0.7	19.3	21.8	2.5	30.0	29.0	-1.0
Mean	*16.7*	*15.7*	*-1.0*	*18.2*	*20.2*	*2.0*	*22.8*	*27.6*	*0*
Heillstugubreen									
Ha	35.0	34.9	-0.1	36.7	36.3	-0.4	50.0	49.0	-1.0
Hb	30.8	30.5	-0.3	30.4	30.6	0.2	38.0	36.0	-2.0
Hc	32.4	31.0	-1.4	33.7	32.1	-1.6	43.0	37.0	-6.0
Hd	30.0	29.9	-0.1	30.0	30.9	0.9	37.0	36.0	-1.0
Mean	*32.1*	*31.6*	*-0.5*	*32.7*	*32.5*	*-0.9*	*42.0*	*39.5*	*-2.5*

Table 7.1. Gradient of ungullied drift slopes (UD) that represent the gradient of deglaciated drift slopes prior to modification by debris flow, gully floors (GF) that represent the gradient of paraglacially-modified valley-side slopes, and the difference between each (Δ) at the Norwegian field sites. For definitions of indices see text.

Profile	Overall mean slope angle (α)			Upper rectilinear slope angle (α_u)			Maximum facet angle (α_m)		
	UD	GF	Δ	UD	GF	Δ	UD	GF	Δ
Western Red Hills									
Wa	31.4	30.8	-0.6	32.1	31.7	-0.4	36.5	33.1	-3.4
Wb	27.8	26.9	-0.9	29.7	27.7	-2.0	32.0	31.5	-0.5
Wc	26.6	25.5	-1.1	29.5	26.2	-3.3	31.5	30.5	-1.0
Wd	29.5	27.6	-1.9	30.2	26.7	-3.5	32.5	32.0	-0.5
We	27.0	26.5	-0.5	28.6	27.4	-1.2	32.5	31.0	-1.5
Wf	22.9	20.4	-2.5	25.7	23.8	-1.9	27.0	27.5	0.5
Mean	*27.5*	*26.3*	*-1.3*	*29.3*	*27.3*	*-2.1*	*32.0*	*30.9*	*-1.1*
Glen Docherty									
Da	24.1	23.3	-0.8	31.6	31.0	-0.6	35.0	36.5	1.5
Db	24.4	23.0	-1.4	31.3	30.2	-1.1	36.0	35.0	-1.0
Dc	25.9	24.4	-1.5	32.7	29.4	-3.3	37.0	36.5	-0.5
Dd	27.0	25.9	-1.1	35.1	33.5	-1.6	38.0	39.3	1.3
De	34.9	33.1	-1.8	36.6	35.8	-0.8	38.0	37.0	-1.0
Mean	*27.3*	*25.9*	*-1.3*	*33.5*	*32.0*	*-1.5*	*36.8*	*36.9*	*0.1*
Glen Einich									
Ea	27.2	26.5	-0.7	31.6	31.1	-0.5	33.5	35.5	2.0
Eb	32.2	31.5	-0.7	35.6	34.2	-1.4	38.5	36.0	-2.5
Ec	33.5	33.3	-0.2	37.7	37.3	-0.4	36.0	40.0	4.0
Ed	31.5	30.7	-0.8	37.4	35.9	-1.5	39.0	38.5	-0.5
Ee	29.6	27.9	-1.7	35.1	32.0	-3.1	36.0	33.5	-2.5
Mean	*30.9*	*30.0*	*-0.8*	*35.5*	*34.1*	*-1.4*	*36.6*	*36.7*	*0.1*
Pass of Drumochter									
PDa	26.8	25.8	-1.0	35.4	34.8	-0.6	37.5	39.5	2.0
PDb	29.7	29.0	-0.7	40.4	40.1	-0.3	39.9	43.5	3.6
PDc	28.9	27.2	-1.7	37.0	36.9	-0.1	36.5	38.5	2.0
PDd	27.2	25.6	-1.6	38.2	38.6	0.4	34.5	47.0	12.5
Mean	*28.2*	*26.9*	*-1.3*	*37.8*	*37.6*	*-0.2*	*37.1*	*42.1*	*5.0*

Table 7.2. Gradient of ungullied drift slopes (UD) that represent the gradient of deglaciated drift slopes prior to modification by debris flow, gully floors (GF) that represent the gradient of paraglacially-modified valley-side slopes, and the difference between each (Δ) at the Scottish field sites. For definitions of indices see text.

between ungullied drift profiles and gully floors varies from 0.1° to 4.5°, and the median reduction in overall gradient at all sites is 1.1°. Profiles that exhibit only a slight reduction in overall gradient (e.g. Fb, Figure 7.2; LEa, Figure 7.4) are

generally characterised by a very shallow failure scarp at the gully head, very close to the level of the ungullied drift surface.

Investigation of the upper rectilinear slope angles of drift adjusting to paraglacial conditions is of particular interest, as this is usually the slope unit where incision is initiated and concentrated. As the data in Table 7.1 show, 32 of the 45 paired profiles show a decrease in the gradient of the upper straight slope after gullying, whilst 13 paired profiles record an increase in the gradient of the upper rectilinear slope after gullying. Plotting the change in rectilinear slope gradient ($\Delta\ \alpha_u$) between ungullied drift slopes and corresponding gully floors against initial rectilinear slope gradient reveals a negative relationship between the two variables significant at $p < 0.001$ (Figure 7.12a). This relationship reveals two trends: First, the upper rectilinear slopes which were initially steep (> *c.* 28° gradient) tend to exhibit a decline in gradient (by as much as 7.2° in profile Lc, Lodalsbreen) due to gully incision. This generally occurs where the upper rectilinear slope rests against a steep rock slope. Conversely, upper rectilinear slopes with initial gradients below *c.* 28° tend to exhibit steepening as a result of gullying. Many of the Leirdal, Visdal and Heillstugubre profiles record an increase in upper rectilinear slope gradient; in profile LEb (Figure 7.4), for example, the gully floor rectilinear slope angle is 5.2° steeper than that of the adjacent ungullied drift slope. At sites where upper slope gradient has increased through gullying, gully heads often take the form of shallow translational failures on gently-sloping drift (Figures 7.4, 7.6 and 7.7). In these cases, where the gully floor meets the ungullied slope at the crest of the drift slope, incision farther downslope inevitably steepens the upper slope gradient. These findings suggest that nature of paraglacial adjustment of slope profile is to some extent determined by initial slope gradient on steep terrain, and by slope configuration on more gently-sloping ground. Slope modification of both steep and relatively gentle drift slopes appears to tend towards a uniform gradient of *c.* 28°, which may

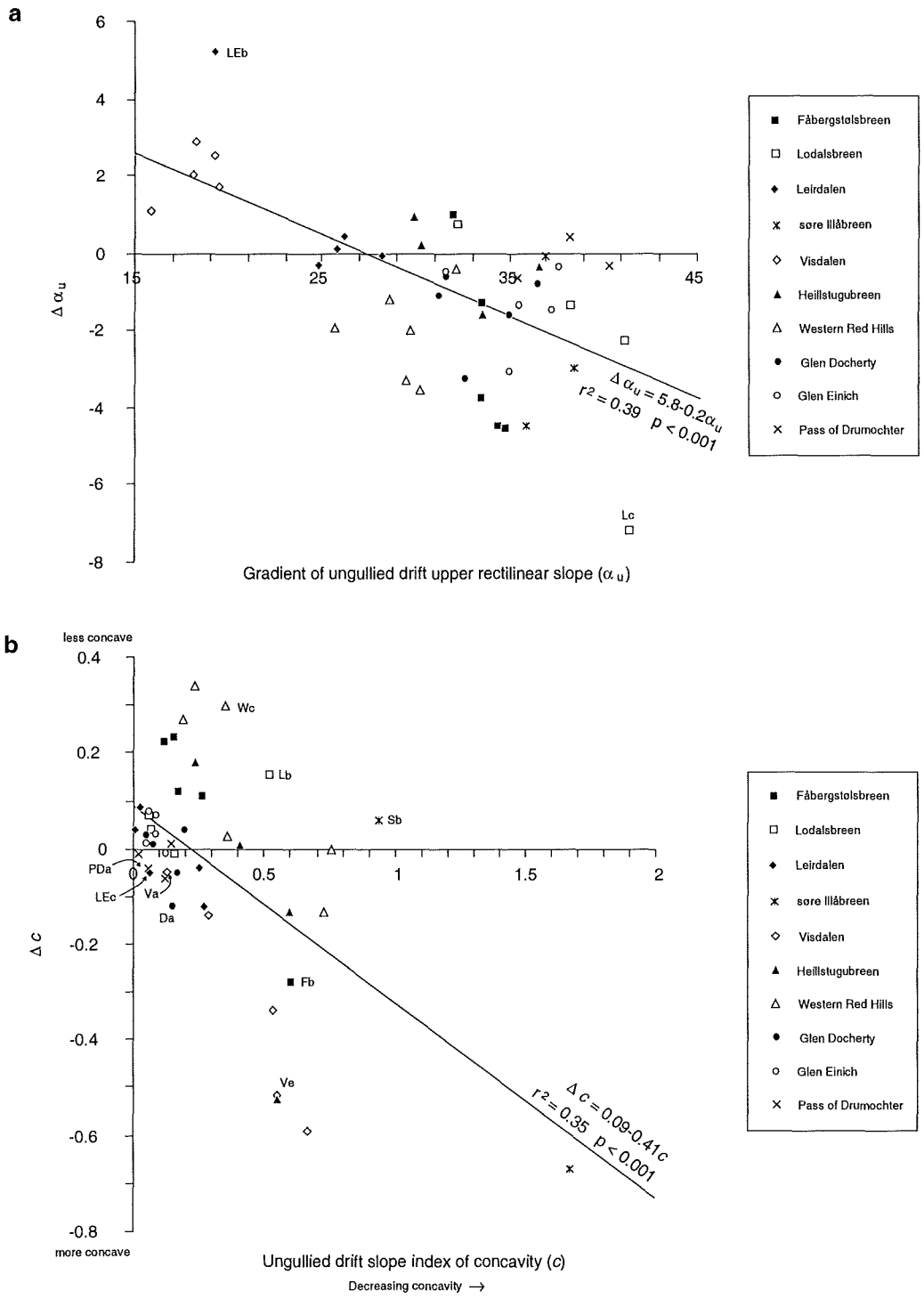


Figure 7.12. Scatterplots showing (a) change in upper rectilinear slope gradients ($\Delta\ \alpha_u$) between paired ungullied and gullied profiles against initial upper rectilinear slope gradients (α_u) at all field sites, and (b) change in the index of concavity ($\Delta\ c$) between paired ungullied and gullied profiles against initial index of concavity (c) values at all field sites. This difference represents the degree of change in (a) slope gradient and (b) slope geometry as a result of gullying. See text for definitions of terms.

represent an 'equilibrium' gradient for the upper rectilinear units of paraglacially-modified drift slopes, possibly related to the threshold gradient of slope failure and related movement of sediment by debris flows (*cf.* Innes, 1983a).

Changes in maximum facet angle (α_m) are much less regular. Gullying has reduced the maximum facet angle at most sections, but there are numerous exceptions (Table 7.1). Accumulation of debris at high angles against the risers of buried rock steps on gully floors commonly results in higher maximum facet angles than those recorded on adjacent ungullied slopes. This is particularly evident in Visdalen profiles Va and Vd (Figure 7.6). Similarly, at the Pass of Drumochter (Figure 7.11), the maximum slope facet gradient in profile PDd has increased by 12.5° as a result of debris flow incision exposing steep (47°) bedrock steps in the gully floor.

7.3.2 Modification of drift slope form

The effects of paraglacial activity on slope geometry are outlined in Tables 7.3 and 7.4 for each of the field sites. As these data show, there is considerable spread in the index of concavity and linearity values for profiles surveyed up ungullied drift, with c values ranging from 0.01 to 1.67, and l values ranging from 0.03 to 0.30. According to the criteria of Church *et al.* (1979), 15 of the 45 ungullied slope profiles are classified as 'concave' ($A/B \leq 0.125$) and 27 are described as 'concave, minor convexity' ($0.125 < A/B \leq 0.75$). Only two ungullied slope profiles (Sb, Figure 7.5, and Wd, Figure 7.8) fall within the 'convex-concave' range ($0.75 < A/B \leq 1.25$), and one (Sa, Figure 7.5) is termed 'convex, minor concavity' ($1.25 < A/B \leq 8.75$). None of the profiles is defined as 'straight' ($A+B < 0.03$). Changes in the concavity and linearity indices due to gully and redeposition of sediment downslope are discussed below.

Profile	Index of concavity (*c*)			Index of linearity (*l*)		
	UD	GF	Δ	UD	GF	Δ
Fåbergstølsbreen						
Fa	0.17	0.29	0.12	0.08	0.10	0.02
Fb	0.60	0.32	-0.28	0.14	0.07	-0.07
Fc	0.26	0.37	0.11	0.10	0.06	0.04
Fd	0.15	0.38	0.23	0.23	0.21	-0.02
Fe	0.12	0.34	0.22	0.19	0.09	-0.10
Mean	*0.26*	*0.34*	*0.08*	*0.15*	*0.11*	*-0.03*
Lodalsbreen						
La	0.16	0.15	-0.01	0.06	0.06	0
Lb	0.52	0.67	0.15	0.03	0.05	0.02
Lc	0.07	0.11	0.04	0.30	0.08	-0.22
Ld	0.06	0.13	0.07	0.15	0.39	0.24
Mean	*0.20*	*0.27*	*0.06*	*0.14*	*0.15*	*0.01*
Leirdalen						
LEa	0.25	0.21	-0.04	0.11	0.18	0.07
LEb	0.27	0.15	-0.12	0.20	0.27	0.07
LEc	0.06	0.01	-0.05	0.22	0.25	0.03
LEd	0.01	0.05	0.04	0.23	0.30	0.07
LEe	0.03	0.12	0.09	0.19	0.23	0.04
Mean	*0.12*	*0.11*	*-0.02*	*0.19*	*0.25*	*0.06*
søre Illåbreen						
Sa	1.67	1.0	-0.67	0.09	0.09	0
Sb	0.94	1.0	0.06	0.07	0.05	-0.02
Mean	*1.31*	*1.0*	*-0.31*	*0.08*	*0.07*	*-0.01*
Visdalen						
Va	0.13	0.08	-0.05	0.24	0.31	0.07
Vb	0.29	0.14	-0.14	0.22	0.33	0.11
Vc	0.67	0.08	-0.59	0.14	0.11	-0.03
Vd	0.53	0.19	-0.34	0.13	0.14	0.01
Ve	0.55	0.03	-0.52	0.26	0.22	-0.04
Mean	*0.43*	*0.10*	*-0.33*	*0.20*	*0.22*	*0.02*
Heillstugubreen						
Ha	0.41	0.42	0.01	0.13	0.09	-0.04
Hb	0.60	0.47	-0.13	0.10	0.05	-0.05
Hc	0.55	0.03	-0.52	0.15	0.09	-0.06
Hd	0.23	0.41	0.18	0.07	0.10	0.03
Mean	*0.45*	*0.33*	*-0.12*	*0.11*	*0.08*	*-0.03*

Table 7.3. Slope geometry of ungullied drift slopes (UD) that represent the geometry of deglaciated drift slopes prior to modification by debris flow, gully floors (GF) that represent the geometry of paraglacially-modified valley-side slopes, and the difference between each (Δ) at the Norwegian field sites. For definitions of indices see text.

Profile	Index of concavity (c)			Index of linearity (l)		
	UD	GF	Δ	UD	GF	Δ
Western Red Hills						
Wa	0.73	0.60	-0.13	0.08	0.08	0
Wb	0.23	0.57	0.34	0.09	0.06	-0.03
Wc	0.35	0.65	0.30	0.11	0.13	0.02
Wd	0.76	0.76	0	0.08	0.06	-0.02
We	0.19	0.46	0.27	0.09	0.06	-0.03
Wf	0.36	0.39	0.03	0.10	0.08	-0.02
Mean	*0.44*	*0.57*	*0.14*	*0.09*	*0.08*	*-0.01*
Glen Docherty						
Da	0.15	0.03	-0.12	0.18	0.22	0.04
Db	0.08	0.09	0.01	0.21	0.28	0.07
Dc	0.17	0.12	-0.05	0.15	0.22	0.07
Dd	0.05	0.08	0.03	0.15	0.22	0.07
De	0.20	0.24	0.04	0.06	0.11	0.05
Mean	*0.13*	*0.11*	*-0.02*	*0.15*	*0.21*	*0.06*
Glen Einich						
Ea	0.06	0.14	0.08	0.18	0.19	0.01
Eb	0.09	0.12	0.03	0.12	0.12	0
Ec	0.09	0.16	0.07	0.13	0.13	0
Ed	0.05	0.06	0.01	0.16	0.13	-0.03
Ee	0.13	0.12	-0.01	0.16	0.15	-0.01
Mean	*0.08*	*0.12*	*0.04*	*0.15*	*0.14*	*-0.01*
Pass of Drumochter						
PDa	0.05	0.01	-0.04	0.23	0.23	0
PDb	0.02	0.01	-0.01	0.20	0.24	0.04
PDc	0.14	0.15	0.01	0.18	0.23	0.05
PDd	0.12	0.06	-0.06	0.30	0.31	0.01
Mean	*0.08*	*0.06*	*-0.03*	*0.23*	*0.25*	*0.02*

Table 7.4. Slope geometry of ungullied drift slopes (UD) that represent the geometry of deglaciated drift slopes prior to modification by debris flow, gully floors (GF) that represent the geometry of paraglacially-modified valley-side slopes, and the difference between each (Δ) at the Scottish field sites. For definitions of indices see text.

