Reconstructing the basal thermal regime of an ice stream in a landscape of selective linear erosion: Glen Avon, Cairngorm Mountains, Scotland

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The Cairngorm Mountain area of Scotland is a classic example of a landscape of selective linear glacial erosion, with sharp contrasts in the intensity of glacial erosion between the deeply incised troughs and valleys and the undulating high plateau. This article examines the Quaternary development of Glen Avon, a 200 m deep glacial trough set within the high plateau of the mountains. Evidence concerning the aggregate basal thermal regimes of the topographically controlled ice streams that formerly developed in this area is reconstructed from the geomorphological record, including bedforms indicative of wet-based, sliding ice and of dry-based ice frozen to its bed. This mapping indicates that basal sliding was not confined exclusively to the troughs but extended towards valley heads and on to parts of the plateau adjacent to troughs. The extent of basal sliding appears to have been greatest beneath pre-Late Devensian ice sheets. Basal ice temperatures are modelled under steady-state conditions for the last ice sheet at *c.* 18 ka BP. Basal thermal regimes are predicted using a reconstruction of the preglacial relief and for the current topography of the area. Convergent flow of ice through the preglacial valley system appears to have been sufficient to induce basal melting and therefore to initiate valley deepening. This effect is enhanced when the model is run across the present topography. Comparison of results of the geomorphological mapping and the modelling reveals significant differences between the actual and predicted extent of basal sliding outside the main ice stream. The overall conclusion is that many ice streams in mountainous terrain are inherited from the locations of preglacial valleys, which serve to accelerate ice flow and promote frictional heating beneath ice sheets.

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Ice streams are regions in a grounded ice sheet that flow much faster than the surrounding ice. These features have traditionally been subdivided into two types: ‘topographic’ ice streams, which lie in bedrock troughs and whose existence is largely controlled by the underlying bedrock topography, and ‘pure’ ice streams that do not lie in bedrock troughs (Stokes & Clark 1999). Pure ice streams can develop over areas of deformable sediment (e.g. Marshall *et al.* 1996) or due to the self-organization of ice flow as the temperature field in an ice sheet evolves through time (Payne & Donglemans 1997; Payne & Baldwin 1999). Most recent research has concentrated on establishing the dynamics of pure ice streams in the modern glacial environment (Blankenship *et al.* 1986; Anandakrishnan *et al.* 1998; Bell *et al.* 1998), and on the reconstruction of palaeo-ice streams (Patterson 1998; Knight *et al.* 1999; Stokes & Clark 1999, 2001). In contrast, there have been relatively few studies of the development and glaciological significance of topographically controlled palaeo-ice streams (Evans 1996; Lian & Hicock 2000).

Topographically controlled ice streams are especially important in mountainous terrain, where they control to a great extent basal thermal conditions and therefore the preservation or erosion of relict glacial and non-glacial landforms and landscapes (Kleman 1992; Kleman & Borgström 1994). Glaciologically, the continuum between protection and erosion is a function of the extent of former frozen-bed (protective) and thawed-bed (erosive) conditions beneath former ice sheets. The controls on the locations of areas of frozen-bed conditions within large ice sheets are strongly scale-dependent (Kleman *et al.* 1999). At the ice sheet scale (10^3 km), the location of frozen-bed areas is a function of dispersal centre location. Ice divides are characterized by low ice velocities and are commonly areas of high bed elevation and therefore limited ice thickness. At the mesoscale (10^2 km) the pattern of basal thermal regime is controlled by headward ice-stream erosion. Topographic control is limited, with the locations of frozen-bed and thawed-bed areas determined principally by the flow patterns that develop within ice sheets at this scale (Payne & Baldwin 1999). Topographic control becomes important at the kilometre scale, especially within mountainous or dissected terrain, where topographically controlled ice streams may develop, creating sharp contrasts in basal thermal regime over small distances (Glasser 1995).

The location of zones of frozen-bed and thawed-bed conditions beneath former mid-latitude ice sheets in North America and Scandinavia are now reasonably well known at the 10^2–10^3 km scale (Sugden 1977; Dyke 1993; Kleman & Stroeven 1997; Kleman &

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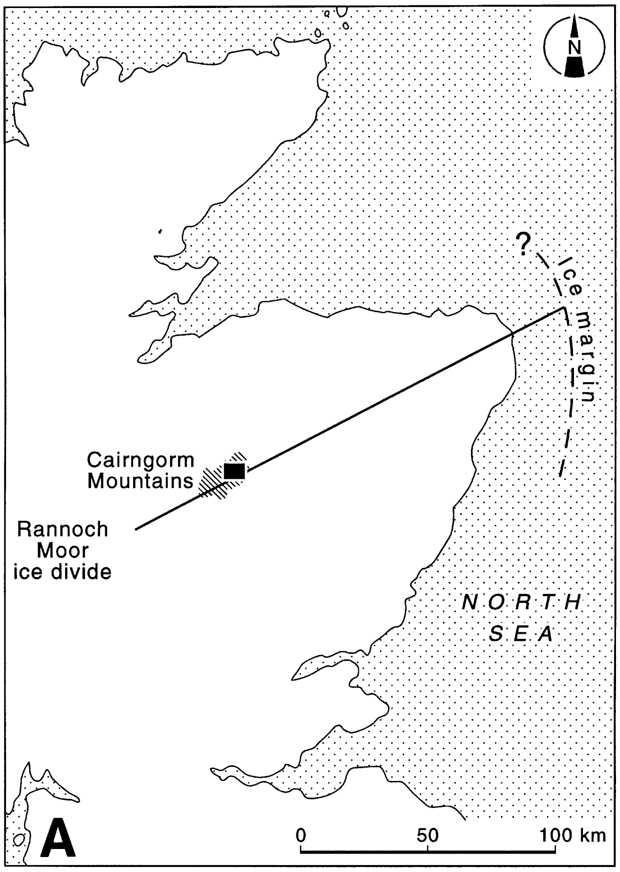
Hättestrand 1999; Kleman *et al.* 1997). However, much less is known about variations in ice-sheet thermal regime at the kilometre scale. Locating frozen-bed and thawed-bed areas beneath former ice sheets is important because of the implications for understanding flow within large ice sheets (Stokes & Clark 2001). The aims of this paper are (1) to examine in detail the Quaternary development of upper Glen Avon, a 200 m deep glacial trough set within the high plateau of the Cairngorm Mountains, Scotland (Figs 1, 2), and (2) to consider the implications for ice-stream development in mountainous areas. The basal thermal regime of the glaciers that formerly covered the study area is reconstructed from geomorphological evidence provided by the mapping of bedforms indicative of wet-based, sliding ice and of dry-based ice frozen to its bed. Basal ice temperatures are then modelled for ice sheets covering a reconstruction of the preglacial relief and for the current topography of the area.



Glacial history of the Cairngorm Mountains

The Cairngorm Mountains (Fig. 1) include the largest area of ground above 1000 m in Britain and have a long and complex history of glaciation (Glasser 1996). Although isolated cirque (corrie) glaciers may have formed at intervals in the Cairngorm Mountains in the period 2.4–0.75 Ma (Clapperton 1997), the first ice sheet to pass beyond the present east coast of Scotland is believed to have developed at around 0.8–0.75 Ma (Gatliff *et al.* 1994). It is probable that this expansion represents the first period of complete ice cover in the Cairngorms. Sediment volumes in the North Sea basin off eastern Scotland indicate that erosion rates were at a maximum in the Middle Quaternary, when ice sheets probably reached their greatest extent and thickness (Glasser & Hall 1997). By reference to the GRIP Summit δ^{18}O record, Clapperton (1997) concluded that temperatures during the last glacial cycle were low enough for the mountains of the Scottish Highlands to support glaciers for much of the last 75 ka. Bond-type cooling cycles of 7–11 ka duration probably produced extensive mountain ice caps in areas like the Cairngorms. Full ice-sheet conditions were probably established from 70–57 ka (Clapperton 1997) and till units of this age have been provisionally identified in northeast Scotland (Hall *et al.* 1995).

The maximum extent of ice in the last glacial cycle probably occurred in the Late Devensian at around 22 ka BP, followed by a withdrawal and a second advance at around 18 ka BP (Sejrup *et al.* 1994). A major ice limit is represented by the Wee Bankie moraine, *c.* 40 km east of Aberdeen (Sutherland 1984), but it is unclear if this represents the maximum extent of Scottish ice in the North Sea in the Late Devensian. The presence of metamorphic erratics at elevations of up to 800 m on the flanks of the northern Cairngorms indicates incursion of ice from Strathspey (Sugden 1970), but it is unclear if this relates to the 18 ka event or to some later advance (Brazier *et al.* 1996). Ice-marginal landforms and deposits at 400–760 m provide evidence of oscillations of the Strathspey ice front during overall ice retreat (Hinxman & Anderson 1915; Brazier *et al.* 1998). It is unclear if complete deglaciation occurred in the Cairngorms during the Lateglacial Interstadial (Gordon 1993) before small corrie and valley glaciers reformed or advanced during the Younger Dryas or Loch Lomond Stadial (Bennett 1996). Several such glaciers, including a small valley glacier at the head of Glen Avon, formed in the study area at this time (Sissons 1979) (Fig. 2).

The absence of schist erratics shows that no external ice from Strathspey penetrated the study area (Bremner 1929; Sissons 1976). The general pattern of ice flow across the plateau was parallel to upper Glen Avon and towards the NE (Bremner 1929) (Fig. 2). Ice that accumulated on Ben Macdui discharged E and NE into the Glen Avon trough and E and SE into Coire Lochain Uaine. The latter ice stream then split, with flow northwards via the Lairig an Laoigh breach and southwards down Glen Derry and SE across the southern part of the Moine Bhealaidh. The general parallelism of flow between valleys and the adjacent

