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A chronostratigraphy of mid- and late-Holocene slope evolution: Creagan a' Chaorainn, Northern Highlands, Scotland

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A HOLOCENE RESEARCH PAPER

Abstract: Unusual exposure of the drift stratigraphy of a typical, vegetated hillslope in the Northern Highlands of Scotland has allowed reconstruction of its Holocene history. Graphic logging of palaeogully fill, sediment analysis, radiocarbon dating and microscopic investigation of horizon boundaries, link periodic slope instability to changes in hydrology and vegetation cover. The oldest preserved organic matter dates from about 7.5 cal. ka BP. Several millennia of subsequent stability were followed, at about 4.3 cal. ka BP, by destabilization whose magnitude and frequency markedly increased after about 2.7 cal. ka BP. Precipitation-driven weathering and erosion offer the best explanation for late-Holocene slope rejuvenation at this site, with change effected through long-lasting shifts in system equilibria as well as high magnitude inputs. Anthropogenic impacts on slope stability are apparently restricted to the last few hundred years. The timing and episodicity of slope evolution suggests climate forcing. Results raise the question of whether the extent of past and potential climate-driven Holocene slope evolution in non-glaciated uplands have been underestimated.

Key words: Slope evolution, Scotland, NW Europe, Holocene climate change, debris flows, colluvium, peat, palaeopodzols, weathering rates.

Introduction

This paper reports the timing, mechanisms and probable triggers for the evolution of a natural slope in NW Scotland, and tests the case for the climatic response which has been established elsewhere in NW Europe.

An increasing weight of evidence indicates that terrestrial Holocene climate change across NW Europe matches the timing of variations in insolation and changes in North Atlantic circulation. For instance, cooling at 8.2 cal. ka BP identified in GRIP and GISP2 records from Greenland ice (Alley *et al.*, 1997) and inferred from marine sediments (Oppo, 1997; Bond *et al.*, 1997; Bianchi and McCave, 1999; Kroon *et al.*, 2000) has equivalents in lake sediments in Norway (Nesje and Dahl, 2001), pollen stratigraphy in N Finland (Seppa and Birks, 2001), and diatom records in N Fennoscandia (Korhola *et al.*, 2000). Repeated southward incursions of ice-rafted debris associated with sea surface cooling of up to 2°C in the eastern N Atlantic as far south as N Scotland, also occurred at about 5.9, 2.8 and 1.4 cal. ka BP (Bond *et al.*, 1997).

On land, a southward-trending, time transgressive, mid-Holocene climate 'optimum' lasting from about 7.5 to 6.0 cal. ka BP, has been identified from, for example, lake levels and glacier fluctuations in Scandinavia (Almquist-Jacobsen, 1995), elevated treelines in Norway (Alm *et al.*, 1996), and dendro-climatological records in Fennoscandia (Eronen *et al.*, 1999). Low lake levels and an elevated treeline between about 6.5 and 4.5 cal. ka BP have also been inferred from combined pine megafossil, pollen and lake level studies in the Swedish Scandes (Barnekow and Sandgren, 2001). There, relatively warm dry summers in the early Holocene (+*c.* 1.5°C) were followed by slightly warmer but wetter conditions (+*c.* 2°C) during the mid-Holocene, with wetter, cooler conditions thereafter. In Britain, climate change has been described as influencing, for example, the mid-Holocene elm (*Ulmus*) decline, between about 6.3 and 5.2 cal. ka BP (Parker *et al.*, 2002), although somewhat later in Scotland. In northern Scotland, descriptions of climate impacts include increased mire-surface wetness after *c.* 6.2 cal. ka BP (Tisdall, 2001); the retreat of pine (*Pinus sylvestris*) after about 4.5 cal. ka BP from its briefly extended

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northwestern limits (Gear and Huntly, 1991; Charman, 1992); and increased mire surface wetness after about 3.9 cal. ka BP (eg, Anderson *et al.*, 1998). Peat humification and speleothem studies in NW Scotland (Charman *et al.*, 2001) suggest peak humidity at 2–300 yr intervals, in phase with GISP2 ice core records, between 2.3 and 1.4 cal. ka BP.

A warm, dry climate in the early Holocene, followed by cooling at about 8.2 cal. ka BP, a 'climatic optimum', then a cooler wetter climate after about 4.5 cal. ka BP, intensifying in the late Holocene, are therefore widely indicated across NW Europe. Some Scottish records appear to be synchronized with N Atlantic sediment records and Greenland ice core proxy climate data, while other data suggest time transgressive effects.