On 24 sets of profiles there has been a decrease in slope concavity (*i.e.* a shift toward convexity, with the index c increasing by 0.01-0.34), whilst on 20 sets of slope profiles, gully floor profiles are more concave than the corresponding ungullied slope profiles, as identified by decreases in c of 0.01-0.67. A decrease

in concavity (i.e. an increase in c) would be expected if incision at the gully head is accompanied with partial infilling of the basal concavity by resedimented drift (*cf.* Ballantyne and Benn, 1994), or if bedrock steps were exposed, increasing the irregularity (area A, Figure 7.1) of the gully floor profile. In contrast, a slope profile would be expected to become more concave (less convex) if the initial slope form contained marked convexities, or if there was limited incision at the gully head and marked downslope extension of slope foot debris, as is evident when debris slopes are modified by repeated by snow avalanche activity (e.g. Luckman, 1971; 1972; 1977; 1978). Such a pattern emerges in a plot of the difference in the concavity index c between ungullied drift slopes and corresponding gully floors against the ungullied drift c values (Figure 7.12b). This reveals a negative relationship between the two variables, significant at $p < 0.001$. This relationship shows that most of the slopes which were initially relatively concave (ungullied drift $c < 0.22$) have become less concave through 'cut' at the top of the slope and 'fill' at the base. Exceptions to this overall trend include individual slope profiles at Leirdalen (LEc, Figure 7.4), Visdalen (Va, Figure 7.6), Glen Docherty (e.g. Da, Figure 7.9) and the Pass of Drumochter (e.g. PDa, Figure 7.11). Each of these exceptions exhibits shallow failure at the gully head, which accounts for an overall increase in concavity. Conversely, slopes which were initially less concave (with ungullied drift c-values > 0.22) tend to exhibit an increase in overall concavity. In many cases, such profiles also exhibit concave failure scars at gully heads, resulting in an increase in concavity (for example, Fb, Figure 7.2; Ve, Figure 7.6). There are exceptions to this trend, however, notably profile Lb (Figure 7.3) and profile Sb (Figure 7.5), which have become less concave despite having relatively high ungullied slope c-values. Whilst the ungullied Lb profile has a high c-value (0.52), the presence of glacier ice at the slope foot has constrained run-out and cone accumulation, and inhibited any increase in concavity. Similarly, the ungullied part of profile Sb adjacent to the gully contains no basal concavity and has a very high c value of 1.67. Despite

these exceptions, the general departure from concavity of the concave slopes, and the increasing concavity exhibited by the less concave slopes show an overall tendency towards a uniform concavity value of *c.* 0.22, which may approximate an 'equilibrium' configuration for paraglacially-modified drift slopes.

No systematic pattern emerged from analysis of the linearity index (*l*) data in Tables 7.3 and 7.4. Whilst 16 paired profiles exhibit an increase in overall linearity, 23 yielded a decrease in overall linearity, and no change was evident in the remainder. Moreover, no relationship was detected between change in linearity between paired ungullied and gullied profiles and initial linearity. It appears that though incision and redeposition of sediment in some cases produce a more linear slope profile, in others the excavation of an increasingly concave gully profile, or a more irregular profile arising from exposure of bedrock steps in the gully floor, results in a reduction in the overall linearity of the profile.

7.3.3 Modification of drift slope profiles: summary

The above analyses show that paraglacial modification of valley-side drift accumulations generally tends to produce slopes with gentler overall gradients, reflecting removal of sediment from upslope and its redeposition on cone and fan surfaces at the slope foot (*cf.* Ballantyne and Benn, 1994; 1996). Detailed survey and analysis have shown, however, that the nature of paraglacial drift-slope adjustment depends considerably on initial slope gradient and configuration. In most cases where unmodified drift slopes have an upper rectilinear slope steeper than *c.* 28° and exhibit pronounced concavity, gully incision has lowered the upper slope gradient and overall concavity has been reduced by redeposition of sediment downslope. Conversely, on less concave drift slopes where the gradient of the initial upper slope lay below *c.* 28°, paraglacial gullying has often steepened the upper rectilinear slope and increased the overall concavity. Although some

exceptions were observed, paraglacial slope profile adjustment therefore appears to be characterised by a convergence of slope form towards an 'equilibrium profile' characterised by an upper rectilinear slope gradient of *c.* 28° and an index of concavity of *c.* 0.22, irrespective of initial slope configuration. This limiting gradient may reflect the threshold for sediment removal by debris flows.

7.4 Gully formation.

Several processes of sediment erosion and transportation operate on the gullied slopes at the field sites investigated, including translational failure of drift, debris flow activity and surface wash on gully floors, collapse of gully sidewalls and debris fall from gully sides. Undoubtedly, a certain amount of erosion also occurs on inter-gully slopes as a result of surface wash and rill erosion (Beaty, 1959). At the most active field sites, a combination of mass movement and surface wash processes has produced sharp gully divides (Figure 7.13) rather than the rounded 'belt of no erosion' divides characteristic of badlands dominated by rainsplash and surface wash processes (Horton, 1945). Whilst it is likely that these processes collectively produce seasonally-distinctive changes in gully slope form (e.g. Schumm, 1956; 1964; Harvey, 1987, 1992), attention is focused here on the nature of gully development over longer timescales, and concentrates on progressive changes in gully size and shape with age. Assessment of gully development was confined to the Norwegian field sites where the temporal development of gullies is better constrained than at the Scottish sites. Profiles describing the cross-sectional form of all surveyed gullies at the Norwegian sites are shown in Figures 7.14 to 7.19. As outlined in section 4.2, these profiles were measured at several points along surveyed gully axes, and correspond to the survey locations marked in Figures 7.2 to 7.7. Results of analysis of these profiles are outlined in Figures 7.20 and 7.21 and are evaluated below by classifying them according to the three time periods described in section 7.2. Examples of gullies

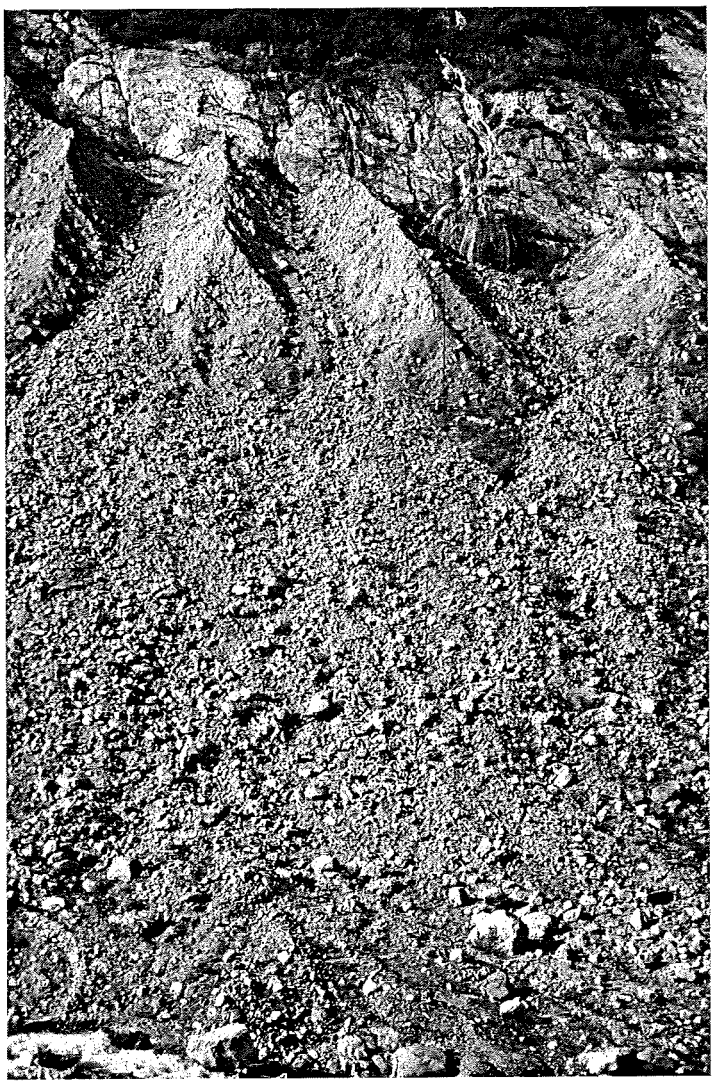


Figure 7.13. Paraglacial gully systems in Fåbergstølsdalen, including gully profile Fd at the right-hand side of the frame. Notice the sharp gully divides, characteristic of incision by highly-active mass movement and surface wash processes.

of these different age categories are shown in Figure 7.22 and represent different stages in the process of paraglacial slope adjustment.

7.4.1 Modification of gully size with time

By comparing the gully profiles illustrated in Figures 7.14 to 7.19, it can be seen that the oldest gullies are generally much larger than the younger ones. For example, the older (T_2) gullies at Fåbergstølsbreen and Lodalsbreen (Fd, Fe

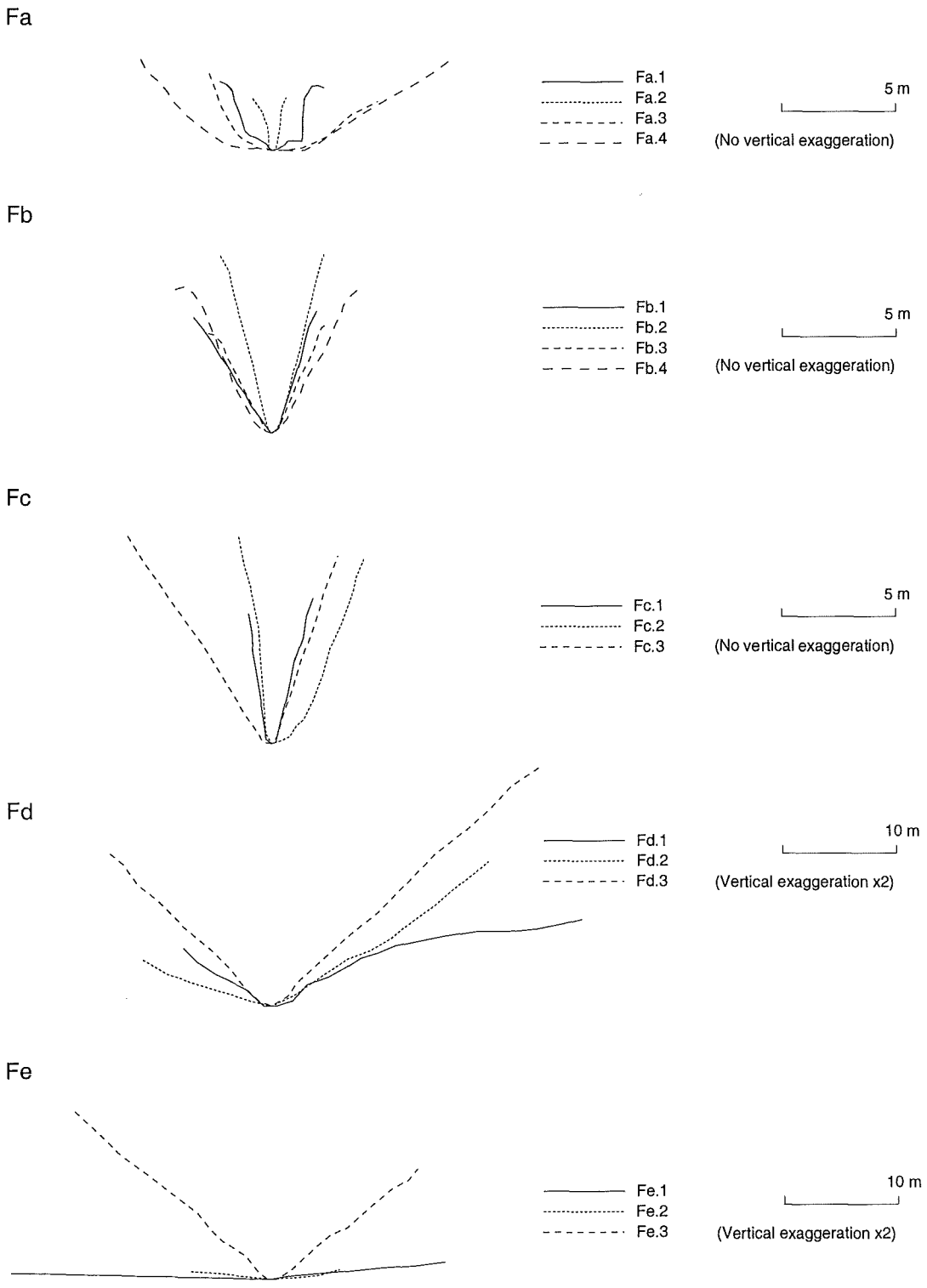


Figure 7.14. Gully cross-section profiles for five gullies surveyed on recently-deglaciated drift in the Fåbergstølsbre foreland, Norway.

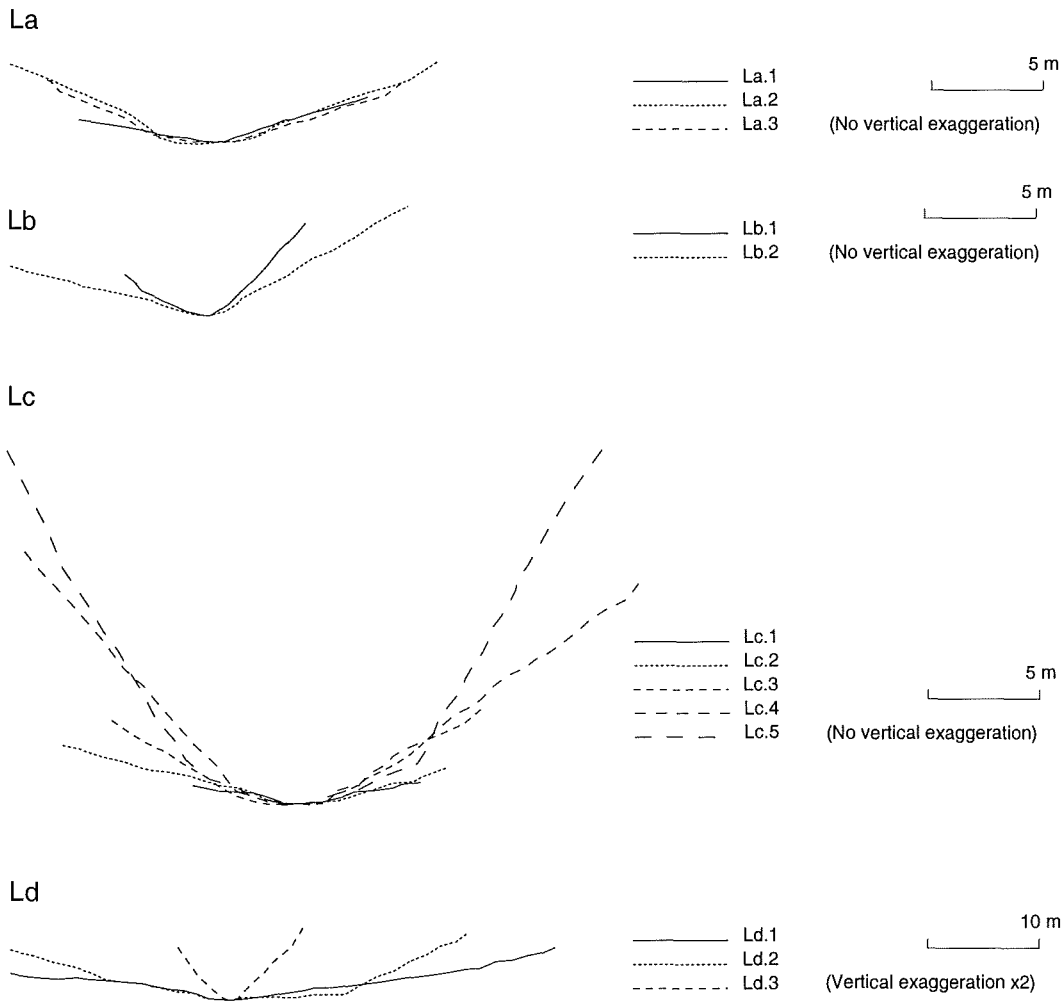


Figure 7.15. Gully cross-section profiles for four gullies surveyed on recently-deglaciated drift in the Lodalsbre foreland, Norway.

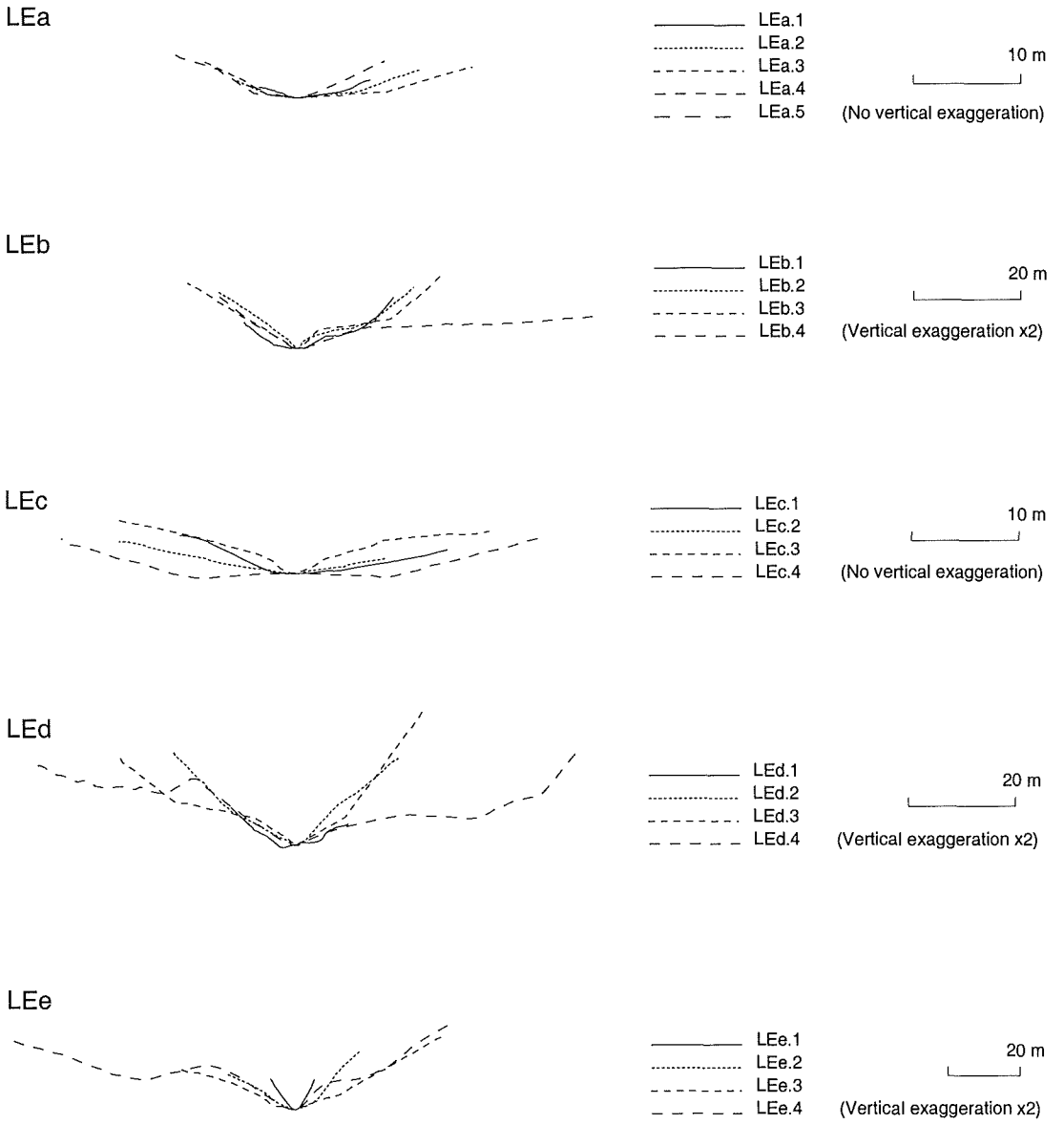


Figure 7.16. Gully cross-section profiles for five gullies surveyed on drift slopes in Leirdalen, Norway.