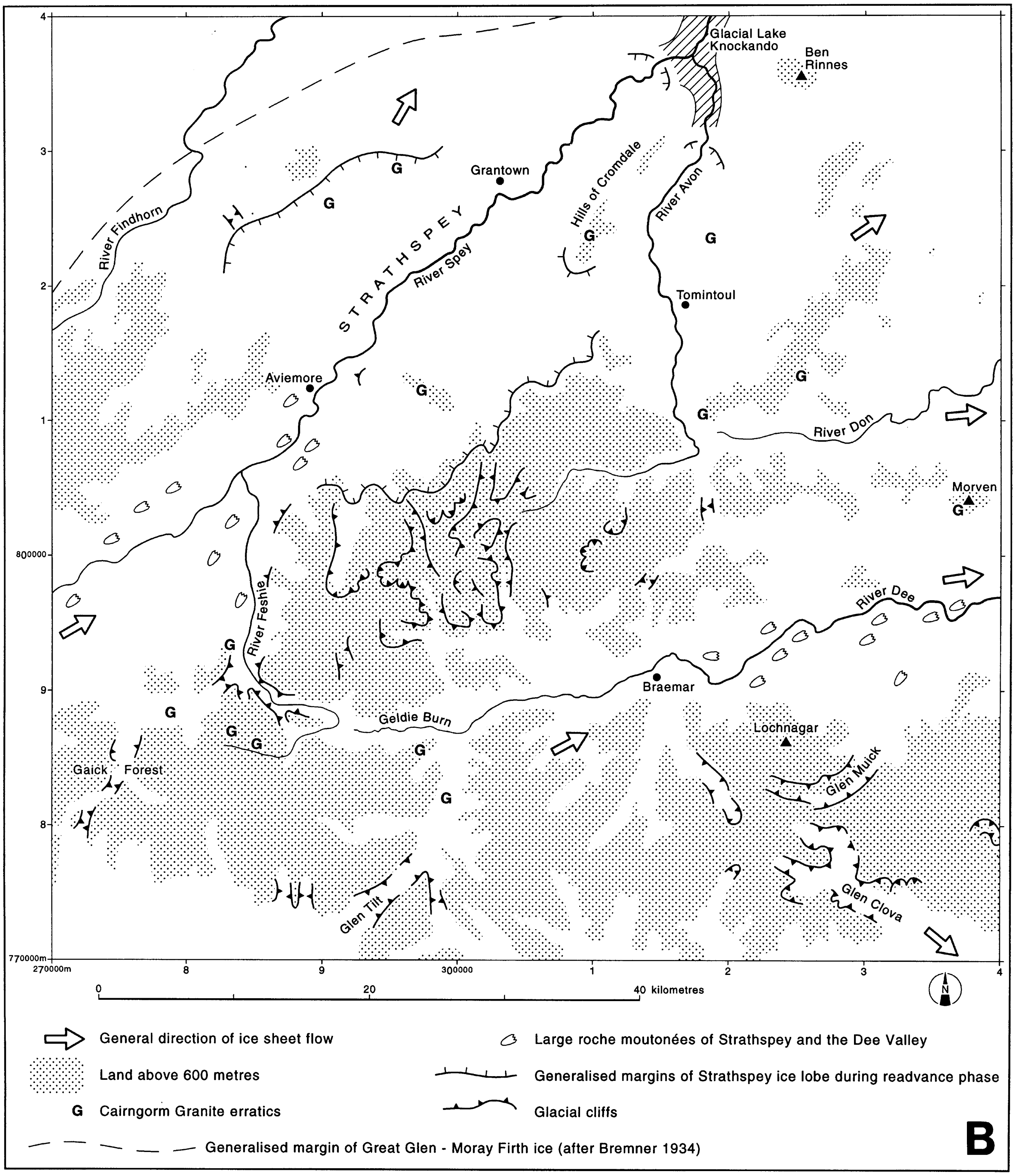


Fig. 1. Location maps of the study area. A (previous page). Map of Scotland showing the location of the field study area and its relationship to the location of the Late Devensian ice divide over Rannoch Moor and a Late Devensian ice margin at the Bosies Bank–Wee Bankie Moraine in the North Sea. The hatched area denotes the Cairngorm Mountain range, while the solid box indicates the area covered by Figs 2, 4, 5, 7, 9, 10. The solid line indicates the flowband within the ice sheet that is used in the reconstruction of ice sheet basal thermal regime in the Cairngorm Mountains. B. Upper Glen Avon and adjacent areas of the central Cairngorm Mountains.

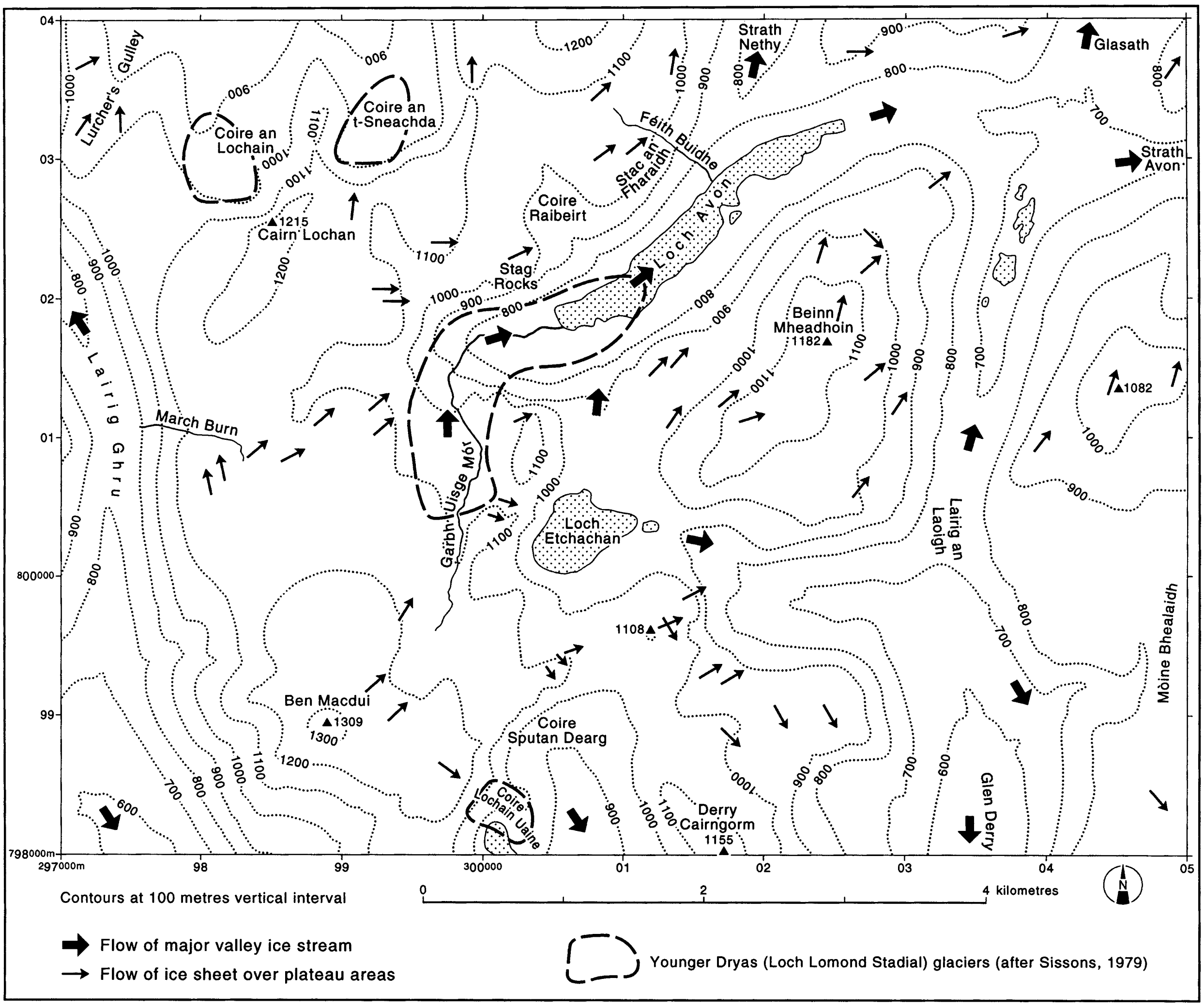


Fig. 2. The central Cairngorm Mountains, indicating patterns of ice flow and localities referred to in the text. All contour lines and spot heights in metres. Numbers around the outside of the frame indicate UK National Grid coordinates.

plateau indicates that successive ice streams that developed across the area were topographically controlled. Significant divergence of flow directions is apparently confined to the area north of Derry Cairngorm, where lee-side plucking indicates flow to the NE and to the SE. This divergence may reflect changing patterns of ice flow during ice cap build-up and decay.

At its maximum thickness the ice sheet covered the entire study area. Evidence for this statement includes the lee-side joint block removal and stripping of blockfields at elevations of up to 1200 m on Ben Macdui, as well as glacially transported tor blocks and stripped blockfields close to the summits of Beinn Mheadhoin (1182 m) and Derry Cairngorm (1155 m) (Figs 3A–F, 4, 5).

Geomorphological evidence of former glacier basal thermal regime

The granite mountains of the Cairngorms represent a classic example of a landscape of selective linear glacial erosion (Sugden 1968; Rea 1998). Within small areas there is a sharp contrast in the intensity of glacial erosion between the deeply incised troughs and valleys and the undulating high plateau. This contrast has been attributed to the fundamental control exercised by the basal thermal regime of successive ice sheets and glaciers during the Quaternary (Sugden 1968; Gordon 1993). Convergent flow of ice along pre-existing valleys and over cols is thought to have raised basal ice temperatures to values above the pressure melting point, promoting basal sliding and enhancing rates of glacial erosion. On plateau areas, however, ice was relatively thin and is believed to have remained cold-

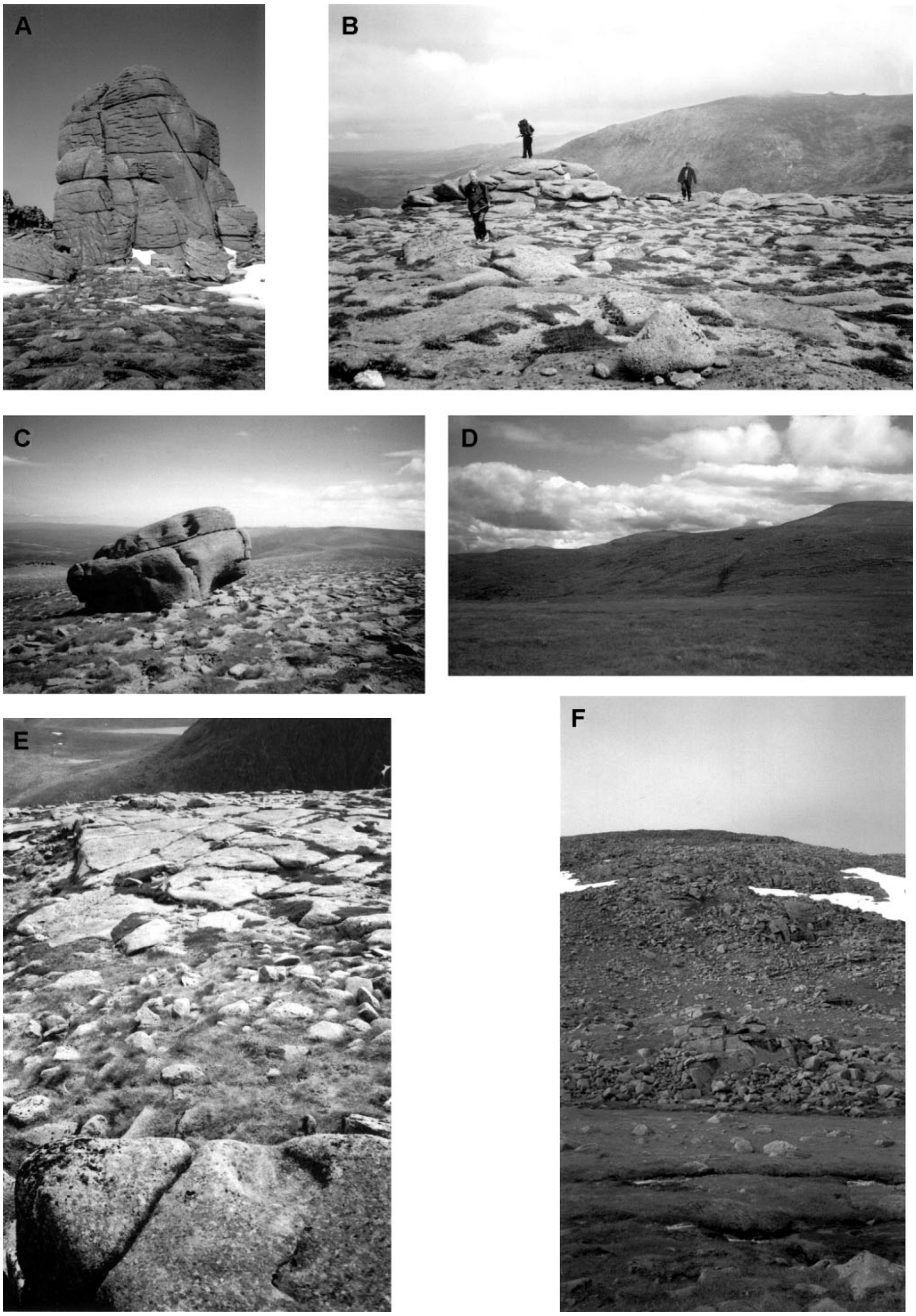


Fig. 3. Landforms indicating progressive modification of the Cairngorm landscape by glacier ice. A. Zone 4: Summit tors and regolith on the Beinn Mheadhoin ridge (NJ023016). B. Zone 4: Summit tor, with few signs of glacial disturbance (NJ024017). C. Zone 4: Glacially transported tor block on Beinn Mheadhoin (NJ024018). Note inclined weathering pits and basal notch. D. Zones 3-1: Panorama showing the transition between ice-moulding and limited erosion (NH985010). E. Zone 2: Tor stump at the former St. Valery refuge (NJ002023). The foreground shows the tor root, together with weathering pit, and stripped granite surfaces behind. F. Zone 1: Ice-moulding of the valley floor of the Garbh Uisge Mhor at 1100 m (NH997001).

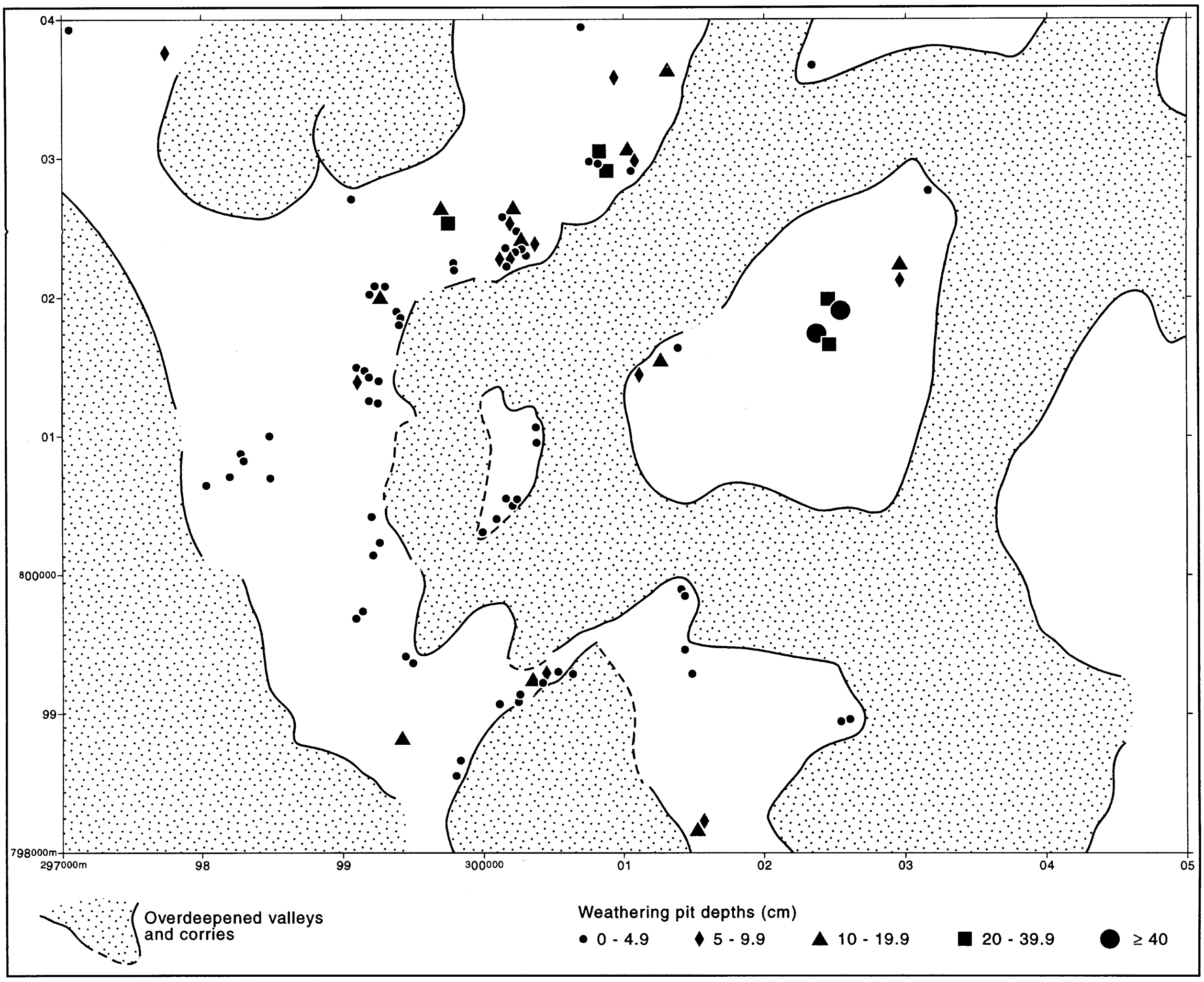


Fig. 4. Location and depths of weathering pits on plateau areas in the central Cairngorm Mountains.

based. This dry-based ice is inferred to have moved predominantly by internal deformation, rather than by sliding. As a result, restricted erosion on the plateau allowed the preservation of a preglacial relief, including delicate features such as tors and chemically weathered rock (Hall 1996).

This conceptual model of landscape evolution has widespread application to other landscapes of selective linear glacial erosion. In microcosm, the Cairngorm landscape is directly comparable to more extensive landscapes of selective linear erosion, including those from Baffin Island (Sugden & Watts 1977), the Torngat Mountains of Labrador (Ives 1958), East Greenland (Sugden 1974) and the Finger Lakes region of eastern North America (Clayton 1974). Selective erosion of plateaux by ice streams flowing within deep glacial troughs is also recognized as an important component of long-term Antarctic landscape evolution (Naslund 2001).

There is no doubt that glacial erosion in the Cairngorms has been selective. A fundamental contrast exists between the glacially over-deepened valleys and the rolling plateaux into which they are cut. The Glen Avon trough and the Lairig an Laoigh breach represent 150–200 m of Quaternary erosion beneath adjacent preglacial valley benches (Glasser 1996). The presence within these valleys of ice-moulded slabs and plucked rock steps indicates that the ice within them was sliding at its base and highly erosive.

The Cairngorm plateau surfaces have long been regarded as little-modified preglacial remnants and it seems likely that the depth of glacial erosion on the plateau is an order of magnitude less than that in the valleys (Sugden 1968; Ballantyne 1994; Hall 1996). Yet there are also widespread signs of glacial modification of bedrock surfaces on parts of the plateau. In particular, the morphology of the granite tors offers important evidence of contrasting degrees of glacial modification.

Tors occur widely on the Cairngorm plateau. All tors have formed in response to long-term differential weathering of the unevenly jointed granite (Ballantyne 1994). The larger tors may have originated under more temperate climates before the Quaternary but small tors have probably developed under periglacial environments during the Quaternary when parts of the Cairngorm plateau were ice-free (Hall 1996). Exposure ages derived from cosmogenic isotope analysis indicate that the summit surfaces of the tors predate at least the last ice sheet (W. Phillips, pers. comm.). The distribution of tors reflects variations in granite lithology (Hall 1996) and in the intensity of glacial erosion across the Cairngorms (Ballantyne 1994). In the western Cairngorms, tors are less frequent and tend to be more subdued forms. In the eastern Cairngorms, and especially on Ben Avon, tor groups occur wherever large joint-bounded blocks occur within the granite. This latter situation mirrors that on Dartmoor, SW England (Ehlen 1991), a granite terrain that experienced intense periglacial activity during the Quaternary but which lay beyond the limit of ice-sheet glaciation. On tor-forming lithologies, differential frost weathering and the removal of surrounding regolith by wash, solifluction and wind action has brought to the Earth's surface areas of massive granite wherever they occur (Palmer & Nielsen 1962).

The absence of tors from areas of massively jointed granite on the plateau above Glen Avon is therefore a reflection of significant glacial erosion (cf. Linton 1952). It requires that ice has removed upstanding blocks to leave plinths or slabs. Such surfaces occur on the lower parts of spurs on the northern rim of Glen Avon. For example, the former St. Valery Refuge was located adjacent to a sole remaining tor block, surrounded by massively jointed slabs (Fig. 3E). It is significant that the higher parts of these spurs carry modified tors that lack summit blocks but which retain upstanding basal blocks, implying a lesser degree of glacial modification (Fig. 3A, B). Displaced tor blocks occasionally can be found nearby, as on Beinn Mheadhoin, where a tor block has been carried 250 m north by ice from the summit tor (Fig. 3C). Only the summit tors of Beinn Mheadhoin (Fig. 3A) can be regarded as largely unmodified by the passage of ice across the study area. Examination of these and other little-modified tors on Ben Avon further east shows that they retain a fragile superstructure of loose blocks and are often deeply sculpted by weathering pits and lapies.

In addition to glacially modified tors, other landforms found on the plateau indicate former basal ice movement (Fig. 5):

a. *Roches moutonnées*. The largest examples occur above Loch Avon at Stac an Fharaidh (Fig. 6A) and show lee-side cliffs up to 20 m high. Areas of ice-moulded terrain, with lee-side plucking occur on the lower parts of the spur above Coire Sputan Dearg, east of The Saddle at the head of Strath Nethy and, most extensively, in the headwater basin of the Garbh Uisge Mhor, high on the slopes of Ben Macdui (Fig. 3F).
b. *Bedrock slabs*. Large, bare granite surfaces, defined by sheet joints and with only limited lee-side joint block removal occur at 1000–1100 m on the SW and NE spurs of Beinn Mheadhoin.
c. *Lee-side joint block removal*. Areas of subhorizontal, massively jointed granite are found on the lower plateau slopes on the north side of upper Glen Avon (Fig. 6B). Typically, the down-ice edge of the blocks ends in a 0.5–1 m high cliff, where the adjacent block has been removed. In a few locations, notably south of the Feith Buidhe, the removed blocks can be seen a few metres to tens of metres downslope.
d. *Glacially disturbed blockfields*. Few blockfields in the study area remain unaffected by the passage of glacier ice. Evidence of glacial disturbance is provided by the presence of perched blocks and by partial stripping of blocks to reveal jointed granite surfaces. The very limited development of blockfields in areas covered by active glaciers during the Loch Lomond Stadial and earlier in the Late Devensian, such as the valley of the Garbh Uisge Mhor, indicates that blockfield formation predates the last glaciation, as elsewhere in the Scottish Highlands (Ballantyne 1998).