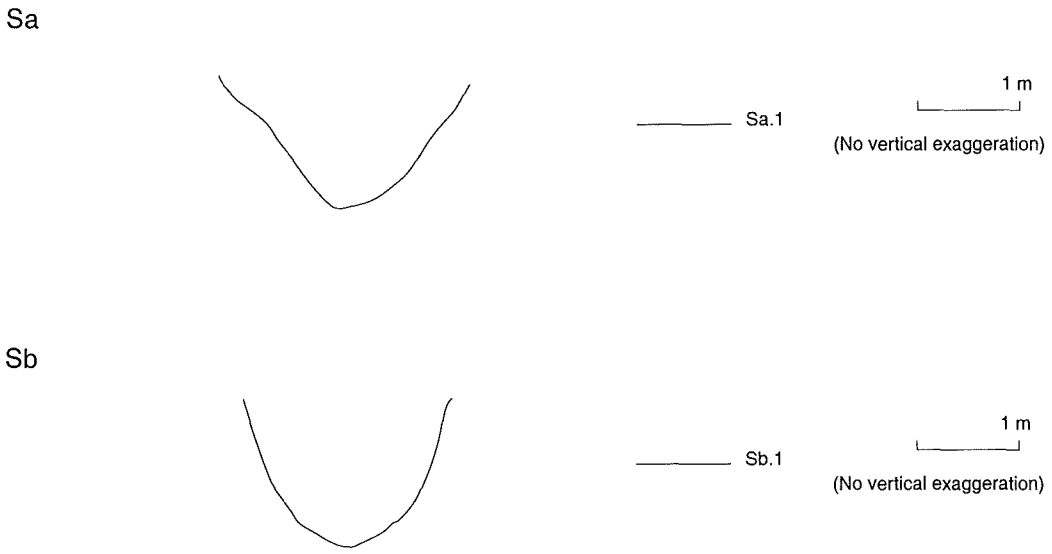


Figure 7.17. Gully cross-section profiles for two gullies surveyed on recently-deglaciated drift slopes in the søre Illåbre foreland, Norway.

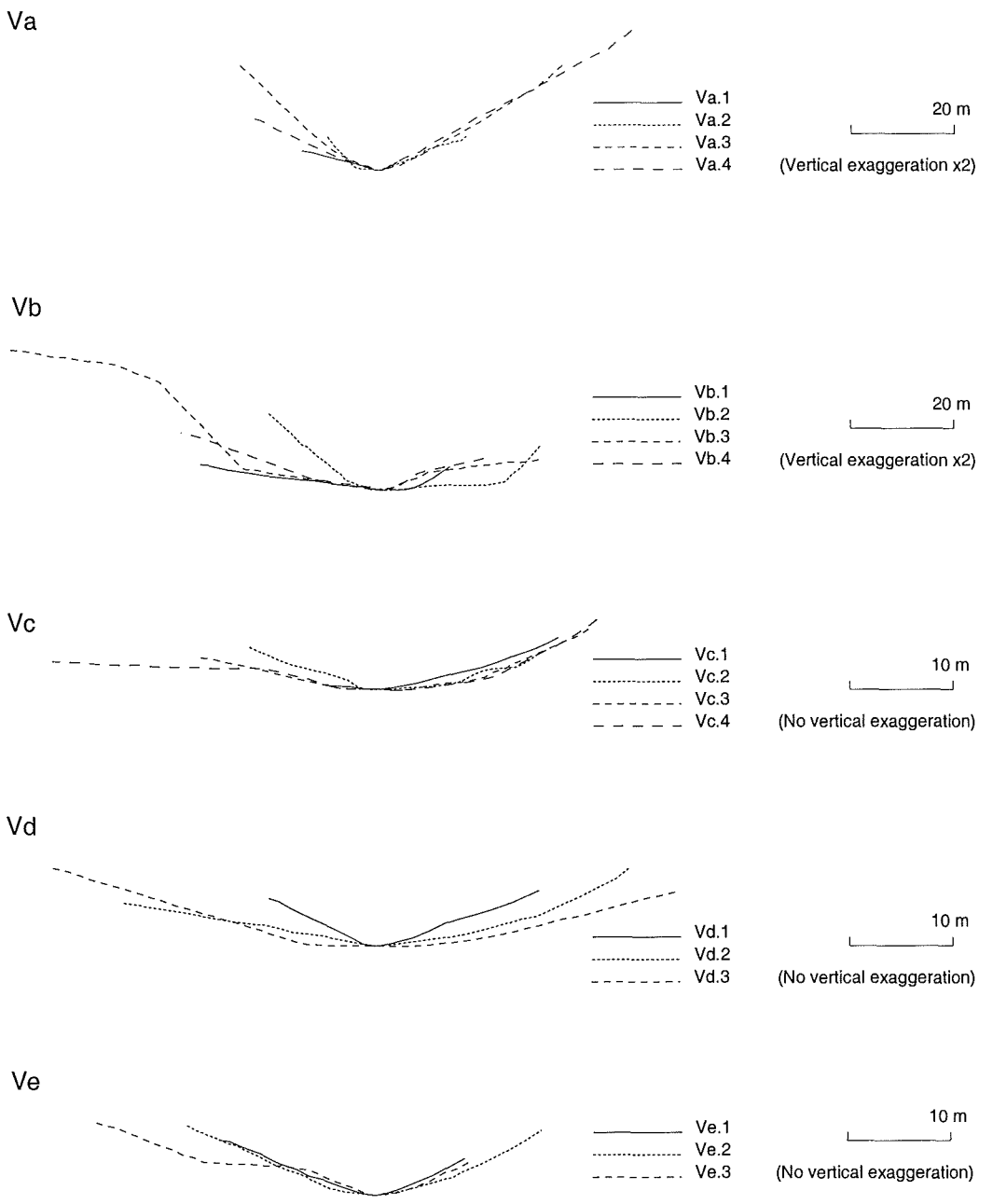


Figure 7.18. Gully cross-section profiles for five gullies surveyed on drift slopes in Visdalen, Norway.

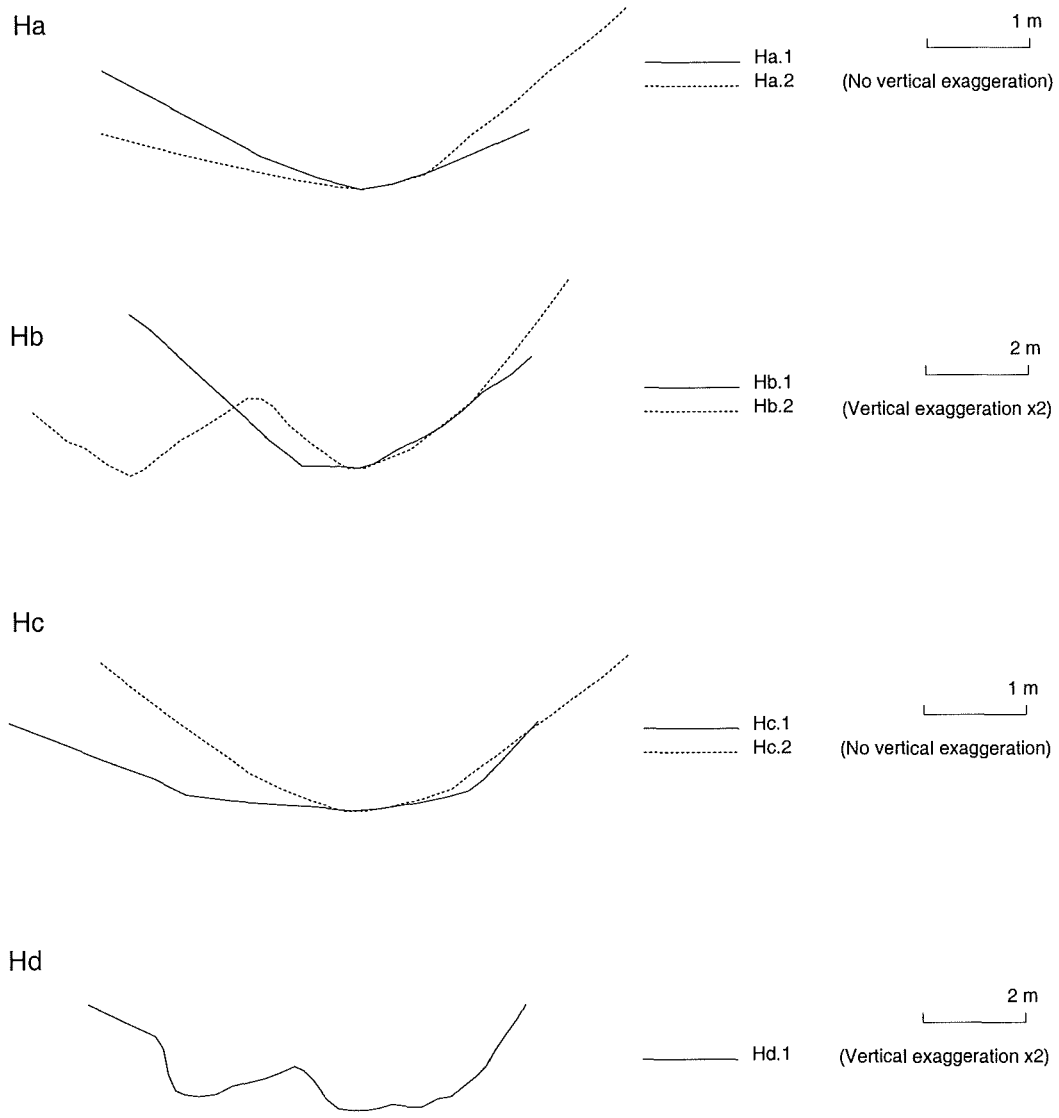


Figure 7.19. Gully cross-section profiles for four gullies surveyed on recently-deglaciated drift slopes in the Heillstugubre foreland, Norway.

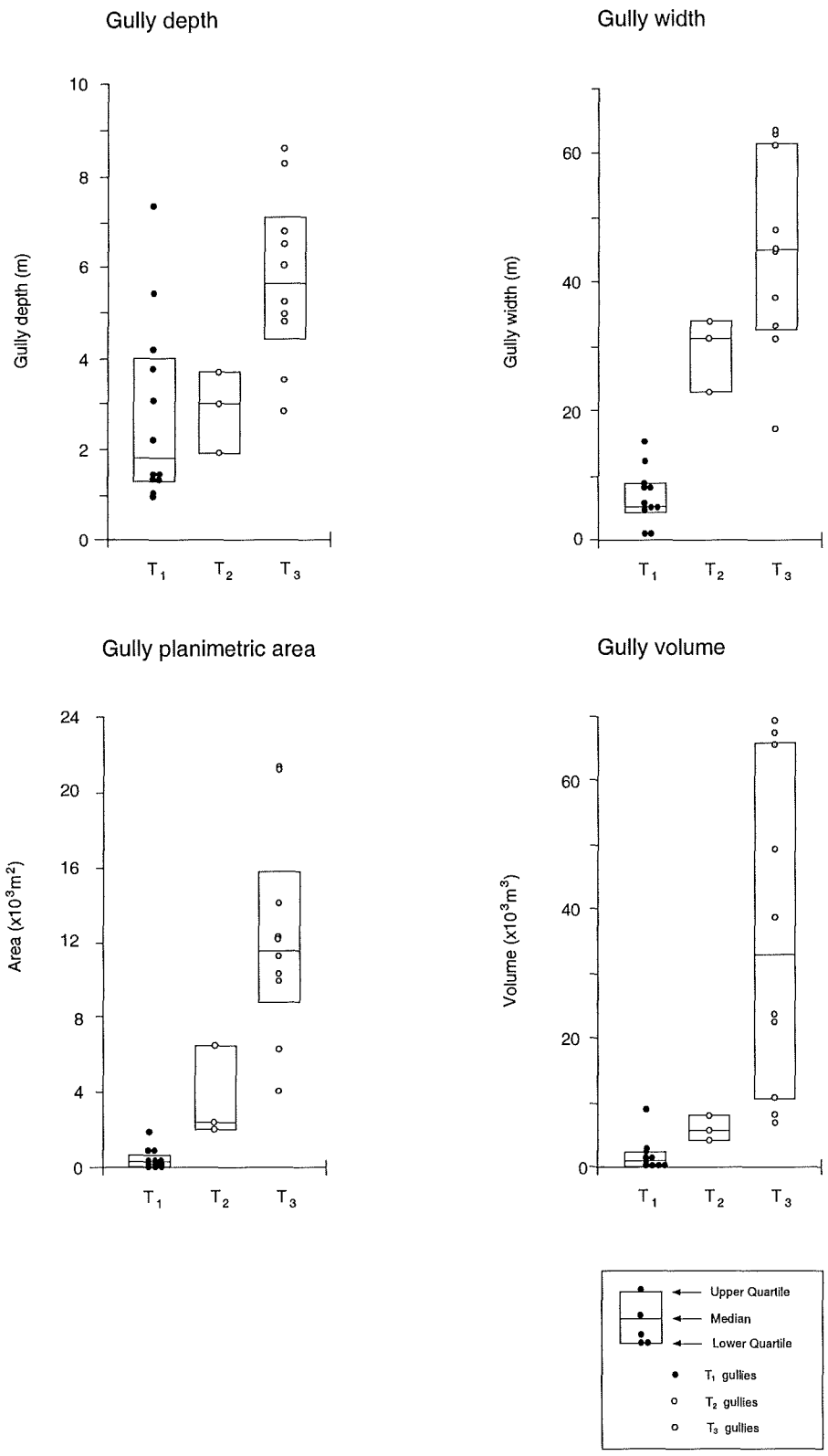


Figure 7.20. Dispersion diagrams illustrating changes in gully size with gully age. T_1 , T_2 and T_3 represent groups of gullies of increasing age at the Norwegian field sites.