The following features, together with little-modified tors, imply very limited erosion and an absence of basal ice movement:

a. *Undisturbed blockfields*. Here the granite structure is obscured by a full cover of blocks with horizontal surfaces that often show weathering pits. Large-scale sorted patterned ground may be evident.
b. *Sandy regolith*. Some areas are characterized by extensive areas of deep sandy regolith with few large blocks, for example on the Cairn Lochan ridge.
c. *Chemically weathered rock*. Granular regolith and subjacent weathered rock is exposed at a number of locations on the plateau (Fig. 5). Stream sections on the floor of Coire Raibert show an association between weathering and linear zones of quartz veining and hematite mineralization in the granite.
d. *Preglacial valleys*. A number of broad, shallow valleys occur on the plateau that show few signs of glacial modification (Fig. 5). Valley alignment is generally controlled by alteration zones within the granite. The valleys often terminate headward in nivation hollows.

The spatial distribution of these features reflects the basal ice conditions of the Quaternary glaciers that covered the study area (Fig. 7):

Zone 1: Ice that flowed into the main valleys was highly erosive. The extensive development of abraded and moulded surfaces on exposed bedrock indicates that the

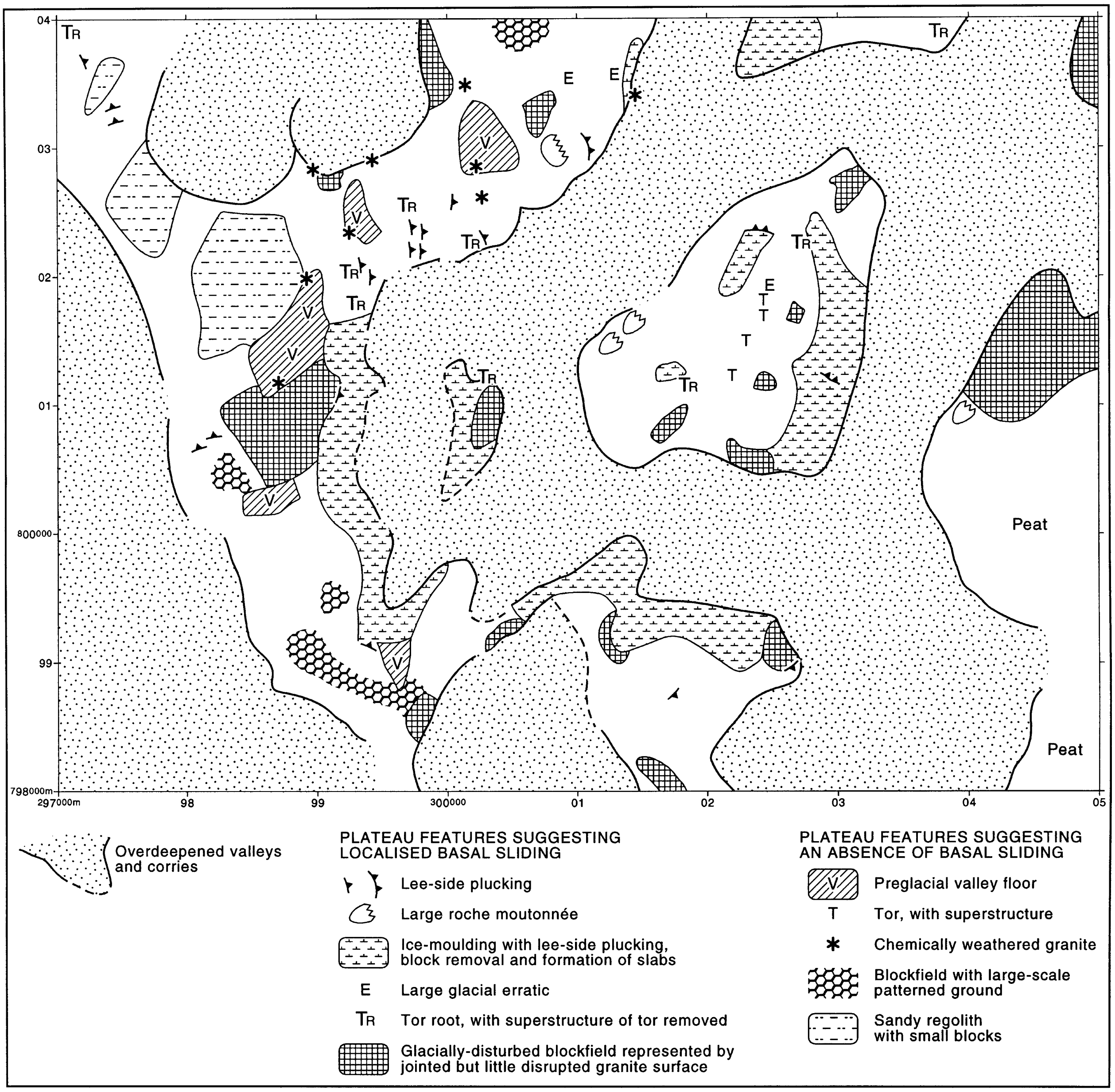


Fig. 5. Geomorphological map of the plateau areas of the central Cairngorm Mountains indicating glacier bedforms and their relation to former basal sliding, together with landforms indicating an absence of basal sliding.

basal ice was above the pressure melting point for extended periods.

Zone 2: Ice that covered the lower parts of the plateau, above the cliffs that define the edges of the main valleys, was periodically and locally sliding at its base. Here joint block removal occurred on the lee slopes of spurs lying across the line of flow. Pre-existing tor blocks on lower spurs were removed to leave slabs or, on the lower flanks of Beinn Mheadhoin, shaped into roches moutonnées. Erosion and hence sliding was concentrated on bedrock highs.

Zone 3: On the upper parts of the plateau, upstanding tors were reduced to plinths by block removal and blockfields were locally disturbed or removed by the passage of ice. There are few signs of glacial erosion on valley floors transverse to ice flow.

Zone 4: Only in a few of the highest parts of the plateau do tors and blockfields show no sign of glacial erosion. Even here there is local evidence of the passage of ice, with minor modification of tors and transport of large granite blocks.

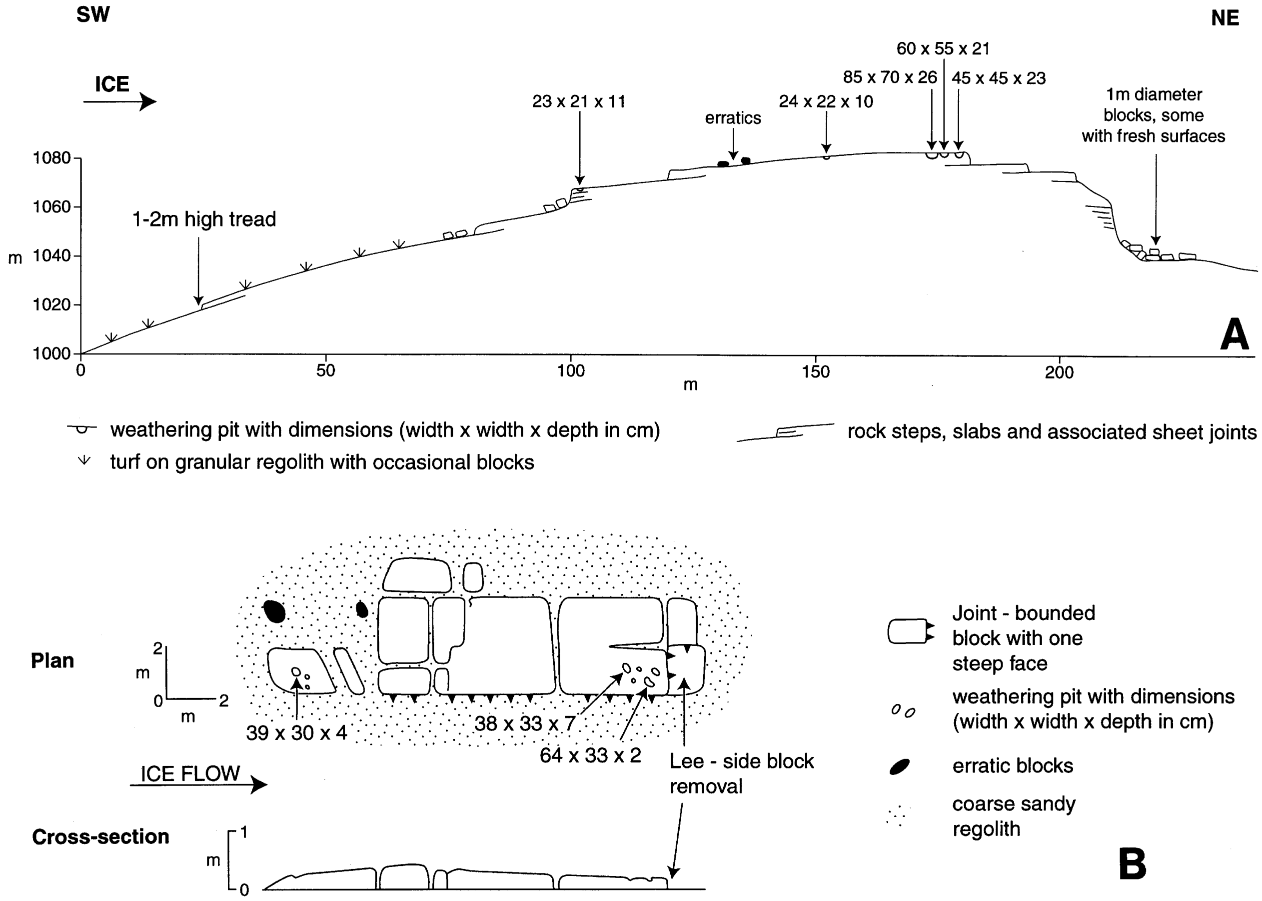


Fig. 6. Examples of glacier bedforms from the plateau above Glen Avon: (A) a large roche moutonnée at Stac an Fhairaidh and (B) glaciated slabs near Stag Rock.

The features indicative of basal sliding in Zones 1 and 2 above require that former glaciers were periodically wet-based. The features in Zones 3 and 4 suggest movement of ice at the glacier base but the amount of differential movement across the ice–rock boundary was small. Limited basal sliding may occur even in the basal layers of glaciers below the pressure melting point (Fitzsimons *et al.* 2000; Cuffey *et al.* 2000) and so it is unclear to what extent this limited modification relates to basal sliding or to internal deformation.

The distribution of weathering pits indicates that in Zones 2 and 3 significant glacial modification of the terrain was completed before the last ice sheet (Fig. 4). In the valley of the Feith Buidhe, above the trough head of Glen Avon, horizontal granite slabs show freshly abraded surfaces, with erratics of grey granite, and well-defined lee-side plucking. Only incipient weathering pits 1–4 cm deep have developed on these surfaces since they were last covered by ice in the Late Devensian. Horizontal surfaces on higher slopes on the north side of Glen Avon also show shallow weathering pits (Fig. 6B). By implication, weathering pit development since Late Devensian deglaciation has been limited.

Much deeper pits occur on the plateau on the surfaces of roche moutonées and glacially modified tors (Fig. 4). Examples include tor plinths and bases, where ice has removed upstanding tor blocks (Fig. 3E). On Stac an Fhairaidh, weathering pits reach depths of 20 cm. It is significant that they are found close to the summit of this large roche moutonnée, above the lee-side cliff, where abrasion would be most limited (Fig. 6A). Around the former St. Valery Refuge, surfaces with shallow pits occur that have been subsequently abraded by the passage of ice. These differences in weathering pit depths occur consistently over a wider area of the Cairngorms (W. Phillips, pers. comm.). Cosmogenic isotope analysis of summit surfaces with deep weathering pits on glacially modified tors, including the summit tor of Cairn Gorm, also indicate that these surfaces predate at least the last ice sheet (W. Phillips, pers. comm.).

In the study area, the deeper weathering pits developed on slabs and roches moutonnées lie within

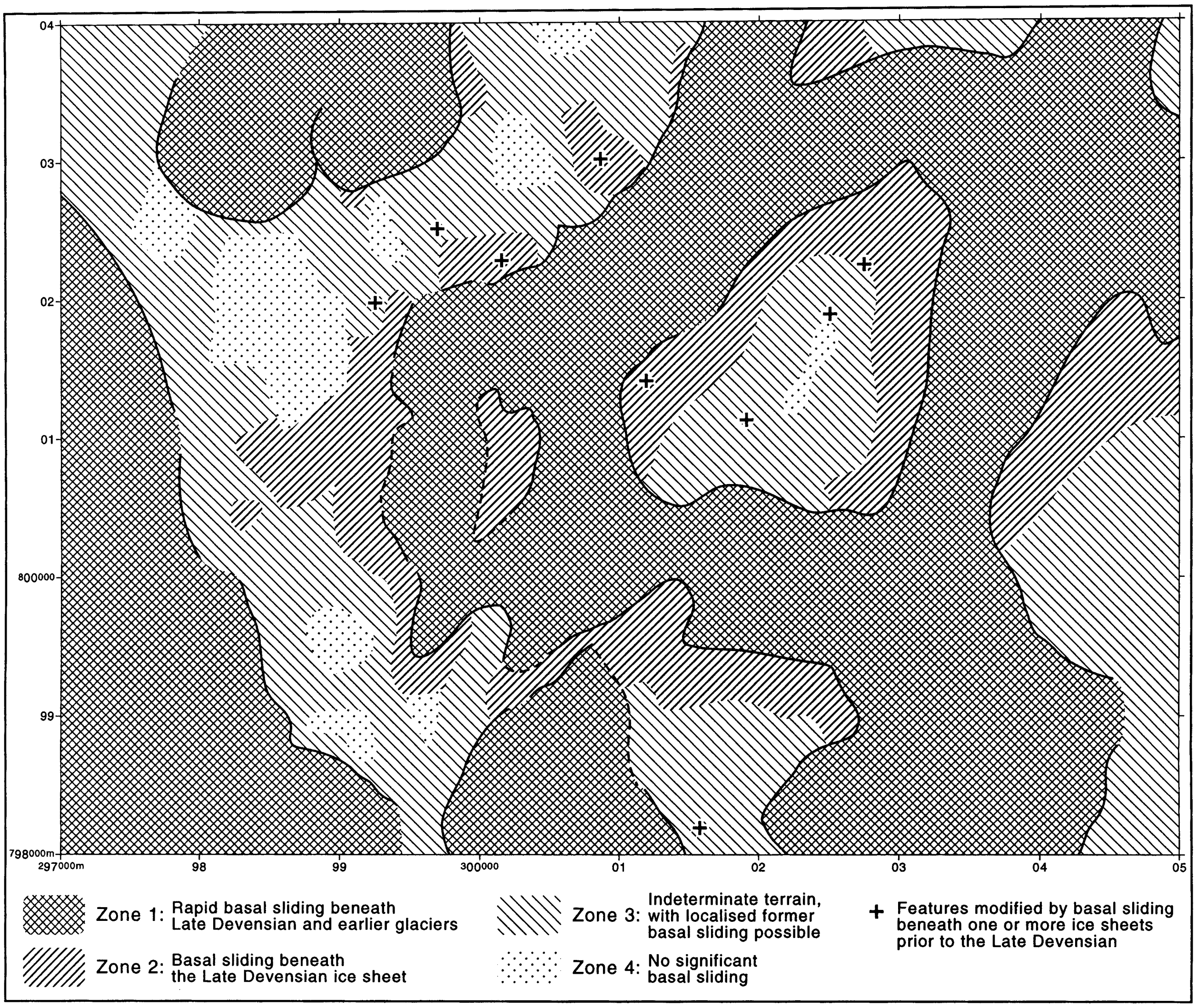


Fig. 7. Inferred patterns of basal sliding in the central Cairngorm Mountains. Modification of the landscape by glaciers rises from Zone 4 (no significant basal sliding) to Zone 1 (rapid basal sliding beneath Late Devensian and earlier glaciers).

Zones 2 and 3 (Figs 4, 7). By implication, abrasion beneath the last ice sheet was insufficient to remove pre-existing pits. The gross form of these features, with lee-side plucking, block removal and widespread abrasion, requires wet-based ice and predates the last ice sheet. Basal sliding beneath an earlier ice sheet or ice sheets was more extensive and reached higher elevations than during OI Stage 2. The mapping also indicates that basal sliding was not confined to the troughs but extended towards valley heads and on to parts of the plateau adjacent to troughs (Fig. 5).

Former basal thermal regime: a numerical model

Calculations of former basal thermal regime in the Cairngorm Mountains are based on the numerical model of Budd *et al.* (1971), in which a heat conduction equation is solved through a series of vertical columns of ice along a given flow line for an ice sheet in steady-state. We recognize that this steady-state model has limitations, in particular because it represents a 'snap-shot' of ice sheet basal thermal regime at a given time and is not time-integrated in its approach. However, the justification for the use of this simple type of model is that it can produce a field-testable output, whereas other more complicated thermomechanical solutions do not produce testable output, using schematic, circular ice sheets resting on flat bedrock at a grid resolution of 25 km^2 (e.g. Payne *et al.* 2000). Other flow-line models exist, but these are only capable of producing broad patterns of thermal regime, resolved for example at 50 km intervals (e.g. Wilch & Hughes 2000) and again are not directly testable against field evidence at the scale required in this study.

The basal ice temperature (T_b) in the model of Budd *et al.* (1971) is given by:

$$T_b = T_s - H\left[Y_b \frac{\mathrm{erf}(y)}{y} - Y_s 2E(y)\right] \quad (1)$$

$$y = \sqrt{\left(\frac{bH}{2k}\right)} \quad (2)$$

$$Y_b = \frac{\wedge g}{K} + \frac{\tau_b u}{K} \quad (3)$$

$$Y_s = \frac{u\alpha\lambda}{b} \quad (4)$$

where T_s is the surface temperature (°C), H is the ice thickness (m), Y_b is the total heat flux at the ice-sheet base (W m^{-2}), Y_s is the heat gradient at the ice-sheet surface (°C m^{-1}), $\wedge g$ is the geothermal heat flux (W m^{-2}) and erf(y) is the error function. The basal shear stress is τ_b (Pa, calculated from ρgHsinα), b is the surface accumulation (m yr^{-1}), α is the surface slope of the ice sheet (in radians), K is the thermal conductivity of ice (2.1 W m^{-1} k^{-1}), λ is the altitudinal air-temperature lapse rate (assumed to be held constant at 7°C/km, following Diamond 1960) and k is the thermal diffusivity of ice (36.29 m^2 yr^{-1}). The balance velocity u is given by:

$$u = \frac{b(x)x}{H} \quad (5)$$

where x is the distance from the ice divide and $b(x)$ is the mean accumulation over that distance. The function E(y) is given by:

$$E(y) = \int_0^y F(\xi)d\xi \quad (6)$$

where F(ξ) is a function known as the Dawson integral. Both the error function (erf(y)) and the Dawson integral are tabulated functions (Abramowitz & Stegun 1965). In addition, Rybicki (1989) provides a numerical recipe for the Dawson integral. Basal ice temperatures are calculated over a two-dimensional grid of 119 cells (17 × 7 data points representing an area of 8 × 6 km at a spacing of 0.5 × 1 km). The grid represents part of a 'flow band' of the last Scottish ice sheet so that ice flow along the long axis of the grid (8 km) approximates flow lines within the ice sheet (Fig. 1).

The model assumes five glaciological conditions. The first is that the ice sheet is in steady state. This means that volumes of ice accumulation and ablation match exactly, and the flux of ice within the ice sheet acts to maintain the surface profile. The second condition is that heat-conducted transverse to ice flow is negligible in comparison to that conducted along the flow line. Third, the advection of ice is assumed to occur in the vertical direction only. The advection rate is also assumed to be constant and equal to the rate of surface warming. The fourth condition is that all friction is assumed to apply at the base of the ice sheet. Finally, the thermal diffusivity of ice (k) and the strain rate of ice are assumed to be constants. The rationale behind these assumptions and the full derivation of the fixed column model are discussed in detail by Budd *et al.* (1971). The basic calculations have been tested against modern ice masses such as the Antarctic ice sheet and proved capable of predicting accurately measured basal ice temperatures to within a few °C (Budd *et al.* 1971). Calculations of former ice sheet thermal regime were made using the spreadsheet method of Glasser & Siegert (2002).

The Budd model has been used to reconstruct steady-state ice-sheet temperatures for the former Laurentide Ice Sheet (Sugden 1977) and former Scottish Ice Sheet (Gordon 1979). A criticism of the application of this model to former ice sheets is the assumption of uniform patterns of ice flow. This assumption is unlikely to hold in areas of complex basal topography, especially where convergent flow occurs in deep troughs. Subsequent applications of the model to a sector of a former Scottish ice sheet (Glasser 1995) and to predict areas of basal melting beneath the modern Antarctic ice sheet for comparison with the spatial distribution of subglacial lakes (Siegert & Glasser 1997) have recognized the importance of convergence and divergence of flow. These authors have therefore refined the model to account for the convergence and divergence of flow at the base of the ice sheet through troughs and basins by multiplying the balance velocity in areas of convergent flow by a Sine function related to position within a trough or basin (Fig. 8). The resulting 'convergent' velocity is then added to that representing the balance velocity.