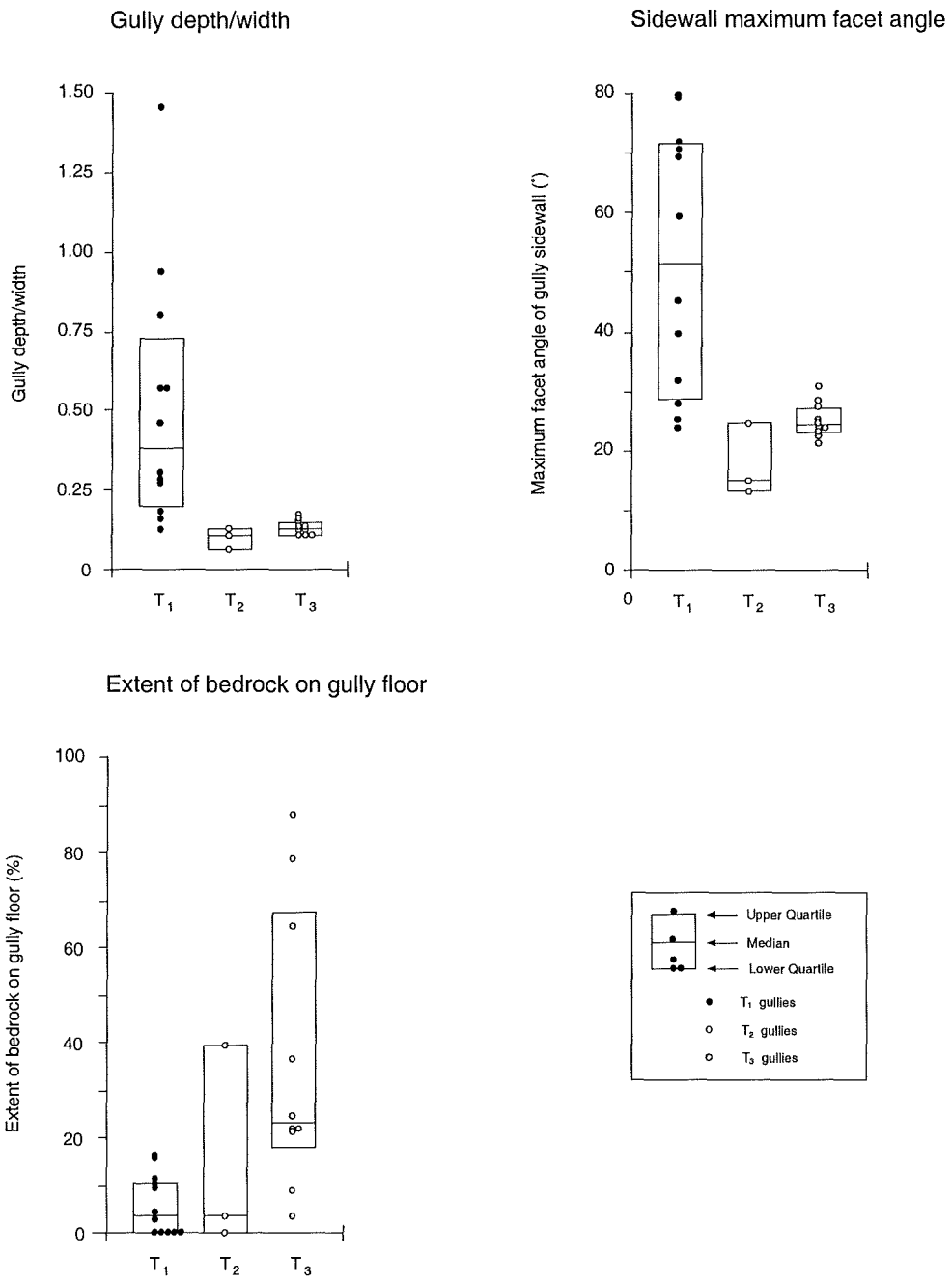


Figure 7.21. Dispersion diagrams illustrating changes in gully shape and extent of bedrock exposed on the gully floor with gully age. T_1, T_2 and T_3 represent groups of gullies of increasing age at the Norwegian field sites.

a

b

c

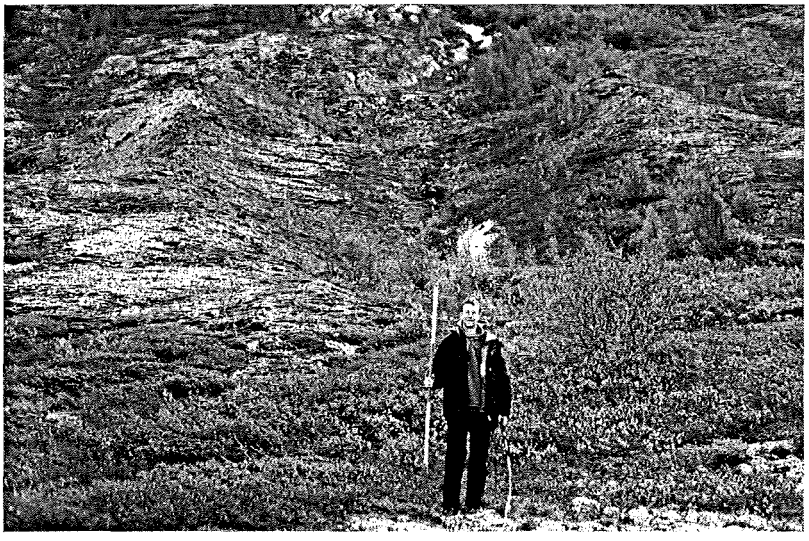


Figure 7.22. Representative examples of (a) T_1 gullies incising recently-exposed drift well within the 'Little Ice Age' glacier limits at Lodalsbreen; (b) T_2 gullies on older drift deposited by the 'Little Ice Age' advance of Fåbergstølsbreen; and (c) mature T_3 gullies incising valley-side drift of Preboreal-age in Leirdalen.

and Ld in Figures 7.14 and 7.15) are considerably wider than the youngest (T_1) gullies surveyed further upvalley at these two sites (Fa-Fc and La-Lc). Similarly, most of the oldest (T_3) gullies measured at the Jotunheim field sites (e.g. LEb, LEd, LEe, Va and Vd, Figures 7.16 and 7.18) are deeper and wider than the younger gullies inside 'Little Ice Age' glacier limits both in Jotunheimen (Sa-Sb and Ha-Hd, Figures 7.17 and 7.19) and at the Jostedalsbre field sites (e.g. Fa, Fb, Fd, Fe and La-Ld, Figures 7.14 and 7.15). These trends suggest systematic changes in gully size with age, and are considered more fully below.

Average values of gully depth and width (weighted according to the proportion of slope represented by each cross-section), planimetric area and gully volume are summarised in Figure 7.20 for each of the three gully age categories. As suggested by inspection of Figures 7.14 to 7.19, all four parameters tend to increase with increasing gully age. In general, the oldest (T_3) gullies are deeper, wider, cover a greater area and have larger volumes than those in the T_2 category, which are in turn deeper, broader and more extensive than the youngest (T_1) gullies. The differences between the T_1 gullies and T_3 gullies are significant at $p < 0.01$ for all four parameters: median depth increases from 1.79 m to 5.61 m; median width from 5.5 m to 44.9 m; median area from 245 m^2 to 11,623 m^2, and median volume from 910 m^3 to 31,330 m^3. Whilst both gully depth and width increase with age, the increase in width is very much greater, and accounts for most of the increase in gully volume. This increase in gully size by widening is well illustrated in Figure 7.22. Other workers describing the development of gully systems have observed upslope extension of gullies through time by gully head retreat (e.g. Brice, 1966; Okuda *et al.*, 1980; Harvey, 1987; 1992; Owen, 1991). Headward retreat was not assessed here, but probably occurs where topographic constraints permit. Indeed, on the oldest valley-side drift in Fåbergstølsdalen, erosion has completely removed the outermost 'Little Ice Age'

lateral moraine (Figure 4.12) and extended the gully system upslope into older drift deposits (section 6.4.1; Ballantyne and Benn, 1994).

7.4.2 Modification of gully shape with time

Temporal trends in gully shape were assessed in terms of depth/width ratio, maximum sidewall facet angle and the extent of bedrock exposed on gully floors (Figure 7.21). The first two parameters exhibit rapid changes through time. The median depth/width ratio for T_1 gullies is 0.38, compared to a median value of 0.11 for the T_2 gullies and 0.13 for the T_3 gullies. Similarly, the median value of maximum sidewall facet angles is 51.9° for the T_1 gullies, 14.7° for the T_2 gullies and 24.2° for the T_3 gullies. For both parameters the T_2 and T_3 values are statistically indistinguishable, but both are significantly lower than the T_1 values at $p < 0.05$ (Mann-Whitney two-sample test). Both parameters therefore indicate that the youngest (T_1) gullies tend to be narrow relative to their depth and steep-sided (Figure 7.22a), but that within a few decades of initial incision both depth/width ratios and maximum sidewall facet angles have declined significantly, and thereafter exhibit limited further change. This pattern implies that following a period of initial rapid incision, gullies undergo progressive widening accompanied by a decline in the gradient of sidewall slopes to < *c.* 25°, after which further widening appears to be accomplished partly by parallel retreat of sidewalls.

These trends in gully development can be seen particularly clearly at Fåbergstølsbreen (Figure 7.14). At this site, the youngest gullies are characterised by steep sidewalls (> 65°) and a relatively narrow cross-sectional form (e.g. Fb and Fc, Figure 7.14). In contrast, older gullies such as Fd and Fe further down-valley have a much broader shape and possess much gentler (< 25°) gully sides. Similarly, of the oldest (T_3) gullies illustrated in Figures 7.16 and 7.18, gullies

LEb, LEc and Vb are especially broad relative to their depth and have sidewalls whose maximum gradients lie between 20° and 25°.

7.4.3 Gully stabilisation

A further point of interest concerning gully development involves slope stabilisation. Whilst focusing on much smaller gullies (generally < *c.* 50 m length) than those described above, Harvey (1992) suggested that gully stabilisation might be controlled by vegetation encroachment. At the Norwegian field sites the most extensive vegetation cover is on the cones and floors of the oldest gullies (Figures 7.2 to 7.7), suggesting that vegetation colonisation may be related to gully age and stabilisation. However, in the case of the large gully systems surveyed in Norway, vegetation colonisation might be expected to be more a response to stabilisation of the ground surface than a causal mechanism. A more likely control on gully stabilisation is depletion of the upslope supply of sediment. As the data in Figure 7.20 reveal, the gullies which have been potentially active for the longest time (the T_3 gullies) have a much higher percentage of bedrock exposed on the gully floor (median = 23.2% of total gully length) than the younger (T_1) gully systems (median = 3.7%), particularly in the gully heads. These differences in bedrock exposure are significant at $p < 0.005$ (Mann-Whitney two-sample test). This finding supports observations made in section 4.6.2, where exhaustion of sediment supply was recognised as a critical factor in terminating paraglacial reworking of drift slopes. Moreover, progressive exposure of bedrock on gully floors explains the reduction of gully deepening relative to widening illustrated above in section 7.4.2, in that deepening of individual gullies is rapidly limited by exposure of bedrock on gully floors.

7.4.4 Gully formation: summary

The grouping of gully shape data into different age categories has allowed trends in gully development to be identified (Figures 7.20 and 7.21). These trends are summarised schematically for a single gully in Figure 7.23. Gullies initially modify valley-side drift by incising vertically to form deep, steep-sided canyons with high depth/width ratios, but very rapidly broaden out and become much larger conduits of sediment transfer. If initiation of debris flows is limited by a minimum gully floor gradient (section 4.5.1), debris flows will continue to remove debris accumulating in gullies as a result of sidewall collapse, but further gully deepening will be limited (*cf.* Ballantyne and Benn, 1994, 1996). The rate of vertical incision is ultimately limited by exposure of bedrock in the gully floor. Subsequent gully development appears to take the form of gully widening, initially by gully-wall slope decline and later by gully-wall slope retreat. The final form of paraglacial gully systems generally comprises an upper bedrock-floored source area, a mid-slope area of broad gullies whose sidewalls rest at relatively gentle (< *c.* 25°) gradients, and a lower slope zone where gullies discharge onto the vegetated surfaces of coalescing debris cones and fans (*cf.*

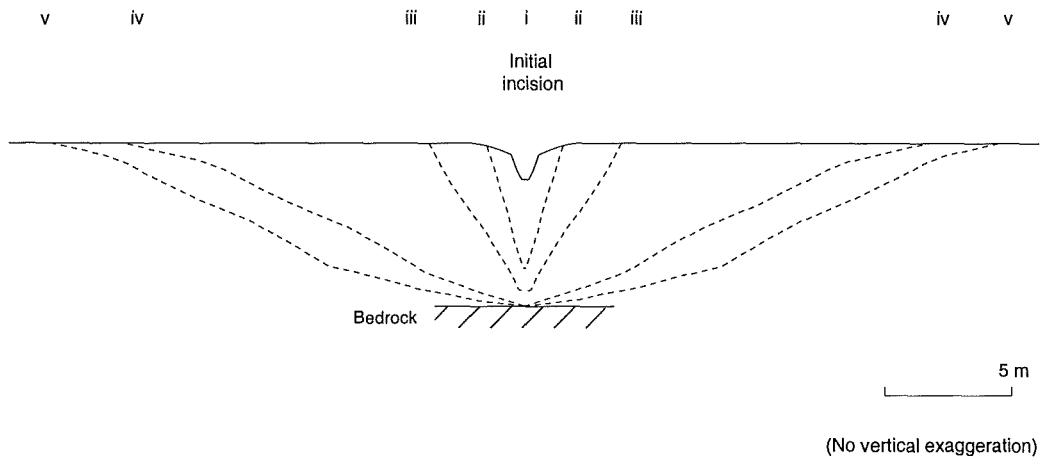


Figure 7.23. Schematic representation of paraglacial gully development, as inferred from measurement of paraglacial gully systems in western Norway: (i) initial incision; (ii) gully deepening; (iii) - (iv) gully widening and sidewall decline; (iv) - (v) sidewall retreat and stabilisation.

Ballantyne and Benn, 1994). Given that the T_2 gullies in Jostedalen are as little as 25 years older than the youngest (T_1) gullies in the same area (section 4.6.1), the progress of gully development from steep-sided canyons to broad, open gullies at these sites is clearly rapid, consistent with the accumulation and stabilisation of paraglacial debris cones on recently-deglaciated terrain within a decadal timescale (Ballantyne, 1995a; Harrison and Winchester, 1997). Rates of ground surface lowering resulting from paraglacial gully systems at the investigated field sites are considered in section 7.5 below.

7.5 Rates of slope adjustment.

Discussion of the rates of geomorphological processes requires some caution, in that the literature on the topic is inevitably biased towards sites of pronounced activity and the effects of high-magnitude events. Nevertheless, it is evident that debris flow represents one of the most effective agents of sediment transfer associated with steep debris slopes in mid-latitude and high-latitude mountain environments (e.g. Rapp, 1960; Gardner, 1979; André, 1986; Ballantyne and Harris, 1994). At the sites investigated in Norway and Scotland, the data on gully dimensions (Tables 4.1, 4.2 and 5.1) can be employed to calculate average ground surface lowering since ice retreat. Where the timing of paraglacial activity is well-constrained, long-term erosion rates have been calculated. These are presented below, together with a comparative case study of short-term slope adjustment within the Fåbergstølsbre glacier foreland.

7.5.1 Long-term rates

Average ground surface lowering within gully systems at the study sites was calculated for individual gullies by dividing total gully volume by gully surface area (Table 7.5). The results show a fairly wide range of individual

Profile	Average surface lowering since gully initiation (m)
Fåbergstølsbreen	
Fa	2.4
Fb	3.6
Fc	5.1
Fd	3.5
Fe	2.0
**	3.1
**	2.5
**	4.7
Mean	*4.1*
Lodalsbreen	
La	1.3
Lb	3.4
Lc	4.8
Ld	0.8
Mean	*2.6*
Leirdalen	
LEa	1.2
LEb	2.0
LEc	1.1
LEd	4.9
LEe	2.4
Mean	*2.3*
søre Illåbreen	
Sa	0.6
Sb	0.8
Mean	*0.7*
Visdalen	
Va	6.1
Vb	3.1
Vc	2.2
Vd	3.2
Ve	2.1
Mean	*3.3*
Heillstugubreen	
Ha	0.4
Hb	0.7
Hc	0.6
Hd	0.6
Mean	*0.6*
Western Red Hills	
Wa	3.3
Wb	0.6
Wc	2.0
Wd	5.1
We	0.8
Wf	4.0
Mean	*2.4*
Glen Docherty	
Da	4.3
Db	2.1
Dc	4.9
Dd	3.6
De	3.6
Mean	*3.7*
Glen Einich	
Ea	8.4
Eb	2.2
Ec	1.7
Ed	1.0
Ee	0.7
Mean	*2.8*
Pass of Drumochter	
PDa	2.8
PDb	2.2
PDc	2.3
PDd	2.3
Mean	*2.4*

Table 7.5. Amounts of ground surface lowering of steep drift slopes at field sites in Norway and Scotland. ** data from Ballantyne and Benn (1994).

values, but with the exception of the Heillstugubre and søre Illåbre sites, where gullies are generally small and immature, the mean site values fall between 2.3 m (Leirdalen) and 4.1 m (Fåbergstølsdalen) of surface lowering (overall median value = 2.45 m), irrespective of age. The gullies on the forelands of Heillstugubreen and (especially) søre Illåbreen may be regarded as atypical and unrepresentative, having failed to develop to the extent of those elsewhere due to unfavourable constraints (section 4.5). On recently-deglaciated terrain around Fåbergstølsbreen and Lodalsbreen, however, ground surface lowering has already occurred to depths similar to those of much older gully systems, such as those which were initially incised several millennia ago in Leirdalen. This suggests that, under propitious circumstances, many paraglacial gully systems approach maturity in less than 50 years (section 4.6.1), implying that though paraglacial slope modification may continue (intermittently) for millennia, the most active period of paraglacial sediment reworking often occurs within decades of deglaciation. This temporal pattern of drift slope response to deglaciation appears similar to the exponential decline in sediment yield inferred for fluvial environments after ice retreat (Church and Ryder, 1972; Church and Slaymaker, 1989; Harbor and Warburton, 1993).

For sites where the approximate age of recent deglaciation is known, corresponding minimum rates of ground surface lowering have been calculated by dividing total ground surface lowering at gully sites by maximum time elapsed since deglaciation (Table 7.6). As might be expected, the implied surface lowering rates at the two Jostedalsbre sites are significantly higher (median value 80 mm yr^{-1}) than those at Heillstugubreen and søre Illåbreen (median value 6.9 mm yr^{-1}), reflecting the limiting constraints on paraglacial activity at the latter. Furthermore, the surface lowering rates associated with the most recent (T_1) gullies at Fåbergstølsbreen and Lodalsbreen are much greater (median = 115 mm yr^{-1}) than those for the T_2 gullies at the same sites (median = 38 mm yr^{-1}),

Profile	Minimum rate of ground surface lowering since gully initiation (mm yr^{-1})	Profile	Minimum rate of ground surface lowering since gully initiation (mm yr^{-1})
Fåbergstølsbreen		*Heillstugubreen*	
Fa	80	Ha	5.5
Fb	119	Hb	8.8
Fc	169	Hc	8.3
Fd	65	Hd	8.2
Fd	38		
		Mean	*7.7*
Mean	*94*	Overall site mean	*0.41*
Overall site mean	*42.3*		
Lodalsbreen		*søre Illåbreen*	
La	67	Sa	2.5
Lb	169	Sb	3.0
Lc	111		
Ld	19	Mean	*2.8*
		Overall site mean	*0.02*
Mean	*92*		
Overall site mean	*73.6*		

Table 7.6 Rates of ground surface lowering of steep drift slopes on recently-deglaciated glacier forelands in Norway. Individual figures and mean figures refer to within-gully ground surface lowering. Overall site mean reflects the average ground surface lowering for all (gullied and ungullied) portions of the slope.

confirming that the rate of paraglacial erosion of drift is greatest immediately after deglaciation. Possible causes of the declining rate of paraglacial gully erosion through time include progressive stabilisation of older gullies as a result of declining sediment supply (section 7.4.3), lowering of gully-floor and/or gully wall gradients (section 7.3.1) or availability of meltwater (e.g. Ryder, 1971a, 1971b; Fitzsimons, 1996).

To give an impression of how these rates compare with rates of surface lowering attributable to debris flow activity in other upland environments, comparative data selected from published sources are presented in Table 7.7. The range of minimum ground surface lowering values within gullies on the

Location	Rate (mm yr^{-1})	Cover	Source
(a) Surface lowering			
Fåbergstølsdalen, Norway	50-100	Glacigenic	Ballantyne and Benn, 1994
Bergsetdalen, Norway	37-94	Glacigenic	Ballantyne, 1995a
Mount Rainier, USA	20-40	Glacigenic	Walder and Dreidger, 1994
Howgill Fells, England	55-210	Glacigenic	Harvey, 1987
Tarfala, Lappland	5	Periglacial	Rapp, 1975
Longyear Valley, Spitsbergen	1	Periglacial	Rapp, 1975
Mangawhara, New Zealand	10-80	Volcanic	Selby, 1976
(b) Debris accumulation			
San Rafael Glacier, Patagonia	300-400	Glacigenic	Harrison and Winchester, 1997
Bergsetdalen, Norway	8-44	Glacigenic	Ballantyne, 1995a

Table 7.7 Rates of (a) ground surface lowering and (b) debris accumulation attributable to debris flow activity on steep slopes in various upland environments. Data have been adjusted where necessary for convenience of representation. Rates of ground surface lowering refer to gullied sites and are not averaged over the whole (gullied and ungullied) slope.

Fåbergstølsbre (38-169 mm yr^{-1}; mean 94 mm yr^{-1}) and Lodalsbre (19-169 mm yr^{-1}; mean 92 mm yr^{-1}) forelands are both mutually consistent and overlap previous estimates of erosion rates in gullies at sites around Jostedalsbre based on more limited surveys of gully dimensions and volumes of debris cones. Ballantyne and Benn (1994) found that ground surface lowering by paraglacial debris flows since AD 1943 in Fåbergstølsdalen was equivalent to a minimum erosion rate of 50-100 mm yr^{-1}, well within the above ranges, though they also estimated that the size of the very largest gully represented an erosion rate of $\geq$ 200 mm yr^{-1}. Similarly, surveys of the dimension of debris cones in Bergsetdalen by Ballantyne (1995a) imply average rates of sediment accumulation of 8-44 mm

yr^{-1}, with an inferred average rate of drift removal of 37-94 mm yr^{-1} at one site. These overlapping ranges of rates of paraglacial gully erosion at the three locations suggest that they may be regarded as being reasonably representative for recently-deglaciated steep drift-covered slopes in western Norway. These rates are also broadly similar to those calculated by Harvey (1987) for ground surface lowering associated with recent gully erosion in glacial deposits in northern England, but are markedly lower than Harrison and Winchester's (1997) surprisingly high estimates for paraglacial sediment accumulation on debris cones in recently-deglaciated environments in Patagonia. Although similar to Selby's (1976) estimated rate of gully erosion in unconsolidated volcanic deposits in New Zealand, the rates of paraglacial debris flow erosion and resedimentation generally exceed those attributable to debris flow activity outside a paraglacial setting.

By collating over 400 published rates of slope processes worldwide, Young and Saunders (1986) presented typical rates of ground loss attributable to surface wash (0.002-0.2 mm yr^{-1}), solution (0.002-0.1 mm yr^{-1}) and landsliding (0.1-10 mm yr^{-1}), and cited 1-5 mm yr^{-1} as representative of rates of overall surface lowering in steep, glacial environments. Comparison of these erosion rates with those calculated for paraglacial gully development at Fåbergstølsbreen and Lodalsbreen highlights the extreme rapidity with which paraglacial erosion of recently-deglaciated terrain occurs, even though this is focused in gully systems and thus not necessarily representative of general rates of ground surface lowering, except on terrain where gully systems coalesce with no intervening ungullied ground. Overall rates of ground surface lowering for both gullied and ungullied slopes at Fåbergstølsbreen and Lodalsbreen were estimated by multiplying the mean ground surface lowering rates within gullies by the proportion of the overall slope occupied by gullies within the uppermost kilometre of each foreland (Table 7.6). This calculation yielded overall values of *c.* 42 mm

yr^{-1} at Fåbergstølsbreen and *c.* 74 mm yr^{-1} at Lodalsbreen, roughly an order of magnitude higher than the general rate for steep glaciated environments estimated by Young and Saunders (1986), and again emphasising the extreme rapidity of drift slope erosion and sediment transfer associated with paraglacial activity shortly after exposure of steep drift slopes from under a cover of glacier ice.

In sum, the new and published data collectively indicate that a very wide range of rates may be associated with paraglacial gully development in recently deglaciated terrain, ranging from the relatively low minimum erosion rates (2.5-8.8 mm yr^{-1}) calculated for sites in Jotunheimen (Heillstugubreen and søre Illåbreen; Table 7.6) to the extremely rapid rates (300-400 mm yr^{-1}) of resedimentation suggested by Harrison and Winchester (1997) for debris cones on the San Rafael Glacier foreland in Patagonia. However, even the relatively low erosion rates calculated for the Jotunheim sites greatly exceed 'normal' erosion rates in many other environments (Saunders and Young, 1983; Young and Saunders, 1986), often by several orders of magnitude.

7.5.2 Short-term rates: Fåbergstølsdalen

Short-term rates of net surface lowering and redeposition were estimated from the erosion pin measurements made over a single year (1996-7) in upper Fåbergstølsdalen. Whilst the limitations of data collected over such a brief time period must be emphasised, these values do help to provide an indication of the magnitude of slope adjustment and the range of internal variation at a single site. The location of the measurement sites is shown on the site map of Fåbergstølsbreen (Figure 4.3). Rates of net erosion and sediment accumulation within two gullies cut into recently-exposed drift above the snout of Fåbergstølsbreen, and on cone surfaces throughout the Fåbergstølsbre foreland, are outlined in Table 7.8.

Pin site	Pin no.	Location	Net erosion (-) or accumulation (+) 1996-7 (mm yr^{-1})
(a) Gullies			
1	1	Crest of gully sidewall	-45
	2	Gully sidewall	-10
	3	Foot of gully sidewall	+5
	4	Gully floor	+7
	5	Gully floor	+92
	6	Gully sidewall	-131
	7	Gully sidewall	-62
2	1	Crest of gully sidewall	-35
	2	Foot of gully sidewall	+164
	3	Gully floor	+232
(b) Debris cones			
3	1	Debris cone surface	+8
	2	Debris cone surface	-36
	3	Debris cone surface	+3
	4	Debris cone surface	-4
4	1	Debris cone surface	+192
	2	Debris cone surface	+279
	3	Debris cone surface	-506
5	1	Slope-foot wash deposits	+3
	3	Slope-foot wash deposits	-3
	5	Slope-foot wash deposits	+4
7	1	Debris cone surface	+2
	2	Debris cone surface	+6
	3	Debris cone surface	+47
	4	Debris cone surface	-27

Table 7.8. Rates of net erosion (-) and accumulation (+) in (a) gullies and (b) on cone surfaces in upper Fåbergstølsdalen between 1996 and 1997. Only 24 pins were re-measured; the remaining 27 are believed to have been buried or removed by debris-flow activity.

The data exhibit a considerable spread in both net erosion and net accumulation rates. Re-measurement of pins inserted in gully sidewalls record surface lowering of 10-131 mm between June 1996 and July 1997, similar to the range of long-term surface lowering rates recorded at Fåbergstølsbreen (Table 7.6). In contrast, pins inserted in gully floors and at the foot of gully sidewalls recorded 5-232 mm of sediment accumulation. This contrast reflects movement of

sediment from the upper parts of gully sidewalls towards gully axes prior to evacuation by debris flow or wash. Given that both gully sites are located on ground deglaciated within the last 30 years (Figure 4.3, section 4.6.1), these data support the inference made above (section 7.4.2) that the transition from gully deepening to erosion of gully sidewalls as the dominant form of gully development tends to occur very soon after initial incision.

On the sampled debris cone surfaces, net accumulation of debris exhibits a very large degree of variability, as might be expected from strongly focused erosion or accumulation of sediment. Many pins indicated small or negligible changes (< 10 mm), that lie within the range of measurement error. At site 4, however, both marked localised erosion (506 mm) and accumulation (192 mm and 279 mm) were recorded, and at site 6 a total of 17 erosion pins inserted to depths of up to 912 mm were lost by removal or burial between 1996 and 1997. The mean value for the recorded sediment accumulation rates on these cone surfaces is 60 mm yr^{-1}, compared to an estimate of 8-44 mm yr^{-1} as an average rate of long-term paraglacial sediment accumulation on debris cones in neighbouring Bergsetdalen (Ballantyne, 1995a). In contrast, the mean rate of net erosion on debris cones in Fåbergstølsdalen was calculated to be *c.* 115 mm yr^{-1}, though this figure is strongly influenced by the very high amount of incision (506 mm) recorded by a single pin at site 4. Interestingly, the average net difference across all 14 debris cone pins is negligible (*c.* -2 mm), suggesting that the pins are recording overall redistribution of sediment, primarily by slopewash, rather than net addition by debris flows, which are liable to bury or remove erosion pins. Although the spatial extent of drift deposits reworked by slopewash appears limited (section 4.4.3), these data highlight the local significance of sediment redistribution by running water on cone surfaces, although in any particular year such processes are liable to be strongly focused by the location of feeder channels and distributaries on cone surfaces. In this context, it is notable that Harrison and

Winchester (1997) observed that debris cones emplaced in Patagonia within *c.* 15 years have apparently been fluvially-incised to depths of 8 m, implying a minimum average rate of incision of *c.* 500-600 mm yr^{-1}.

7.6 Summary.

1. Paraglacial slope adjustment operates primarily through the development of gully systems cut into steep valley-side drift deposits. The overall pattern is one of stripping of glacigenic sediment from the upper parts of the drift slope and redeposition of this sediment in debris cones downslope. The net result is an overall lowering of average gradient (by up to 4.5°) along gully axes, evident at all the sites investigated.

2. Further paraglacial activity may lower the overall gully-floor gradient below the threshold gradient of debris flow initiation, thereby leading to progressive atrophy and ultimate stabilisation.

3. In many cases where ungullied drift slopes possess an initial upper rectilinear slope unit steeper than *c.* 28° and exhibit pronounced concavity, gully incision has lowered the upper slope gradient and resulted in partial infill of the basal concavity. Conversely, on less concave drift slopes where the initial upper slope unit gradient is below *c.* 28° and initial failure is shallow, slight steepening of the upper slope and an increase in overall concavity have often occurred.

4. In general, paraglacial slope profile adjustment appears to be characterised by a convergence of slope profiles towards an 'equilibrium form' with an upper rectilinear slope gradient at *c.* 28° and an index of concavity of *c.* 0.22. It is notable that several authors have observed that the minimum

threshold for the initiation of hillslope debris flows is often around 28-30° (e.g. Ballantyne, 1981; Innes, 1983a).

5. The three categories of gully age investigated in Norway may be seen as representing different stages in gully development. The youngest gullies form deep, steep-sided canyons, but very rapidly broaden out and become much larger conduits of sediment transfer. After initial incision, further gully deepening is limited, but gullies become progressively wider as sidewall collapse and other processes move sediment towards the gully floor, where it is evacuated by frequent debris flows. Initial gully widening takes the form of sidewall decline, but after sidewalls have relaxed to a gradient of *c.* 25°, parallel retreat appears to predominate. Gully widening progressively reduces the width of intervening ridges of ungullied drift.

6. The final form of mature paraglacial gully systems consists of an upper bedrock-floored source area, a mid-slope area of broad gullies whose sidewalls rest at stable, moderate (< *c.* 25°) gradients, and a lower slope zone where gullies discharge onto the partially vegetated surfaces of coalescing debris cones and fans. Some gullies appear to have attained this final form and stabilised within decades of initiation, following exhaustion of readily-entrainable sediment on the upper part of the slope.

7. At all but the least active field sites, paraglacial activity has transformed steep drift-mantled valley-sides into gullied slopes where an average of *c.* 2-4 m of surface lowering has taken place within the gullied area. Under favourable conditions, gullies cut into recently-deglaciated drift slopes have reached these levels of surface lowering in less than 50 years.

8. At the most active sites in the Jostedalsbre area, these average amounts imply minimum erosion rates averaging *c.* 90 mm yr^{-1} since gully initiation, though in the Jotunheim foreland sites average erosion rates have been much lower, possibly as low as *c.* 2.5 mm yr^{-1}. However, even these relatively low rates greatly exceed 'normal' erosion rates in other environments, often by several orders of magnitude.

9. Short-term estimates of net erosion and accumulation at one site highlight the variability of highly focused paraglacial erosion and resedimentation of drift, and the importance of slopewash in locally redistributing sediment on the surfaces of debris cones.

In general, the findings reported above concerning the nature of slope modification due to paraglacial reworking of steep valley-side drifts support the sequence of changes hypothesised by Ballantyne and Benn (1994) on the basis of their more limited investigation of paraglacial hillslope modification in Fåbergstølsdalen. Their research, however, dealt mainly with changes in slope profiles and did not consider the nature of gully development, and thus the three-dimensional consequences of paraglacial drift slope modification. In a later paper, however, they presented a schematic model (Ballantyne and Benn, 1996, Figure 52.6) in which they illustrated the possible sequence of hillslope changes from gully initiation to final stabilisation. The results presented above and in previous chapters allow the development of a more detailed model of paraglacial drift slope modification; this forms the topic of the following chapter.

Chapter 8

A model of paraglacial slope modification and resedimentation

8.1 Introduction.

The research reported in this thesis has furnished a body of new information regarding the characteristics, causes and sedimentological and morphological consequences of paraglacial modification of steep, valley-side drift slopes. In most previous work on slope evolution, field measurements, laboratory experimentation and mathematical modelling of hillslope development due to mass movement have largely concerned situations where slope form changes slowly, thereby presenting problems for research (Brooks *et al.*, 1993a). One of the attractions of studying paraglacial slope modification is that the timescale over which significant changes occur is relatively brief, permitting the development of empirically-based conceptual models of hillslope evolution and modification. However, existing models of paraglacial landscape change (see chapter 2) provide only a partial explanation of the modification of drift-covered slopes, and there is scope for developing a new or refined model of paraglacial drift-slope adjustment. This chapter first outlines the limiting parameters for such a model on the basis of the findings of previous chapters (section 8.2), then presents a new model of the inferred stages of paraglacial slope modification and resedimentation (section 8.3).

In their definitive paper considering paraglacial sedimentation, Church and Ryder (1972) suggested a model to represent the paraglacial sediment cycle associated with fluvial sediment transport. The underlying assumption of the paraglacial concept as defined by Church and Ryder was that deglaciation represents a fundamental change in the terrestrial erosional environment. Sediment that may have attained stability within the glacial depositional

environment may be potentially unstable with respect to the subaerial environment that succeeds it. Thus for periods of several thousands of years or more, sediment yield from deglaciated terrain may be unrelated to the primary production of debris by nonglacial processes (Church and Slaymaker, 1989; Harbor and Warburton, 1993; Brooks, 1994). Deglaciated terrains consequently fail to conform with either dynamic equilibrium models of landscape evolution (e.g. Johnson, 1984a; Slaymaker, 1984; Church and Slaymaker, 1989) or conventional concepts of drainage basin sediment delivery (e.g. Church *et al.*, 1979; Leonard, 1986a; Harbor and Warburton, 1993; Ashmore, 1993; Leemann and Niessen, 1994). Indeed, Benn and Evans (1998) have recently suggested that it is possible for a glaciated landscape to fail to reach its nonglacial sediment equilibrium before the onset of the next glaciation. Only Ballantyne and Benn (1996) have attempted a schematic explanation of how drift-mantled slopes within paraglacial landsystems might respond to deglaciation (Figure 2.5). However, their model was based mainly on two-dimensional changes in slope profiles and limited qualitative observations at a single site (upper Fåbergstølsdalen) in Norway.

8.2 Limiting parameters.

The conclusions reached in previous chapters are here used as the basis for modelling the sequence of paraglacial slope modification and resedimentation. Withdrawal of glacier ice at the study sites has exposed valley-side drift slopes that usually exhibit a steep upper rectilinear slope, interpreted as predominantly resulting from the collapse of lateral moraines, and basal concavity. At all but the least active sites, drift surface relief is characterised by gully incision and the associated development of debris cones at the slope foot (chapters 4 and 5). The dominant process of paraglacial sediment transfer is debris flow, although surface wash and snow avalanches play a secondary role in sediment redistribution at some sites. Gullying of drift tends to be widespread on slopes where gradient,

sediment availability and water supply are favourable. Analysis of sub-surface sediment units and depositional structures (chapter 6) has shown that many drift slopes have been extensively reworked by debris flows and slopewash, consistent with the interpretation of drift surface relief. The presence of ancient paraglacial deposits underlying *in situ* till in gully sections implies cyclic alternation of glacial and paraglacial sediment transfer during glacial advances and retreats respectively. At the most active sites, paraglacial deposits have accumulated very rapidly following deglaciation, often within decades of exposure to nonglacial processes, and at some locations the juxtaposition of *in situ* glacigenic deposits and reworked valley-side sediment indicates penecontemporaneity of deglaciation and paraglacial remobilisation of drift. Most significantly, analysis of surveyed slope profiles and gully dimensions on drift-mantled slopes on both recently- and formerly-deglaciated terrains (chapter 7) permits the detailed reconstruction of three-dimensional paraglacial slope evolution and associated sediment redeposition to be integrated in a sequential model based on field data.