The effects of increasing ice velocity in areas of convergent flow have been investigated by Siegert & Glasser (1997), who conducted sensitivity tests on this method. They concluded that this method satisfactorily accounts for convergence of flow in topographically complex areas. The modelling strategy adopted in this article is to run the model through an initial 'reference' state across the present-day topography (Fig. 9A), where the convergent flow of ice is not accounted for, before altering the velocity within the flowband in response to changes in velocity introduced by topography. The model is also run across a reconstructed preglacial topography (Fig. 10A) in order to gain insight into patterns of basal thermal regime beneath early Quaternary ice sheets in this area. The basal boundary condition is not changed when the ice reaches the melting point, so areas of basal melting are identified by positive temperatures.

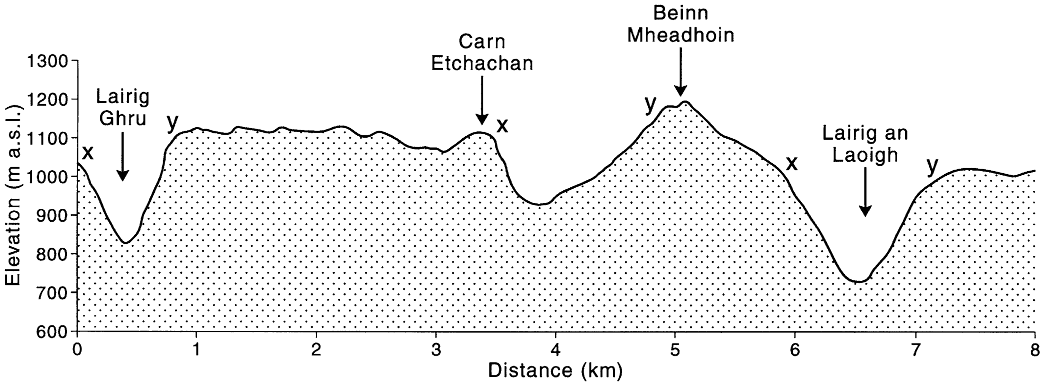


Fig. 8. Cross-section depicting the present-day topography from NGR 9701 to 0501. The markers x and y denote the lateral extent of the deep troughs in which convergent flow of ice is assumed to occur.

Derivation of input parameters

The inputs to the model are those chosen to represent a Scottish ice sheet at 18 ka BP. Full discussion of the input parameters are provided by Glasser (1991, 1995). These are the ice-sheet profile, the geographical extent of the ice sheet, the atmospheric temperature regime, altitudinal lapse rate and accumulation rate. The ice-sheet surface profile is a parabola generated by an assumed basal shear stress of 100 kPa. The margin of the ice sheet is set at the Bosies Bank moraine, 25 km offshore from the present-day North Sea coast of Scotland (Hall & Bent 1990). This moraine reflects the position of the ice margin at 18 ka and is used in preference to the Late Devensian maximum position at *c.* 22 ka when Scottish ice was probably convergent with Scandinavian ice. Annual average temperature at the margin is assumed to be −9°C (Coope 1975) with an accumulation rate of 0.75 ma^{-1} (see discussion in Hubbard 1999). Ice thickness is determined by subtracting the present-day topography from the calculated ice-sheet profile and corrected for isostatic adjustment using an ice/Earth mantle density conversion factor of 1.36. This produces an ice thickness of between 300 and 400 m over the highest summits in the area.

Sensitivity tests of the model

Sensitivity tests using the model have shown that it is more sensitive to changes in some input parameters than others (Glasser 1991). For example, it is very sensitive to changes in ice thickness and therefore to changes in the ice-surface profile used to depict the ice sheet. The general rule is that the thicker the ice sheet, the warmer the basal temperature. The model is less sensitive to changes in the accumulation rate. This is because under steady-state conditions higher accumulation rates lead to higher vertical velocities and therefore decrease the basal temperature. High accumulation rates nearer the margin produce higher basal temperatures as a result of increased velocity and increased frictional heating. Since there is a direct relationship between the ice surface temperature and the basal temperature, the model is sensitive to the mean annual temperature imposed at the margin. The warmer the imposed ice marginal temperature, the greater the extent of basal melting.

Model results

Present-day topography

The basal temperature distribution for the reference model run across the present-day topography (Fig. 9A) indicates that the area is dominated by basal freezing (Fig. 9B). Basal temperatures range from −11°C over the summit of Ben Macdui to −6°C in the deep troughs of Glen Avon and Lairig an Laoigh. Velocities predicted in the model are *c.* 5 ma^{-1}, and as a result frictional heat production is generally low. Basal ice temperatures are therefore primarily a function of the underlying topography, which determines ice thickness. This frozen-bed thermal regime is altered by the inclusion of the enhanced velocities in areas likely to experience convergent flow (Fig. 9C). This model run predicts extensive areas of basal melting in the deep troughs, although ice over the mountain summits and higher slopes remains frozen at the base. Velocity increases of between 4 and 6 ma^{-1} are required to initiate basal melting in areas previously predicted to be frozen-based in the reference models run.

Preglacial topography

The basal temperature distribution for the reference model run across a reconstructed preglacial topography (Fig. 10A) indicates that the area is again dominated by basal freezing (Fig. 10B). Basal temperatures range from −12°C over the plateau surrounding Ben Macdui

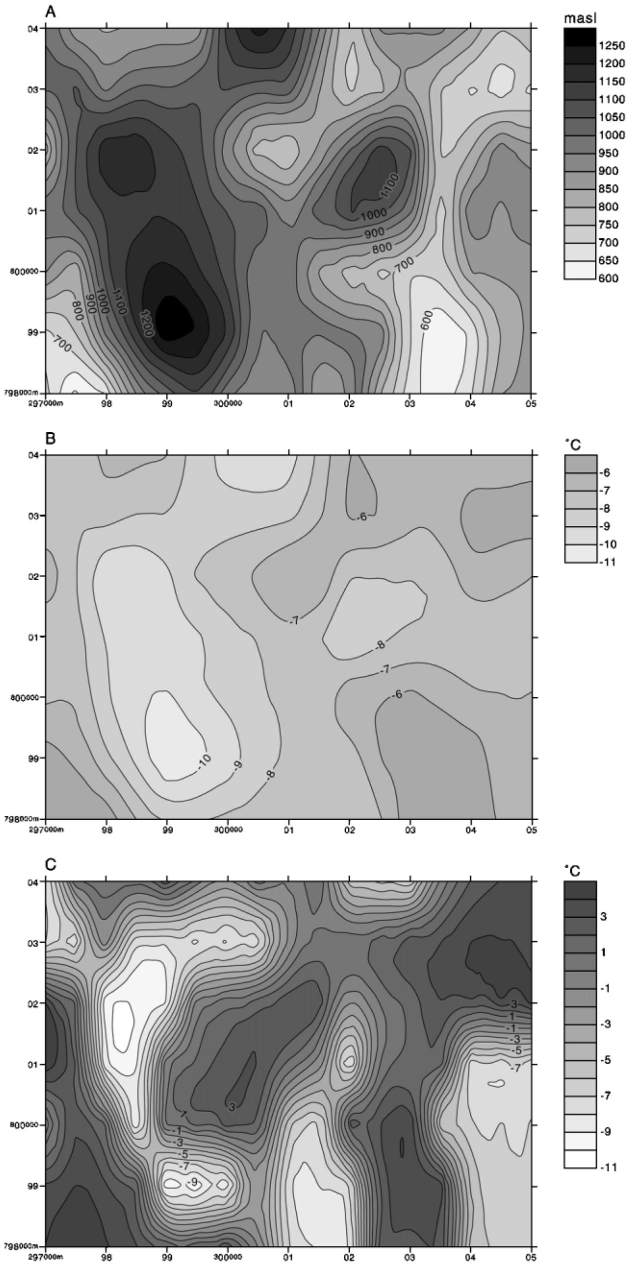


Fig. 9. Modelled basal ice temperatures in the study area: A. The basal topography based on present-day relief. B. Reference model run, indicating frozen-bed conditions. C. Model run including the effects of convergent flow, indicating areas of basal melting in troughs and other topographic basins.

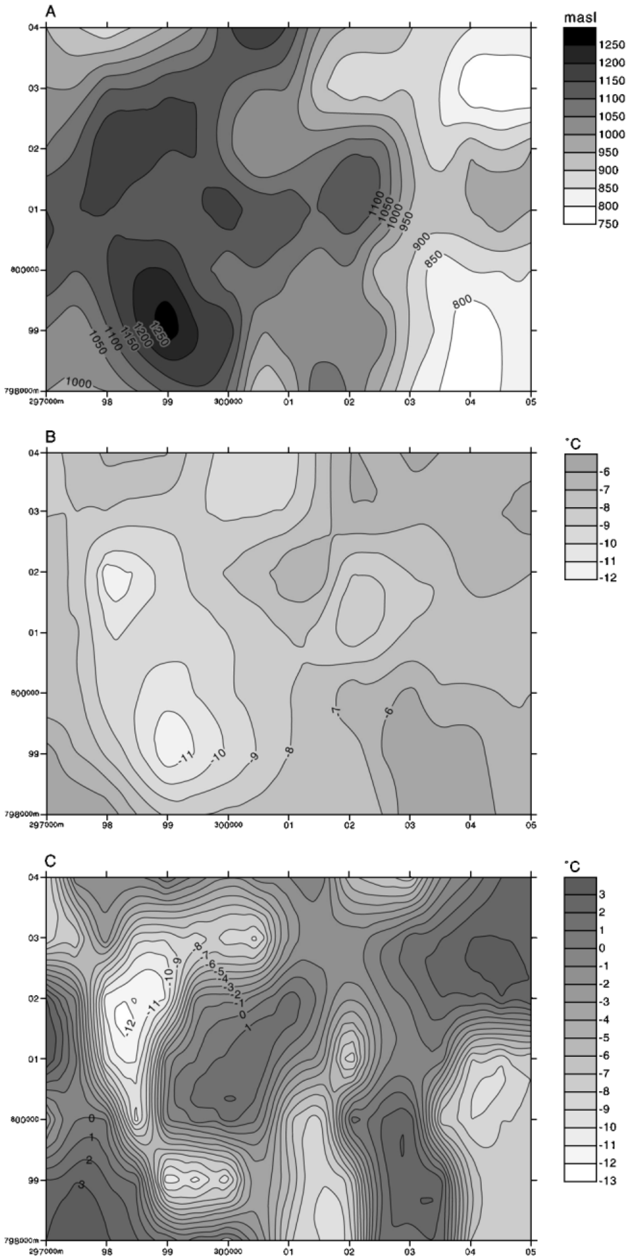


Fig. 10. Modelled basal ice temperatures in the study area: A. The basal topography based on a reconstruction of the preglacial relief. B. Reference model run, indicating frozen-bed conditions. C. Model run including the effects of convergent flow, indicating areas of basal melting in preglacial valleys.

to −6°C in the deep troughs of Glen Avon and Lairig an Laoigh. Velocities predicted in the model are again *c.* 5 $\mathrm{m\,a^{-1}}$, so that frictional heat production is still generally low. This frozen-bed thermal regime is altered by the inclusion of the enhanced velocities in areas likely to experience convergent flow (Fig. 10C). This model run predicts extensive areas of basal melting in the deep troughs, although ice over the mountain summits and higher slopes remains frozen at the base. The overall extent of basal melting is less than that predicted by the present-day topography (Figs 9C, 10C). This is largely because the preglacial valleys are less deep than those of the present day so that topographic contrasts are less pronounced.