The morphological, sedimentological, historical and process evidence presented in chapters 4-7 therefore suggests that modelling of paraglacial slope behaviour must consider the modification of drift by repeated downslope mass transport of a steep, upper rectilinear slope and basal concavity, often extremely rapidly after deglaciation. However, there is evidence at some sites for multiple phases of drift reworking. Buried soils developed on the tops of individual sediment units within accumulations of reworked drift in the Scottish Highlands indicate that episodic paraglacial reworking was delayed or renewed several thousands of years after deglaciation. Consequently, an appropriate model of drift slope adjustment also needs to take into account possible delayed or renewed episode(s) of reworking within a long history of intermittent modification of drift deposits.

8.3 Proposed stages of paraglacial slope modification and resedimentation.

The research reported in this thesis has focused primarily on the characteristics, causes and consequences of paraglacial modification of drift slopes. In the following discussion, the sequence of paraglacial slope adjustment and sediment redeposition is divided into several stages that represent progressive hillslope response following deglaciation. It is important to note that the discrete phases of paraglacial modification illustrated below are schematic, as investigations of the extent and timing of paraglacial reworking of valley-side drifts have shown that the behaviour of such slopes has been locally variable (chapters 4 and 5). The model outlined here may therefore be considered as a generalised conceptual framework for paraglacial slope modification and resedimentation. It treats two scenarios separately: the first concerns paraglacial response immediately after deglaciation, and the second concerns delayed or renewed paraglacial slope modification. For each scenario the model is illustrated with reference to particular examples from the field sites.

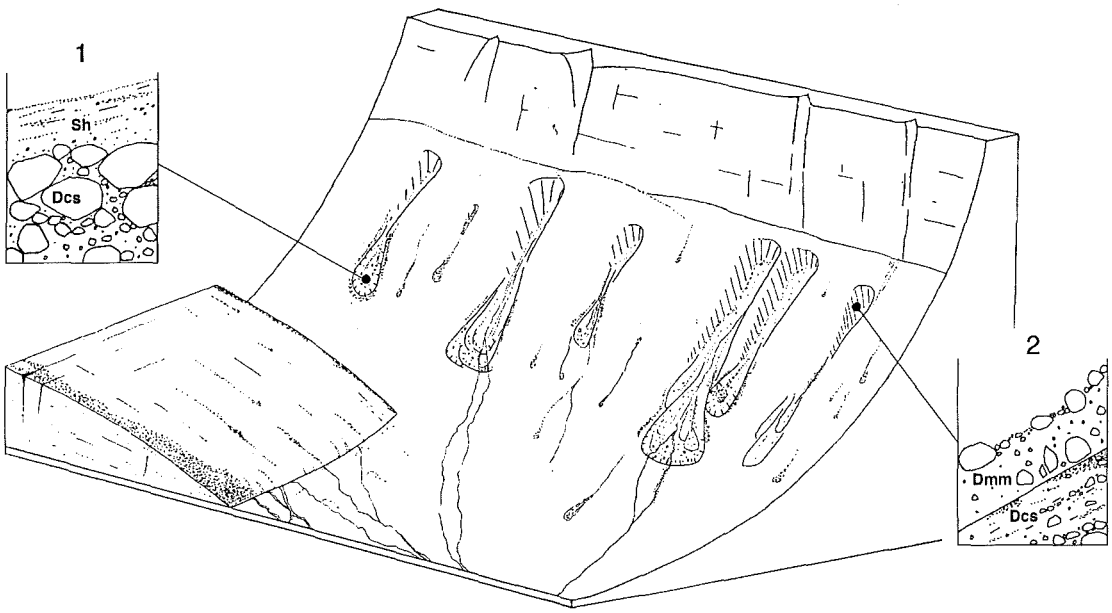
8.3.1 Scenario 1: Paraglacial response immediately following deglaciation

The stages envisaged in paraglacial response of drift are described and illustrated below in a series of schematic block diagrams (Figures 8.1 and 8.2), then summarised in graphs that show how individual parameters change through time (Figure 8.3).

Stage 1: Gully incision

Deglaciation of upland valleys exposes a valley-side cover of steep, unvegetated glacigenic deposits. Such deposits are essentially in a metastable

Stage 1



Stage 2

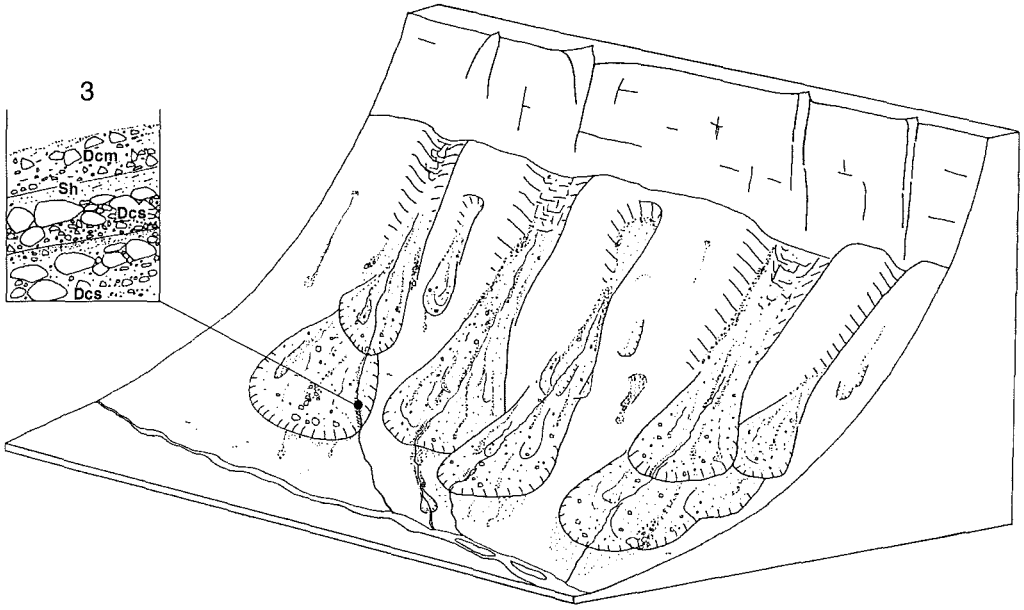
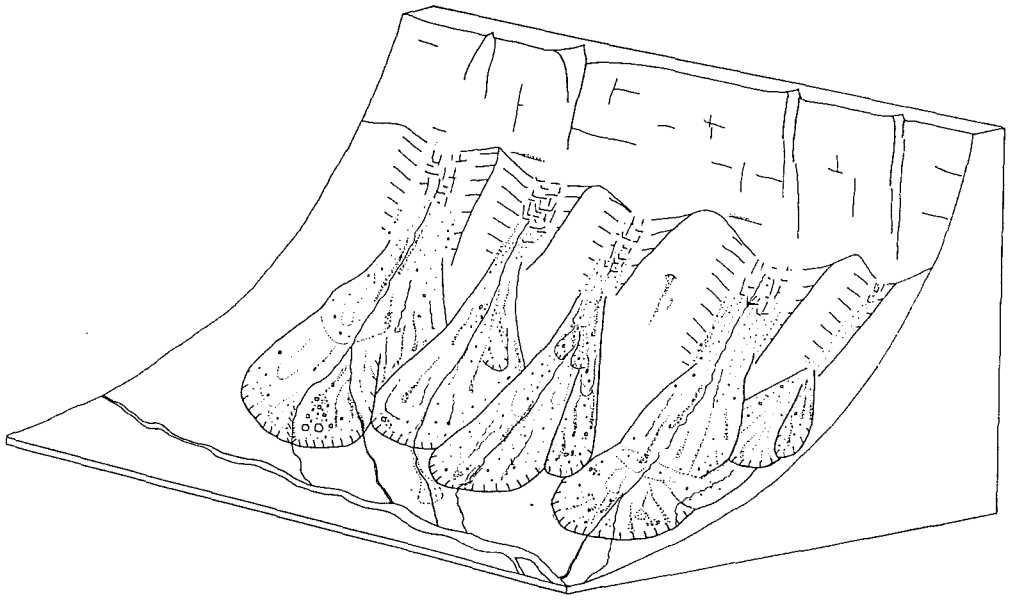


Figure 8.1. Stages 1 and 2 of paraglacial slope modification and resedimentation immediately following deglaciation. Key for sections in Figure 5.13.

Stage 3



Stage 4

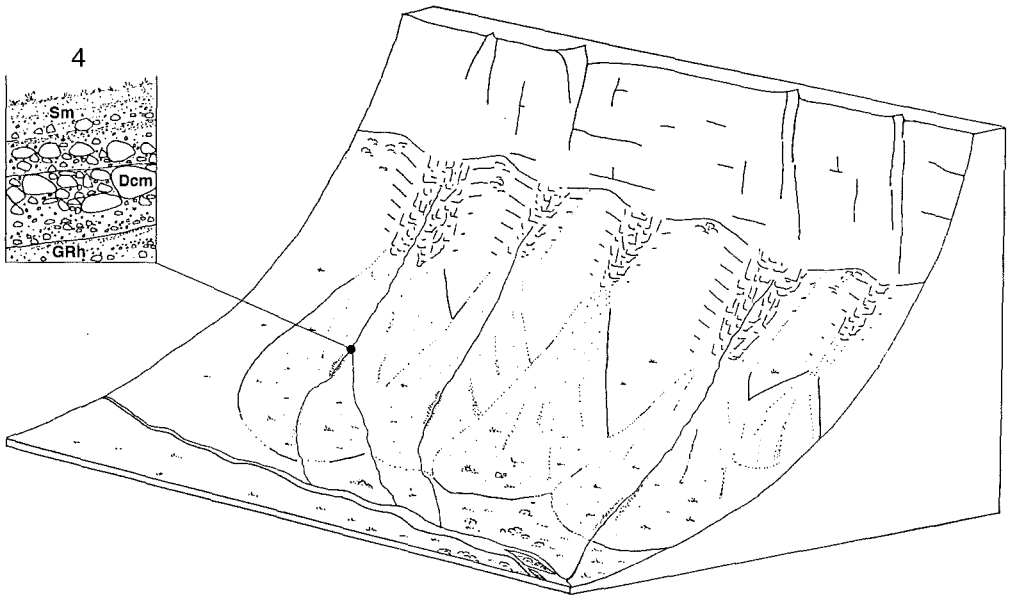


Figure 8.2. Stages 3 and 4 of paraglacial slope modification and resedimentation immediately following deglaciation. Key for sections in Figure 5.13.

state, with a low threshold of resistance to erosion and failure, and are thus vulnerable to small extrinsic changes or perturbations, particularly increases in runoff or porewater pressure (Ryder, 1971a). Stage 1 (Figures 8.1) depicts the onset of paraglacial modification of valley-side drift, primarily by slope failures and associated debris flows triggered during extreme rainstorm or snowmelt events. The location of initial failures may be partly determined by that of rockwall gullies that focus delivery of water to the drift surface downslope. Once established, failure scars and gullies concentrate both surface and subsurface flow of water, and provide sites for snow accumulation and late snow-lie. In the early stages of paraglacial drift-slope modification, gullies deepen rapidly through removal of gully-floor sediment by successive debris flows and by snowmelt or storm runoff. Rapid incision produces very steep gully sides that are prone to undermining and collapse, providing loose sediment to the gully floors, from where it is removed by wash and debris flows. Within gullies, there is usually an overall decline in slope gradient, arising in part from headward gully deepening and in part from the redeposition of sediment downslope in the form of debris lobes and nascent debris cones. Sediment structures within such lobes may reveal one or more debris flow units and/or fine-grained slopewash deposits overlying *in situ* till (Figure 8.1, section 1). Gully deepening may reveal the subsurface stratigraphy of the valley-side drift, possibly including earlier paraglacial sediments underlying *in situ* till (Figure 8.1, section 2; see Figures 6.22 and 6.23). This initial stage of drift slope modification is exemplified on the most recently-deglaciated terrain close to the snouts of Fåbergstølsbreen and Lodalsbreen (Figure 7.22a). This stage represents the waxing period of paraglacial drift-slope modification: as the number of active gullies progressively increases, so the pace of sediment reworking by debris flows and runoff consequently accelerates.

Stage 2: Gully extension and widening

Stage 2 of paraglacial drift-slope modification is dominated by headward extension and widening of gullies, and the accumulation of debris cones at the slope foot (Figure 8.1). As reduction in gully-floor gradient and localised exposure of bedrock progressively limit further gully-floor deepening, gully evolution tends to take the form of headward extension to the foot of the rockwall and gully widening through failure and collapse of sidewalls. Both processes contribute sediment to gully floors, where it is evacuated by intermittent debris flows, runoff from late-lying snowbeds at gully heads, rainstorm runoff and, more locally, by snow avalanches. During this stage, further slope failures may initiate new gullies on previously-ungullied slopes, so that gullies at different stages of development coexist. In the oldest gullies, bedrock is increasingly exposed at gully heads, where downcutting ceases completely (Figure 8.2). Where mature gullies are closely spaced, progressive widening in the form of parallel retreat of gully walls reduces the width of intervening areas of *in situ* drift until adjacent gullies are separated by sharp-crested arêtes, and ultimately erosion of such divides may result in the development of complex gully systems, lateral merging or 'capture' of gullies, and the formation of broad 'master' gullies. Downslope from the oldest gullies, substantial debris cones accumulate, with a surface microrelief of debris-flow levées and lobes. Such cones are composed of stacked diamictons emplaced by successive debris flows with occasional intercalated slopewash horizons (Figure 8.1, section 3). The latter represent reworking of loose sediment by surface runoff during rainstorms or spring snowmelt. Observations at sites of recent glacier retreat in Norway suggest that the transition from stage 1 to stage 2 is often rapid. In Fåbergstølsdalen, for example, landforms characteristic of stage 2 have developed after as little as 25 years (Figure 7.22b). This stage marks the peak of paraglacial sediment reworking, being characterised

by a high density of gullies and an abundant supply of sediment from collapsing sidewalls.

Stage 3: Maximum development of gullies

The third stage of paraglacial modification of drift-mantled slopes is reached when gullies achieve their maximum lateral extent. Bedrock is now extensively exposed in funnel-shaped gully heads, and vertical incision has largely ceased. Sediment transport is limited to reworking of debris supplied from gully sidewalls, which progressively decline to stable angles of *c.* 25°. As sediment supply to gully axes declines, the rate of sediment transport by intermittent debris flow events and surface runoff decreases. In consequence, less sediment reaches the valley floor, resulting in increased sediment accumulation near the apices of debris cones. As a result, these tend to extend headward, backfilling the lower parts of gullies and ultimately reaching the zone of exposed bedrock. On the valley floor the debris cones at the slope foot coalesce to form a continuous apron, into which shallow stream courses are incised. This stage marks the waning phase of paraglacial drift-slope modification, during which the supply of debris to gully systems progressively diminishes, and consequently the rate of sediment reworking slows down. At this stage, too, significant vegetation colonisation begins on inactive debris cone surfaces and on gully sidewalls, further reducing the amount of sediment in transport.

As even the oldest gullies in recently-deglaciated areas of Fåbergstølsdalen have not yet achieved this stage, it seems likely that it is only achieved after more than 50 years after gully initiation. However, some features of stage 3 (coalescing debris cones, degraded gully sidewalls and extensive exposures of bedrock at gully heads) are evident even within 50 years of exposure by deglaciation, and the presence of inactive paraglacial debris cones that formed within *c.* 100-200 years

of glacier retreat at nearby Bergsetdalen (Ballantyne, 1995a) suggests that stage 3 is reached, at least in this area, within 50-100 years after deglaciation.

Stage 4: Stabilisation

The final stage of paraglacial modification of steep drift slopes is characterised by sediment exhaustion and cessation of debris flow activity. This occurs when gully sides stabilise through slope decline and vegetation encroachment, so that very limited sediment reaches gully floors (Figure 8.2). Debris cone accretion consequently ceases, and through time the surface microrelief on cone surfaces degrades so that features such as debris flow levées and lobes can no longer be distinguished. Cone surfaces became largely or completely vegetated, except where they are incised by small stream channels. Geomorphic activity at this stage is confined to localised slopewash and limited fluvial erosion, evident in exposures in the form of a final layer of colluvial sediment overlying debris-flow deposits (Figure 8.2, section 4). Morphologically, this final stage is dominated by a broad apron of relict coalescing debris cones that extend upslope to meet extensive areas of exposed bedrock, with limited survival of unmodified drift between some cones. This final stage of paraglacial slope development is illustrated by some sections of valley-side slopes in Leirdalen and Visdalen (Figure 7.22c).

Changes in individual parameters

The sequence of events outlined above can also be expressed in terms of changes in individual parameters (Figure 8.3). Rates of sediment transfer climb steeply in stage 1 due to rapid vertical incision of newly-formed gullies, peak in stage 2, when the gully system reaches maturity and gully sidewalls supply abundant sediment, and decrease more slowly in stage 3, as sidewalls retreat from

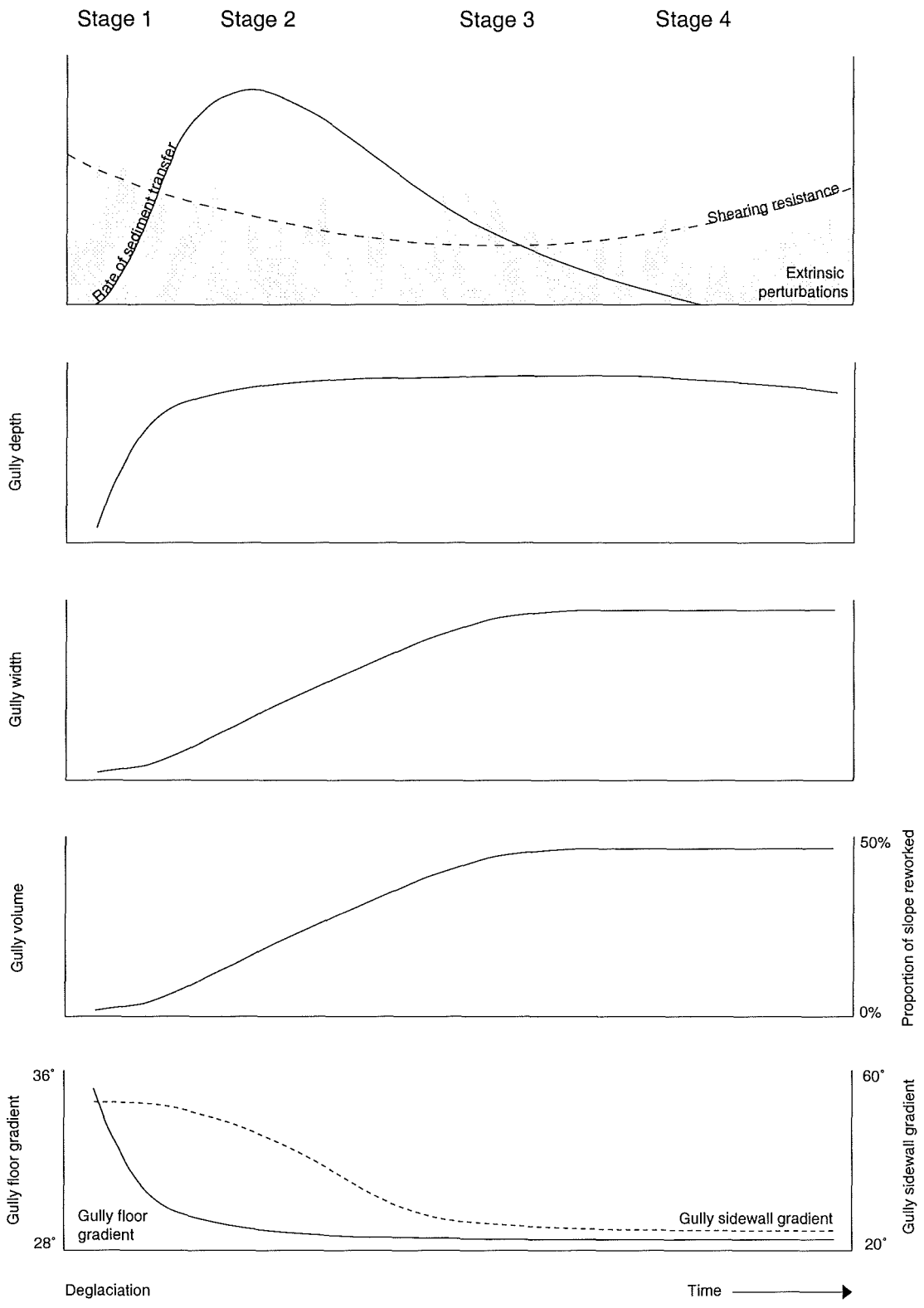


Figure 8.3. A model of paraglacial slope modification illustrating four stages of slope adjustment and resedimentation immediately following deglaciation.

the gully axes and decline to gentler gradients. Gully depth increases dramatically during stage 1, during the period of rapid downcutting, but then changes little thereafter due to lessening of overall gully-floor gradient and increasing exposure of bedrock in gully floors. In contrast, gully width increases more gradually throughout stage 2 and, more slowly, during stage 3. Changes in gully width cease only in stage 4 as gully sides stabilise. As changes in gully width rather than gully depth dominate changes in gully volume (chapter 7), the latter display a similar temporal pattern. Upper drift slope gradient in gullies tends to decline rapidly during stage 1, reflecting rapid incision of gullies, but to remain largely unchanged thereafter. In contrast, gully sidewall gradient initially remains unchanged throughout stage 1 and the earlier part of stage 2, as undercutting maintains steep sidewall slopes, then exhibits a gradual decline until it reaches a stable gradient in stage 3.

8.3.2 Scenario 2: Delayed or renewed paraglacial modification of drift-mantled slopes

Although Ryder (1971a) and Church and Ryder (1972) related the term 'paraglacial' to the period of relatively rapid sedimentation during or immediately after deglaciation, when deglaciated terrain is in the process of transition from predominantly glacial to dominantly-fluvial conditions, they also acknowledged that paraglacial sediment transfer continues as long as drift remains accessible to fluvial erosion and transportation, which may be long after the initial 'paraglacial period'. Their original definition thus left the term somewhat open to interpretation. If 'paraglacial' processes are temporally constrained by a 'paraglacial period' of rapid reworking and redeposition of unstable sediment immediately after deglaciation, then further drift remobilisation centuries or millennia after deglaciation may not be considered a strictly 'paraglacial' response. However, within this thesis and elsewhere the continued influence of a supply of

glacigenic sediment rather than any ill-defined temporal constraint is considered the defining criterion of paraglacial activity. It follows that the environment under which such sediment is reworked or the length of the delay between drift deposition and subsequent reworking is immaterial. This view implies that 'paraglacial' *sensu lato* refers to the long-term inheritance of glacial landforms and sediments in the landscape, and their influence on subsequent geomorphic activity, irrespective of length of time elapsed since deglaciation. The second scenario of the model of drift slope adjustment outlined below aims to illustrate the effects of one or more possible episodes of delayed or renewed drift reworking.

At locations where paraglacial modification of drift-covered hillslopes has been delayed (rather than renewed) the general form of the model outlined in section 8.3.1 will still apply, though the onset of sediment reworking (stage 1) will be preceded by a prolonged period of slope stability and hence represent incision of previously stable, and often completely vegetated drift slopes. Consequently, for delayed paraglacial activity, each of the curves illustrated in Figure 8.3 is simply displaced to the right along the horizontal axis representing time elapsed since deglaciation (Figure 8.4). In other respects, the stages of delayed paraglacial drift slope modification are likely to resemble those illustrated schematically in Figures 8.1 and 8.2, though it is possible that gully sidewall decline may be delayed owing to the influence of a stabilising vegetation cover on intact slopes between adjacent gullies. The Drumochter Pass site (Figure 5.9) offers a possible type site for delayed paraglacial reworking of steep-drift-mantled slopes (Ballantyne and Benn, 1996), though it is uncertain whether drift reworking at this site represents delayed paraglacial activity *sensu stricto*, or recent reactivation of a paraglacial gully system of much greater antiquity.

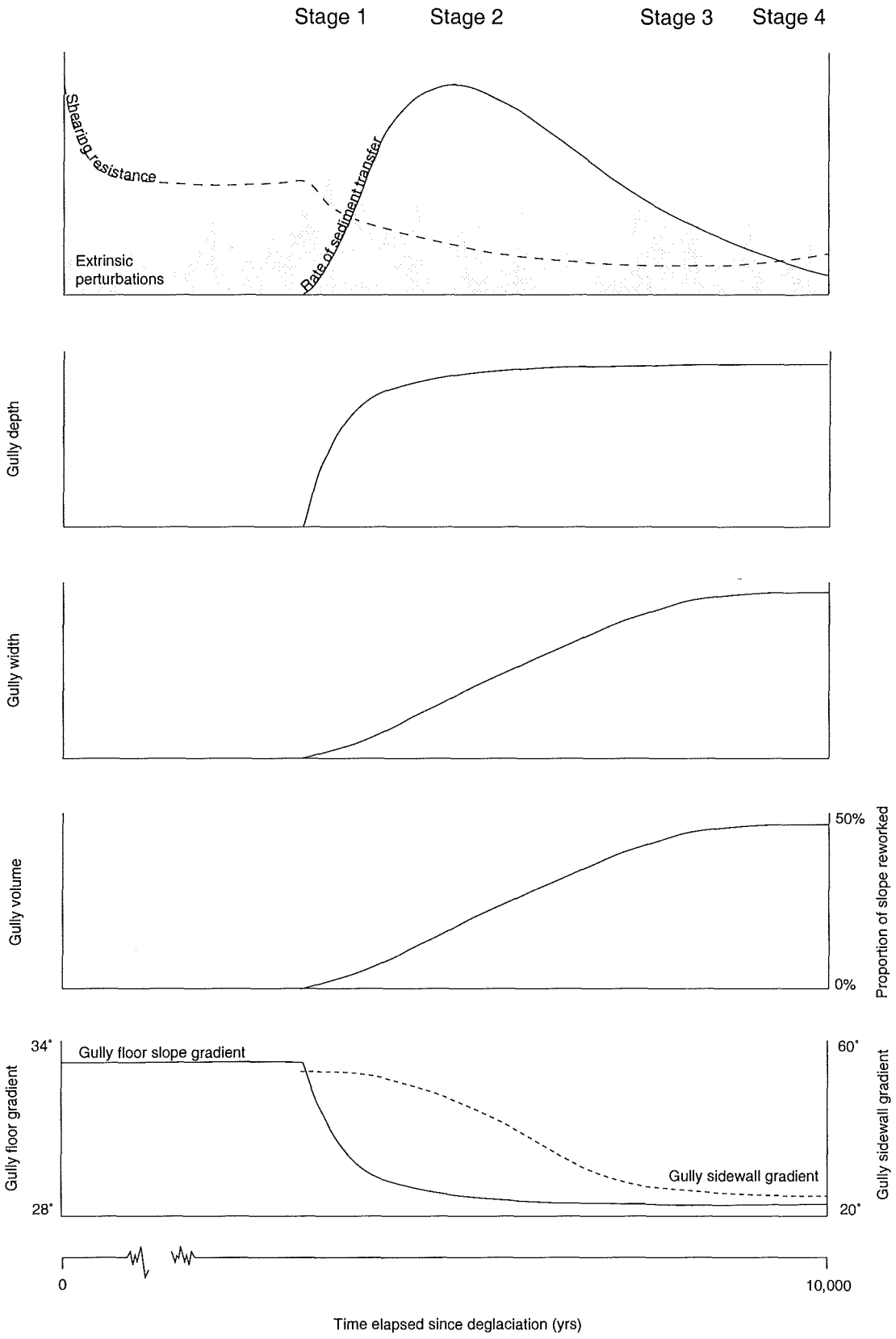


Figure 8.4. A hypothesised model of delayed paraglacial slope modification illustrating four stages of slope adjustment and resedimentation.

Two possible explanations may account for such a delayed response to deglaciation. First, the shearing resistance of such slopes may have declined through time, due, for example, to the influence of pedogenesis (Brooks, 1997; Brooks *et al.,* 1993b, 1995; Brooks and Richards, 1993, 1994) or possibly anthropogenic influences such as changes in vegetation cover, burning or introduction of grazing animals (Innes, 1983b, Ballantyne, 1991a). Second, (and possibly in combination with the first), the magnitude of extrinsic perturbation in the form of exceptional rainstorm or snowmelt events may have been insufficient to trigger initial slope failure and thus gully development at the time of deglaciation, but to have reached greater levels at some time in the Holocene.

Alternatively, the onset of Late Holocene paraglacial resedimentation of valley-side drifts (as outlined in section 5.6.2) may represent renewed, rather than delayed, paraglacial reworking. In this case, a fifth stage of paraglacial slope modification may be invoked to illustrate this more complicated scenario of hillslope modification. Stage 5 of the model of paraglacial slope modification (Figures 8.5) represents a phase of rejuvenation of previously stable gully systems. This is most likely to occur at sites where sediment supply was not completely exhausted from upper drift slopes following an initial phase of paraglacial reworking, and to be triggered by unusually extreme rainstorm events that cause a rise in porewater pressures to critical levels, causing failure of previously-stable slopes or gully walls. Slope failure and erosion by debris flows resulting from such a storm may strip vegetation cover from gullies, thus lowering the threshold for subsequent debris flow events and remobilisation of loose sediment (Brazier and Ballantyne, 1989).

Figure 8.5 illustrates the landforms characteristic of renewed paraglacial modification of drift-mantled hillslopes. Reactivation is evident in the form of unvegetated gullies, often located within larger, vegetated gullies, and fresh

Stage 5

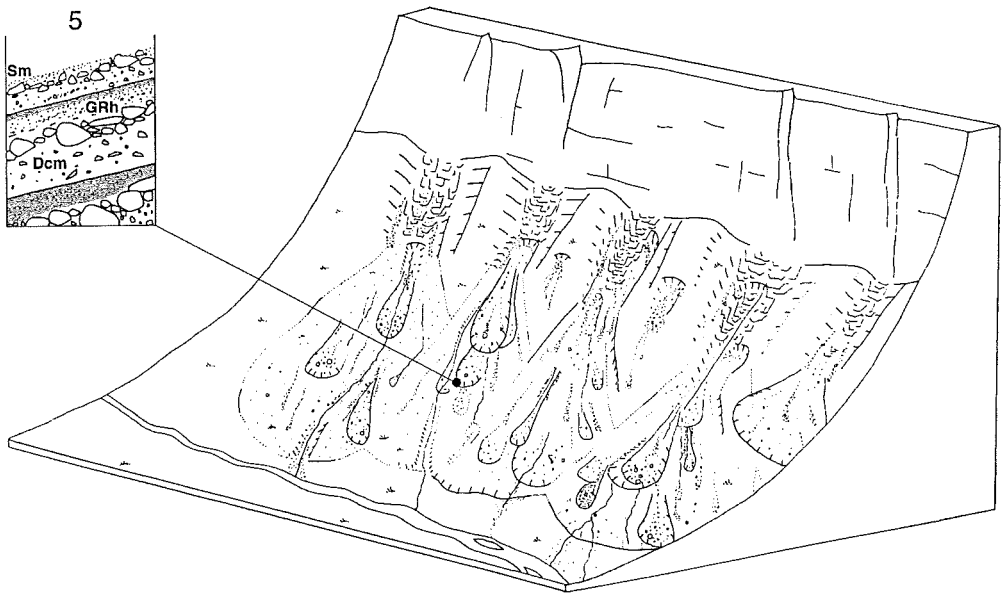


Figure 8.5. Stage 5 of paraglacial slope modification and resedimentation representing renewed paraglacial activity. Key for section in Figure 5.13.

accumulations of reworked sediment in the form of debris levées and lobes and splays of slopewash sediments on the vegetated surfaces of debris cones. Sections cut through such cones may reveal not only the stacked debris-flow units and intercalated slopewash deposits characteristic of mature cones in recently-deglaciated environments, but also buried palaeosols or peat layers. Such organic layers provide evidence for multiple phases of reactivation of ancient gully systems, and intervening periods of stability characterised by soil formation or peat growth. The active gully systems in Leirdalen and Visdalen and at some of the Scottish sites probably represent Late Holocene rejuvenation of paraglacial gullies that initially formed following deglaciation in Lateglacial or Preboreal times (e.g. Figure 5.11).

Probable changes in various parameters associated with renewed paraglacial modification of drift slopes are illustrated in Figure 8.6. Rate of sediment transfer increases rapidly after a prolonged period of dormancy. As ancient gullies are likely to have been excavated to the underlying rockhead, reactivation is unlikely to result in further gully deepening beyond initial excavation of debris that has accumulated in the interim. The predominant change is likely to be reflected in renewed and initially rapid gully widening, and an accompanying increase in gully volume. As renewed incision is negligible, the gradient of gully floors is unlikely to undergo significant change, but gully sidewalls experience rapid initial steepening through collapse and undermining. Renewed paraglacial activity is likely to be terminated in one of two ways, either when sediment supply to gullies is exhausted (as in the model of paraglacial reworking of recently-deglaciated terrain), or when a prolonged respite from geomorphologically-significant storm events permits relaxation of gully walls to stable angles and recolonisation by vegetation.

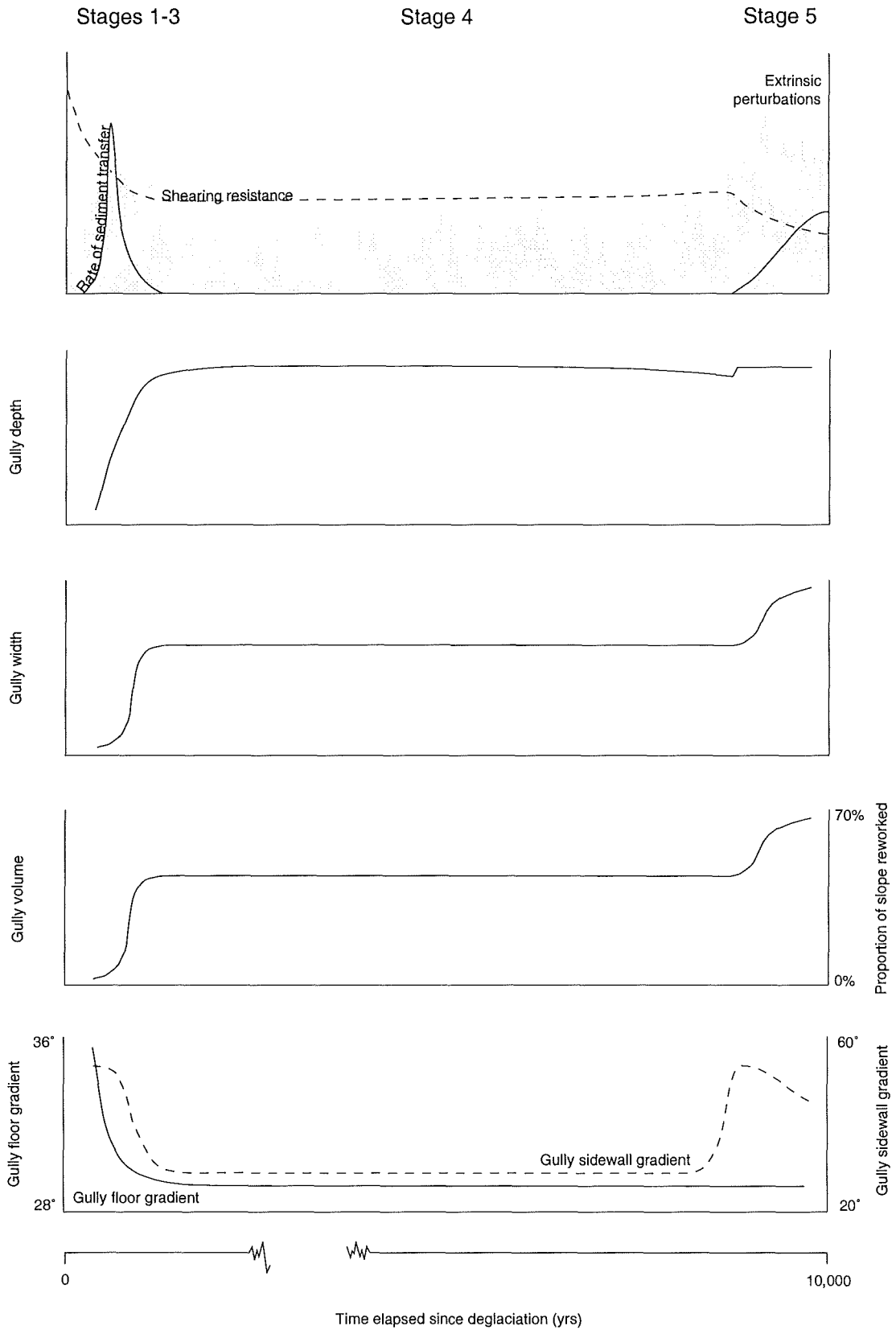


Figure 8.6. A hypothesised model of renewed paraglacial slope modification illustrating five stages of slope adjustment and resedimentation.