Discussion

The reconstruction of the basal thermal regimes of the glaciers that covered the study area provides an aggregate picture of conditions through the Quaternary. Ice masses of different thickness have formed and decayed over the Cairngorms under variable and largely unknown climatic conditions. Nonetheless, the reconstruction reveals a zoned patchwork of former basal ice conditions:

Zone 1: periodically covered by wet-based ice, including during the Late Devensian
Zone 2: covered by wet-based ice in a period or periods prior to the Late Devensian
Zone 3: local development of wet-based ice is possible during a period or periods prior to the Late Devensian
Zone 4: dry-based ice throughout the Quaternary

The selective nature of glacial erosion in the study area can be compared to evidence from other glacial landscapes. For example, a patchwork of preserved non-glacial and glacial landforms has been identified on plateau surfaces in northwestern Sweden (Clarhäll & Kleman 1999), where this transition is interpreted as the result of the rates of change from dry- to wet-based conditions beneath ice during ice recession. McDougall (2001) has also identified a transition from cold-based ice on plateaux to ice-moulded bedrock, documenting wet-based erosive conditions, on plateau edges in the mountains of the English Lake District. Modification of preglacial bedrock structures by glacial erosion, including the modification of sheet structures into roches moutonnées, has been described in the Dee Valley to the east of the study area (Sugden *et al.* 1992) and in glaciated areas of Sweden (Lindström 1988; Olvmo *et al.* 1999) and the USA (Patterson & Boerboom 1999).

Comparison of the results of field mapping with the model predictions indicates that there is a good fit for Zone 1 (Figs 7, 9C). The deep glacial valleys of Glen Avon, the Lairig Ghru and the Lairig an Laoigh correspond with the location of the predicted zones of basal melting beneath ice sheets modelled on both the existing terrain and the preglacial terrain. The model helps to explain the overdeepening of these valleys from their fluvial precursors. The preglacial valleys, when filled with ice, were of sufficient depth to induce basal melting through a combination of greater ice thickness and the funnel effect of their geometry to accelerate ice flow and promote frictional heating (Fig. 10C). This effect is enhanced as the valleys are deepened by glacial erosion.

The model adds to a growing body of literature that points to the importance of basal thermal regime in the development of ice streams, especially in topographically complex terrain. First is the observation that modern Antarctic ice streams, such as Ice Stream D, follow deep bed channels that may have an influence on their subglacial thermal regime (Bindschadler *et al.* 2000). Thus the onset of Ice Stream D occurs downstream of an area where there is an increase in longitudinal strain rate after the ice has passed the centre of an overdeepening in the bed channel. This relationship has also been demonstrated mathematically by Hindmarsh (2001), who demonstrated that flow-parallel channelling generates large quantities of heat within bedrock lows (i.e. valleys) beneath an ice sheet. Second is the acquisition of radar data, which indicates that the heads of many modern Antarctic ice streams are areas where large quantities of water are stored in subglacial lakes (Siegert & Bamber 2000). These authors suggest that basal melting is an important factor in determining the onset location of ice streams. Third is the work of Bourgeois *et al.* (2000), who identified that ice streams draining the Pleistocene ice sheets in Iceland were located over areas of high geothermal anomalies, indicating that ice stream activity was due to bed lubrication by meltwater generated in areas of high geothermal heat flux. Finally, there is geomorphological and sedimentological evidence, including large-scale ice-flow indicators, which suggests that the interior of the Cordilleran Ice Sheet consisted of local accumulation areas that were drained by rapidly flowing glaciers, interpreted as topographically controlled ice streams (Lian & Hicock 2000).

The modelled basal thermal regime does not match closely the geomorphological evidence in Zones 2–3. The model generally overestimates the overall extent of dry-based ice on the plateau in areas where field mapping indicates that only small parts of the study area (Zone 4) have not been modified to some extent by the passage of ice (Figs 5, 7). Zones 2 and 3 generally lie outside, although adjacent to, the zones of basal sliding predicted by the model. There are a number of possible reasons for this mismatch:

1. The sample grid of the model (0.5×1 km) is too coarse to identify the local effects of topography on basal thermal regime. For example, the spurs that run perpendicular to ice flow north of Loch Avon have lost blockfields and tors from their lower parts and have been modified by lee-side block removal. The intervening valleys, Coire Domhain and Coire Raibert, show few signs of glacial modification and retain mantles of chemically weathered granite. Here it seems that localized basal sliding has been induced where ice has been forced to flow uphill over the spurs.
2. Edge effects and interaction with other components of ice flow within the ice sheets are ignored. This sector of the former Cairngorm ice cap was not isolated from adjacent areas, including the Lairig Ghru and Strathspey. It is possible that these other components of ice flow added or subtracted to the mass budget of ice passing over the study area.
3. The model parameters were different for ice sheets

prior to the Late Quaternary. In another article, we have suggested that more extensive, thicker and more erosive ice sheets covered Scotland during the Middle Quaternary (Glasser & Hall 1997). It is probable that the valley glaciers and smaller mountain ice fields that existed in the Cairngorms in the Early and Middle Quaternary developed under different climatic conditions to those modelled here. Thus the patterns of frozen-bed and thawed-bed conditions are spatially and temporally variable during earlier glaciations.

In addition, this study has examined a plateau area incised by only one major valley. To fully explain the patterns of ice flow at this local scale there is clearly a need for further modelling studies incorporating a more sophisticated treatment of former ice dynamics and basal thermal regime. In particular there is a need for modelling studies of the evolution of ice-sheet thermal regime over a full glacial cycle, with which to compare the time-integrated geomorphological record.

Conclusions

The main conclusions of this study are as follows:

- Geomorphological mapping indicates that basal sliding beneath former ice sheets in this part of the Cairngorm Mountains was not confined exclusively to ice streams in the troughs, but that basal melting also extended towards valley heads and on to parts of the adjacent plateau. The modification of weathering pits developed on granite surfaces in the study area indicates that the extent of basal sliding was greater beneath pre-Late Devensian ice sheets.
- Numerical modelling of ice-sheet thermal regime under steady-state conditions suggests that the location of preglacial valleys is a strong influence on ice-sheet thermal regime. These valleys, when filled with ice, were of sufficient depth to induce basal melting through a combination of greater ice thickness and the funnel effect of their geometry. We propose that many topographically controlled ice streams may inherit their location from preglacial valleys, since these valleys serve to accelerate ice flow and promote frictional heating.
- Convergent flow of ice streams through preglacial valley systems in this part of the Cairngorm Mountains appears to be sufficient to induce basal melting and to initiate valley deepening beneath ice sheets. Thus a promising avenue for further research is to examine the relationships between valley forms and their control on patterns of ice flow at this scale.
- There is clearly a need for time-integrated numerical modelling studies that incorporate a more sophisticated treatment of former ice dynamics and basal thermal regime, and in particular for time-dependent models of basal thermal regime that can be tested against the time-integrated geomorphological record. This is especially the case at the kilometre scale investigated here, where topography is a strong influence on patterns and rates of ice flow.

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References

Abramowitz, M. & Stegun, I. A. 1965: *Handbook of Mathematical Functions with Formulas, Graphs and Mathematical Tables*. 1046 pp. Dover Publications, New York.

Anandakrishnan, S., Blankenship, D. D., Alley, R. B. & Stoffa, P. L. 1998: Influence of subglacial geology on the position of a West Antarctic ice stream from seismic observations. *Nature 394*, 62–65.

Ballantyne, C. K. 1994: Scottish Landform Examples – 10: the tors of the Cairngorms. *Scottish Geographical Magazine 110*, 54–59.

Ballantyne, C. K. 1998: Age and significance of mountain-top detritus. *Permafrost and Periglacial Processes 9*, 327–345.

Bell, R. E., Blankenship, D. D., Finn, C. A., Morse, D. L., Scambos, T. A., Brozena, J. M. & Hodge, S. M. 1998: Influence of subglacial geology on the onset of a West Antarctic ice stream from aerogeophysical observations. *Nature 394*, 58–62.

Bennett, M. R. 1996: The Loch Lomond Readvance in the Cairngorm Mountains. *In* Glasser, N. F. & Bennett, M. R. (eds.): *The Quaternary of the Cairngorms*, 54–68. Quaternary Research Association, Cambridge.

Bindschadler, R., Chen, X. & Vornberger, P. 2000: The onset area of Ice Stream D, West Antarctica. *Journal of Glaciology 46*, 95–101.

Blankenship, D. D., Bentley, C. R., Rooney, S. T. & Alley, R. B. 1986: Seismic measurements reveal a saturated porous layer beneath an active Antarctic ice stream. *Nature 322*, 54–57.

Bourgeois, O., Dauteuil, O. & Van Vliet-Lanoë, B. 2000: Geothermal control on flow patterns in the last glacial maximum ice sheet of Iceland. *Earth Surface Processes and Landforms 25*, 59–76.

Brazier, V., Gordon, J. E., Kirkbride, M. P. & Sugden, D. E. 1996: The Late Devensian ice sheet and glaciers in the Cairngorm Mountains. *In* Glasser, N. F. & Bennett, M. R. (eds.): *The Quaternary of the Cairngorms*, 28–53. Quaternary Research Association, Cambridge.

Brazier, V., Kirkbride, M. P. & Gordon, J. E. 1998: Active ice-sheet deglaciation and ice-dammed lakes in the northern Cairngorm Mountains, Scotland. *Boreas 27*, 297–310.

Bremner, A. 1929: The glaciation of the Cairngorms. *The Deeside Field 4*, 29–37.

Budd, W. F., Jenssen, D. & Radok, U. 1971: *Derived Physical Characteristics of the Antarctic Ice Sheet*. ANARE Interim Reports, Ser A. Glaciology. Publication No. 120.

Clapperton, C. M. 1997: Greenland Ice Cores and North Atlantic Sediments: Implications for the Last Glaciation in Scotland. *In* Gordon, J. E. (ed.): *Reflections on the Ice Age in Scotland*, 45–58. Scottish Natural Heritage, Edinburgh.

Clarhäll, A. & Kleman, J. 1999: Distribution and glaciological implications of relict surfaces on the Ultevis plateau, northwestern Sweden. *Annals of Glaciology 28*, 202–208.