8.4 Summary.

1. Following ice retreat, glacial sediments and landforms in upland environments are commonly subject to extensive reworking by paraglacial processes, owing to the abundance of steep slopes, the action of high-energy processes, and the widespread availability of unconsolidated sediment (*cf.* Owen *et al.*, 1995; Benn and Evans, 1998). The stages of paraglacial slope adjustment and resedimentation outlined above describe the progressive modification of valley-side slopes and accompanying resedimentation of drift that result from these changes.

2. The principal effects of repeated reworking of valley-side drifts are overall lowering of the gradient of the slope, stripping of drift cover from the uppermost part of the slope, and the formation of a broad apron of coalescing debris cones built from accumulating debris-flow and colluvial deposits.

3. The models outlined above illustrate the sequence of morphological and sedimentological changes that occur during overall slope transformation, not only for recently-deglaciated terrain, but also for environments where paraglacial reworking of drift was delayed or renewed after a long period of dormancy. Their value lies in providing a set of morphological and sedimentological criteria that will allow the characteristics of both recent and ancient paraglacial activity to be identified, thus contributing to fuller appreciation of paraglaciation as a component of landscape history in glaciated upland environments.

Chapter 9

Conclusions and future prospects

9.1 Introduction.

The aims of the research reported in this thesis have been essentially threefold: first, to establish the nature and extent of paraglacial modification of deglaciated drift slopes, and its sedimentological implications; second to identify the factors that condition or trigger paraglacial reworking of drift-mantled hillslopes; and third, to model the temporal pattern of paraglacial modification of drift-mantled hillslopes. The main findings relating to these original aims are summarised below, and the most promising areas for further research are briefly explored.

9.2 Principal findings.

9.2.1 The nature and extent of paraglacial slope modification

The first aim of this thesis was to 'identify and establish the nature and extent of paraglacial modification of deglaciated drift slopes, and its sedimentological implications'. The most conspicuous legacy of drift slope modification at all sites studied takes the form of debris cones, fans and aprons located along the flanks of valleys downslope of gullies. These features support a surface microrelief of paired levées, thick terminal lobes and poorly sorted deposits which demonstrate that debris flow has been the dominant process of paraglacial sediment transfer. Secondary paraglacial processes include slopewash and snow avalanching.

The extent of paraglacial activity varies considerably between sites, but may be extremely widespread where conditions are favourable. At four recently-deglaciated sites investigated around Jostedalsbreen, gully density is of the order of 10-100 gullies per kilometre, reflecting resedimentation of an estimated 44% of all valley-side drift. Drift slopes at sites in Jotunheimen are generally much less extensively modified by paraglacial processes (gully density nowhere exceeds 8 km^{-1} at the sites studied), and no more than *c.* 11% of hillslope drift appears to have been reworked by gully erosion. At sites in the Scottish Highlands, gully density is highly variable, but nowhere exceeds 25 gullies per kilometre, reflecting remobilisation of less than 7% of valley-side drift.

Comparison of the sedimentological characteristics of glacigenic drift reworked by debris flow with those of *in situ* till demonstrates that the former retains many of the sedimentological characteristics of the latter, and that the two cannot be readily distinguished in terms of clast shape, angularity or texture, fine-fraction granulometry or packing (void ratio). The combined use of clast fabric analysis (preferred orientation) and large-scale structural and lithofacies relationships, however, appears to offer a reasonably robust method for differentiating reworked from *in situ* glacigenic deposits. Application of these diagnostic criteria to the interpretation of valley-side drift deposits at localities in southern Norway has shown that in some cases these comprise recent *in situ* glacial drift deposits overlying crudely stratified diamictons that represent paraglacial reworking of glacigenic deposits by debris flows following deglaciation in late Preboreal times, indicating cyclic alternation of glacial and paraglacial sediment transfer. In the Western Red Hills on the Isle of Skye, sedimentological and stratigraphic evidence indicates that paraglacial remobilisation of drift accompanied deglaciation at the end of the Loch Lomond Stade.

9.2.2 Factors influencing paraglacial slope modification

The second aim of this thesis was to 'identify the factors that condition or trigger paraglacial reworking of drift-mantled hillslopes'. This is important if the constraints on, and causes of, paraglacial slope activity are to be understood. Factors controlling paraglacial drift modification by debris flows include gradient, sediment characteristics and water supply. At sites studied in Norway, gully density and hence paraglacial reworking of glacigenic sediment is greatest where initial drift deposits form slopes steeper than *c.* 30° and are thicker than *c.* 10 m, and where the void ratio of unreworked sediment exceeds *c.* 0.35. Widespread gullying is also favoured at sites of high and focused water input, in particular where melting snow and ice are involved. At sites studied in Scotland, gullying is extensive only where drift slopes are steeper than *c.* 30° and drift cover exceeds *c.* 3 m, and where the void ratio of unreworked sediment exceeds *c.* 0.4. Both the amount of precipitation and focusing of water delivered onto valley-side drift are seemingly instrumental in conditioning gullying intensity, and may determine the extent of delayed or renewed paraglacial reworking at sites where the intrinsic constraints on drift reworking are satisfied. Rainstorms and rapid snowmelt are considered the most likely triggers of drift-slope failure and reworking of sediment by debris flow at the sites investigated.

9.2.3 The temporal pattern of paraglacial slope modification

The third aim of this thesis was to 'model the temporal pattern of paraglacial modification of drift-mantled hillslopes'. The aim of this part of the research was to establish the developmental characteristics of paraglacial hillslope adjustment, and to determine whether these follow a consistent pattern.

Paraglacial slope adjustment operates primarily through the development of gully systems cut into steep valley-side drift deposits. The overall pattern is one of stripping of glacigenic sediment from the upper parts of the drift slope and redeposition of this sediment in debris cones downslope. The net result is an overall lowering of average gradient (by up to 4.5°) along gully axes, evident at all the sites investigated. Gully deepening may lower the overall gully-floor gradient below the threshold gradient for debris flow initiation, thereby leading to progressive atrophy and ultimate stabilisation. However, the detailed form of slope adjustment varies considerably. At most sites where ungullied drift slopes consist of an initial upper rectilinear slope unit steeper than *c.* 28° and a pronounced basal concavity, gully incision has lowered the upper slope gradient, and sediment redeposition has resulted in partial infill of the concavity. Conversely, on less concave drift slopes where the initial upper slope unit gradient is below *c.* 28° and initial failure is shallow, slight steepening of the upper slope and an increase in overall concavity have often occurred. Though there are exceptions to this pattern, in general paraglacial slope profile adjustment appears to be characterised by a convergence of slope profiles towards an 'equilibrium form' with an upper rectilinear slope gradient at *c.* 28° and an index of concavity of *c.* 0.22. It is notable that several authors have observed that the minimum threshold for the initiation of hillslope debris flows is often around 28-30°, suggesting that this represents the fundamental control on the final form of paraglacially-modified hillslopes.

The process of gully development associated with paraglacial reworking of drift-mantled slopes has been reconstructed by comparative analysis of slopes deglaciated at different times. The youngest gullies form deep, steep-sided canyons, but very rapidly broaden out and become much larger conduits of sediment transfer. After initial incision, further gully deepening is limited by exposure of bedrock on gully floors, but gullies become progressively wider as

sidewall collapse and other processes move sediment towards gully axes, where it continues to be evacuated by intermittent debris flows. Initial gully widening takes the form of sidewall decline, but after sidewalls have relaxed to a gradient of *c.* 25°, parallel retreat of sidewalls appears to predominate. Gully widening progressively reduces the width of intervening ridges of ungullied drift, sometimes forming steep-sided arêtes as neighbouring gullies converge, and occasionally resulting in the partial or complete removal of gully divides. The final form of mature paraglacial gully systems consists of an upper bedrock-floored source area, a mid-slope area of broad gullies whose sidewalls rest at stable, moderate (< *c.* 25°) gradients, and a lower slope zone where gullies discharge onto the partially vegetated surfaces of coalescing debris cones and fans. Some gullies appear to have attained this final form and stabilised within several decades of initiation, following exhaustion of readily-entrainable sediment on the upper part of the slope.

In Norway, at all but the least active field sites, paraglacial sediment reworking has transformed steep drift-mantled valley-sides into gullied slopes where an average of *c.* 2-4 m of surface lowering has taken place within the gullied area. Under favourable conditions, gullies cut into recently-deglaciated drift slopes have reached these levels of surface lowering in less than 50 years. At the most active sites in the Jostedalsbre area, erosion rates have averaged *c.* 90 mm yr^{-1} since gully initiation, though in the Jotunheim foreland sites average erosion rates have been much lower, possibly as low as *c.* 2.5 mm yr^{-1}. However, even these relatively low rates greatly exceed 'normal' erosion rates in other environments, often by several orders of magnitude, thus highlighting the extreme rapidity of paraglacial erosion of recently-deglaciated drift-mantled slopes.

An additional feature of this investigation was to examine the timing and causes of delayed or renewed paraglacial slope modification and resedimentation,

with particular reference to the Scottish Highlands. Analysis of eighteen radiocarbon dates suggests that delayed or renewed paraglacial modification of valley-side drift at the Scottish field sites represents localised, intermittent activity since *c.* 6.5 cal ka BP. When combined with a broader dataset of hillslope reworking of glacigenic drift and rockfall talus slopes throughout the Scottish Highlands, however, broad coincidences of timing are evident at *c.* 2.7 to 1.7 cal ka BP and after *c.* 0.7 cal ka BP. In combination with analysis of sub-fossil pollen and charcoal, this chronology suggests that recent reworking of ancient drifts may partly reflect the influence of extreme rainfall events during periods of climatic deterioration at *c.* 2.7 to 2.3 cal ka BP and after *c.* 0.3 cal ka BP, and at one site, anthropogenic interference with the vegetation cover in recent centuries. The possibility that such delayed or renewed activity was initiated by destructive storms of random occurrence, however, cannot yet be dismissed. Pending further research, these possible links remain conjectural.

In general, the findings reported in this thesis concerning the temporal pattern of slope modification due to paraglacial reworking of steep valley-side drifts support the sequence of two-dimensional changes hypothesised by Ballantyne and Benn (1994, 1996). However, the results presented here have allowed the development of a more detailed model of paraglacial drift slope adjustment and resedimentation, the stages of which describe the progressive modification of valley-side slopes and accompanying resedimentation of drift that result from these changes. The principal effects of repeated reworking of valley-side drifts are overall lowering of the gradient of the slope, stripping of drift cover from the uppermost part of the slope, and the formation of a broad apron of coalescing debris cones built from accumulating debris-flow and colluvial deposits. The model proposed in this thesis illustrates the sequence of morphological and sedimentological changes that occur during overall slope transformation, not only for recently-deglaciated terrain, but also for environments

where paraglacial reworking of drift was delayed or renewed after a long period of dormancy.

9.3 Future research.

The findings reported in this thesis suggest that paraglacial reworking of steep, drift-mantled slopes by debris flows during and immediately after deglaciation is a widespread phenomenon, and appears to be the 'normal' geomorphic response to deglaciation where favourable intrinsic controls and extrinsic triggers exist. Whilst debris flows have been observed on drift-mantled slopes in mountainous terrain in Europe, North America, the Himalayas, Japan, New Zealand and Antarctica, few studies (Li Jijun *et al.*, 1984; Owen *et al.*, 1995) have hitherto sought to assess the spatial extent of paraglacially-reworked drift deposits. Consequently, there is potential for further research to establish the generality of these findings, particularly with regard to valley-side drifts developed on a range of lithologies or under different tectonic, topographic and climatic circumstances.

Furthermore, by establishing the sedimentological effects of paraglacial reworking of drift, more accurate assessment of the spatial extent of paraglacial deposits may be permitted, particularly in areas where the morphological effects of sediment reworking are equivocal or indistinct. Sedimentological analyses employed during the current research have confirmed previous suggestions (e.g. Lawson, 1988; Ballantyne and Benn, 1994) that preferred clast orientation and structural and lithofacies relationships are the most valuable criteria for distinguishing *in situ* drift deposits from those reworked by debris flow activity. However, where exposures display ambivalent sedimentary features or highly-complex sedimentologies, correct interpretation may still be problematic. Consequently, there is potential for further research to establish whether other

criteria may be more successful. In recent years an increasing number of workers have applied thin section analyses and microfabric techniques to the study of slope deposits (e.g. Derbyshire and Owen, 1990; Bertran, 1993; Elliott, 1996; Harrison, 1996; Harris, 1998). In this context, identification of micromorphologic characteristics diagnostic of paraglacially-reworked drift deposits and unmodified drift offers a promising avenue for further research, and may also have wider process and stratigraphic implications relating to paraglacial reworking.

Hypothesis testing based on the relationships between gully density and possible controls on the extent of paraglacial drift modification during this research has focused on contrasting gully density at a range of field sites. This has allowed the key physical constraints that govern paraglacial modification of steep drift slopes to be identified with greater confidence than hitherto. Further detailed analysis of within-site and between-site variation in paraglacial activity therefore represents a fruitful area for further research. In particular, investigation of the geotechnical properties of vulnerable, valley-side drift accumulations may shed further light on the conditions required for failure of drift, and help explain why different valley-side slopes have responded in different ways and at different times to paraglacial reworking.

The findings reported in this thesis concerning the history of delayed or renewed erosion and reworking of drift deposits in the Scottish Highlands suggest that ancient glacial drift slopes are highly sensitive to climatically-induced instability, possibly resulting from changing climate. However, the results relating to the timing and causes of delayed or renewed paraglacial modification of drift slopes in the Scottish Highlands are based on a limited number of reliable radiocarbon dates, and are thus provisional. More generally, there is a paucity of published information regarding the age, frequency and causes of phases of accelerated drift slope reworking in upland Britain. Future research on the

chronology and causes of debris flow activity on drift slopes in upland Britain would thus be a valuable contribution to the wider research surrounding the causes of accelerated Late Holocene geomorphic activity, and may ultimately yield a useful proxy record for reconstructing former climatic changes. In particular, additional timing data would help determine whether Mid-Late Holocene reworking of slopes exhibits random timing, or whether there is significant temporal clustering that may reflect general climatic deterioration, periodic increases in extreme events, or possible anthropogenic influences on slope stability. Lacustrine sediments from Loch Einich and other locations, for example, may yield a record of intermittent inwashing of minerogenic slopewash and coarse-clastic sediment from the adjacent steep drift slopes. Radiocarbon dating of organic-rich sediment intercalated with reworked deposits within such lake basins could yield a chronology of inwashing of debris, and thus of delayed or renewed paraglacial modification of glacigenic drift upslope. Furthermore, there is a need to identify and investigate sites where reworked drift containing intercalated organic horizons directly overlies *in situ* till relating to final deglaciation, to help determine whether recent episode(s) of reworking of drift slopes in upland Britain constitute delayed or renewed paraglacial activity. Meanwhile, this question remains unresolved.

Finally, the model of paraglacial slope modification and resedimentation presented in this thesis is based on data from a limited number of field sites within a fairly similar climatic environment, relief amplitude and tectonic setting. Further testing of this model in other deglaciated upland environments would be useful to establish its validity in providing morphological and sedimentological criteria of general applicability for the identification of both recent and ancient paraglacial activity. Moreover, the value of such a model would be significantly enhanced if it attempted to integrate the responses of both hillslope and fluvial systems to deglaciation.

9.4 Conclusion.

The term *paraglacial* was defined by Church and Ryder (1972, p. 3059) as referring to "nonglacial processes that are directly conditioned by glaciation", and encompassing both proglacial processes and those occurring around and within the limits of former glaciation. Subsequently, use of the term has broadened to incorporate numerous aspects of geomorphological activity that are in some way influenced by changes on the landscape following deglaciation (e.g. rock slope failure) or by inheritance of glacial landforms and deposits (e.g. fluvial sediment transport or development of coastal features in areas of thick glacigenic drift). The relevance of the paraglacial concept is particularly evident in the context of recent retreat of mountain glaciers, which has resulted in marked changes in many mountain geomorphic systems and radically disrupted many upland communities. In this thesis an attempt has been made to understand the implications of glacial inheritance on the postglacial development of drift-mantled hillslopes, irrespective of whether these changes occurred immediately after deglaciation (i.e paraglacial *sensu* Church and Ryder (1989)) or following a prolonged period of relative stability. The reworking of glacigenic sediment is thus considered the defining criterion of paraglacial activity, irrespective of the time elapsed since deglaciation. This work has demonstrated the usefulness of the paraglacial concept as an explicit geomorphic response within deglaciated mountain environments which helps to explain landscape evolution over a wide range of temporal and spatial scales. In particular, this research has addressed hitherto unresolved questions concerning the characteristics, causes and consequences of paraglacial activity on steep, drift-mantled slopes. Whilst there is inevitably scope for further development of our understanding of the mechanics and significance of paraglacial drift reworking and landscape modification in other settings, the results outlined above have demonstrated the importance of the concept for the

interpretation of formerly-glacierized mountain landsystems in passive continental margins.

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