Clayton, K. M. 1974: Zones of glacial erosion. *Institute of British Geographers Special Publication 7*, 163–176.

Coope, G. R. 1975: Climate fluctuations in Northwest Europe since the last interglacial. *In* Wright, A. & Moseley, F. (eds.): *Ice Ages: Ancient and Modern*, 153–168. Seel House Press, Liverpool.

Cuffy, K. M., Conway, H., Gades, A. M., Hallet, B., Lorrain, R., Severinghaus, J. P., Steig, E. J., Vaughn, B. & White, J. W. C. 2000: Entrainment at cold glacier beds. *Geology 28*, 351–354.

Diamond, M. 1960: Air temperature and precipitation on the Greenland ice sheet. *Journal of Glaciology 3*, 558–567.

Dyke, A. 1993: Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography 17*, 223–247.

Ehlen, J. 1991: Significant geomorphic and petrographic relations with joint spacing in the Dartmoor Granite, Southwest England. *Zeitschrift für Geomorphologie 35*, 425–438.

Evans, I. S. 1996: Abraded rock landforms (whalebacks) developed under ice streams in mountain areas. *Annals of Glaciology 22*, 9–16.

Fitzsimons, S. J., Lorrain, R. D. & Vandergoes, M. J. 2000: Behaviour of subglacial sediment and basal ice in a cold glacier. *In* Maltman, A. J., Hubbard, B. & Hambrey, M. J. (eds.): *Deformation of Glacial Materials*, 181–190. Geological Society Special Publication 176, London.

Gatliff, R. W., Richards, P. C., Smith, K., Graham, C. C., McCormac, M., Smith, N. J. P., Long, D., Cameron, T. D. J., Evans, D., Stevenson, A. G., Bulat, J. & Ritchie, J. D. 1994: *The Geology of the Central North Sea*. HMSO, London.

Glasser, N. F. 1991: *Modelling the Effects of Topography on Ice Sheet Erosion, Scotland*. Ph.D. dissertation, University of Edinburgh.

Glasser, N. F. 1995: Modelling the effect of topography on ice sheet erosion, Scotland. *Geografiska Annaler 77A*, 67–82.

Glasser, N. F. 1996: Landforms of glacial erosion in the Cairngorm Mountains. *In* Glasser, N. F. & Bennett, M. R. (eds.): *The Quaternary of the Cairngorms*, 104–113. Quaternary Research Association, Cambridge.

Glasser, N. F. & Hall, A. M. 1997: Calculating Quaternary glacial erosion rates in North East Scotland. *Geomorphology 20*, 29–48.

Glasser, N. F. & Siegert, M. J. 2002: Calculating basal temperatures in ice-sheets: an Excel spreadsheet method. *Earth Surface Processes and Landforms 27*, 673–680.

Gordon, J. E. 1979: Reconstructed Pleistocene ice sheet temperatures and glacial erosion in Northern Scotland. *Journal of Glaciology 22*, 331–344.

Gordon, J. E. 1993: The Cairngorms. *In* Gordon, J. E. & Sutherland, D. G. (eds.): *Quaternary of Scotland*, 259–276. Chapman and Hall, London.

Hall, A. M. 1996: The paleic relief of the Cairngorm Mountains. *In* Glasser, N. F. & Bennett, M. R. (eds.): *The Quaternary of the Cairngorms*, 13–27. Quaternary Research Association, Cambridge.

Hall, A. M. & Bent, A. J. A. 1990: The limits of the last British ice sheet in northern Scotland and the adjacent shelf. *Quaternary Newsletter 61*, 2–12.

Hall, A. M., Whittington, G., Duller, G. A. T. & Jarvis, J. 1995: Late Pleistocene environments in lower Strathspey, Scotland. *Transactions of the Royal Society of Edinburgh: Earth Sciences 85*, 253–273.

Hindmarsh, R. C. A. 2001: Influence of channelling on heating in ice-sheet flows. *Geophysical Research Letters 28*, 3681–3684.

Hinxman, L. W. & Anderson, E. M. 1915: *The Geology of Mid-Strathspey and Strathdearn*. Memoir of the Geological Survey of Scotland.

Hubbard, A. 1999: High-resolution modeling of the advance of the Younger Dryas ice sheet and its climate in Scotland. *Quaternary Research 52*, 27–43.

Ives, J. D. 1958: Glacial geomorphology of the Torngat Mountains, northern Labrador. *Geographical Bulletin 12*, 47–75.

Kleman, J. 1992: The palimpsest glacial landscape in northwestern Sweden – Late Weichselian deglaciation landforms and traces of older west-centred ice sheets. *Geografiska Annaler 74A*, 305–325.

Kleman, J. & Borgström, I. 1994: The boulder fields of Mt Fulufjället, west-central Sweden – Late Weichselian boulder blankets and interstadial periglacial phenomena. *Geografiska Annaler 72A*, 63–78.

Kleman, J. & Stroeven, A. 1997: Preglacial surface remnants and Quaternary glacial regimes in northwestern Sweden. *Geomorphology 19*, 35–54.

Kleman, J. & Hättestrand, C. 1999: Frozen-bed Fennoscandian and Laurentide ice sheets during the Last Glacial Maximum. *Nature 402*, 63–66.

Kleman, J., Hättestrand, C., Borgström, I. & Stroeven, A. 1997: Fennoscandian palaeoglaciology reconstructed using a glacial geological inversion model. *Journal of Glaciology 43*, 283–299.

Kleman, J., Hättestrand, C. & Clarhäll, A. 1999: Zooming in on frozen-bed patches: scale-dependent controls on Fennoscandian ice sheet basal thermal zonation. *Annals of Glaciology 28*, 189–194.

Knight, J., McCarron, S. G. & McCabe, A. M. 1999: Landform modification by palaeo-ice streams in east-central Ireland. *Annals of Glaciology 28*, 161–167.

Lian, O. B. & Hicock, S. R. 2000: Thermal conditions beneath parts of the last Cordilleran Ice Sheet near its centre as inferred from subglacial till, associated sediments, and bedrock. *Quaternary International 68–71*, 147–162.

Lindström, E. 1988: Are roches moutonnées mainly preglacial forms? *Geografiska Annaler 70A*, 323–331.

Linton, D. L. 1952: The significance of tors in glaciated lands. *17th International Congress*, 354–357. Washington, DC.

Marshall, S. J., Clarke, G. K. C., Dyke, A. S. & Fisher, D. A. 1996: Geologic and topographic controls on fast flow in the Laurentide and Cordilleran ice sheets. *Journal of Geophysical Research 101* (*B8*), 17827–17839.

McDougall, D. A. 2001: The geomorphological impact of Loch Lomond (Younger Dryas) Stadial plateau icefields in the central Lake District, northwest England. *Journal of Quaternary Science 16*, 531–543.

Naslund, J. O. 2001: Landscape development in western and central Dronning Maud Land, East Antarctica. *Antarctic Science 13*, 302–311.

Olvmo, M., Lidmar-Bergström, K. & Lindberg, G. 1999: The glacial impact on an exhumed sub-Mesozoic etch surface in southwestern Sweden. *Annals of Glaciology 28*, 153–160.

Palmer, J. A. & Nielsen, R. A. 1962: The origin of granite tors on Dartmoor, Devonshire. *Proceedings of the Yorkshire Geological Society 33*, 315–339.

Patterson, C. J. 1998: Laurentide glacial landscapes: the role of ice streams. *Geology 26*, 643–646.

Patterson, C. J. & Boerboom, T. J. 1999: The significance of pre-existing, deeply weathered crystalline rock in interpreting the effects of glaciation in the Minnesota River valley, USA. *Annals of Glaciology 28*, 53–58.

Payne, A. J. & Donglemans, P. W. 1997: Self-organization in the thermomechanical flow of ice sheets. *Journal of Geophysical Research 102* (*B6*), 12219–12233.

Payne, A. J. & Baldwin, D. J. 1999: Thermomechanical modelling of the Scandinavian ice sheet: implications for ice-stream formation. *Annals of Glaciology 28*, 83–89.

Payne, A. J. & 10 others 2000: Results from the EISMINT model intercomparison: the effects of thermomechanical coupling. *Journal of Glaciology 46*, 227–238.

Rea, B. R. 1998: The Cairngorms – a landscape of selective linear erosion. *Scottish Geographical Magazine 114*, 124–129.

Rybicki, G. B. 1989: Dawson's integral and the sampling theorem. *Computers in Physics 3*, 85–87.

Sejrup, H. P., Haflidason, H., Aarseth, I., King, E., Forsberg, C. F., Long, D. & Rokoengen, K. 1994: Late Weichselian glaciation history of the northern North Sea. *Boreas 23*, 1–13.

Siegert, M. J. & Bamber, J. L. 2000: Subglacial water at the heads of Antarctic ice-stream tributaries. *Journal of Glaciology 46*, 702–703.

Siegert, M. J. & Glasser, N. F. 1997: Convergent flow of ice within the Astralobe subglacial basin, Terre Adélie, East Antarctica: an hypothesis derived from numerical modelling experiments. *Polar Research 16*, 63–72.

Sissons, J. B. 1976. *Scotland*. 150 pp. Methuen, London.

Sissons, J. B. 1979: The Loch Lomond Advance in the Cairngorm Mountains. *Scottish Geographical Magazine 95*, 66–82.
Stokes, C. R. & Clark, C. D. 1999: Geomorphological criteria for identifying Pleistocene ice streams. *Annals of Glaciology 28*, 67–74.
Stokes, C. R. & Clark, C. D. 2001: Palaeo-ice streams. *Quaternary Science Reviews 20*, 1437–1457.
Sugden, D. E. 1968: The selectivity of glacial erosion in the Cairngorm Mountains, Scotland. *Transactions of the Institute of British Geographers 45*, 79–92.
Sugden, D. E. 1970: Landforms of deglaciation in the Cairngorm Mountains, Scotland. *Transactions of the Institute of British Geographers 51*, 201–219.
Sugden, D. E. 1974: Landscapes of glacial erosion in Greenland and their relationship to ice, topographic and bedrock conditions. *Institute of British Geographers Special Publication 7*, 177–195.
Sugden, D. E. 1977: Reconstruction of the morphology, dynamics and thermal characteristics of the Laurentide Ice Sheet at its maximum. *Arctic and Alpine Research 9*, 21–47.
Sugden, D. E. & Watts, S. H. 1977: Tors, felsenmeer, and glaciation in northern Cumberland Peninsula, Baffin Island. *Canadian Journal of Earth Sciences 14*, 2817–2823.
Sugden, D. E., Glasser, N. F. & Clapperton, C. M. 1992: Evolution of large roches moutonnées. *Geografiska Annaler 74A*, 253–264.
Sutherland, D. G. 1984: The Quaternary deposits and landforms of Scotland and the neighbouring shelves: a review. *Quaternary Science Reviews 3*, 157–254.
Wilch, E. & Hughes, T. J. 2000: Calculating basal thermal zones beneath the Antarctic ice sheet. *Journal of Glaciology 46*, 297–310.