However, although some climate-related, episodic patterns of slope failure have been reported, our understanding of the impacts of that climate variability on mountain slope evolution in NW Europe suffers from a scarcity of high-resolution dates (Ballantyne, 1991; Berrisford and Matthews, 1997; Brunsden and Ibsen, 1997). For instance, after a long period of stability, debris flow frequency in W Norway increased markedly between about 4.4 and 2.7 cal. ka BP, decreasing again during the next millennium (Matthews *et al.*, 1997a,b). Marked increases in slope failure frequency in Norway, related to changes in precipitation and temperature, occurred after 5.4 cal. ka BP, with slope response indicating particularly severe winter conditions between 4.3 and 3.3 cal. ka BP, and a further period of instability in the fifteenth to eighteenth centuries AD (Blikra and Selvik, 1998). The timing of these events is broadly matched by glacier advance in Iceland at 5–4.5, 4.2, 3.0, 2.0 and 1.3–1.2 cal. ka BP (Gudmundsson, 1997), but less clearly related to Bond *et al.*'s (1997) description of southward incursions of cold Arctic surface waters.

In the Scottish Highlands there is evidence suggestive of rapid postglacial slope stabilization (Ballantyne and Benn, 1996), but dated records of natural slope evolution are too sparse to evaluate the influence of climate (Table 1). A number of studies have indicated that mid- and late-Holocene slope evolution is the result of continuing paraglacial sediment redistribution (*sensu* Church and Ryder, 1972) in a sediment-rich, but transport-limited, system (Ballantyne, 1986, 1991; Brazier *et al.*, 1988; Innes, 1997; Ballantyne and Whittington, 1999). In the absence of contrary evidence, this slope activity has been attributed to storm-generated erosion, randomly distributed in space and time (Innes, 1983a; Ballantyne, 1986, 1993; Ballantyne and Whittington, 1999; Curry, 2000), and anthropogenic impacts on sediment-stabilizing vegetation cover (Edwards and Rowntree, 1980; Innes, 1983b, 1997; Ballantyne and Whittington, 1987; Brazier *et al.*, 1988; Tipping, 1995; Curry, 2000), superimposed on intrinsic development (eg, Brunsden, 1993; Phillips, 1999; Thomas, 2001).

Although related to climate, the influence of Holocene weathering on slope stability remains a neglected topic and rates are unquantified (Bain *et al.*, 1990, 1993; Ballantyne, 1991). However, two studies from NW Scotland (Innes, 1983a; Hinchcliffe *et al.*, 1998) describe the impact of postglacial weathering of cliff faces on talus slopes. Mechanisms include mechanical fragmentation resulting from pressure release and freeze-thaw, and chemical weathering (mainly hydrolysis) which is active in these wet, oceanic environments.

The study site

Creagan a' Chaorainn (Figure 1) rises above lower Gleann Chaorainn (5°55′W, 57°30′N) near its junction with Strath Conon, at the eastern margin of the Younger Dryas icecap (Bennett and Boulton, 1993). Its vegetated, north-facing slope is typical of many in the Scottish Highlands. The upper slope consists of a low, near-vertical cliff, above which the summit of Creagan a' Chaorainn rises to 550 m in a series of broad, low-angle steps. Below this, foliated bedrock dipping at an angle of 25–40° has a thin, patchy cover of sandy, bouldery diamicton and eroding blanket peat. Bedrock is feldspar-quartz-hornblende gneiss cut by pegmatites and folded amphibolite dykes, closely fractured in places, probably because of the proximity of the (Caledonian) Strathconon fault zone (Johnstone and Mykura, 1989). It is foliated, jointed and lithologically heterogeneous, giving rise to numerous hydraulically linked fissures and good hydraulic connectivity with the drift cover. Sustained groundwater seepages through the outcrop are maintained in very dry summers, indicating a minimum groundwater residence time of five weeks, and moderately long flow paths.

The mid slope lies at an angle of 15–18° and is dominated by subparallel, vegetated and largely infilled palaeogullies cut in diamicton up to 4 m thick (Figure 2). Depressions marking stabilized palaeogullies are about 50 cm deep. The palaeogullies terminate in a gently convex, 10–12° lower slope and accumulation zone, underlain by undulating bedrock with protruding knolls.

Slope stratigraphy is exposed by recent, high oblique, cross-slope incision to bedrock by a modern gully of comparable scale to the palaeogullies (Figure 3). Further exposures occur in the walls of palaeogullies, which exhibit some consequent headward erosion. Depth to bedrock generally increases downslope, but reduces again downslope at the altitude of the joint-controlled change in strike of the modern gully. Thick blanket peat is confined to the base of the slope where till thins over rising bedrock. Lower slope sedimentation is decoupled from fluvial processes because the Holocene river, Allt Gleann Chaorainn, abruptly cuts down through some 10 m of glaciofluvial outwash into a fault-controlled gorge. Protection of the mid and lower slope sediment sink from fluvial processes has simplified and preserved slope depositional history.

Palaeogullies are typically 3–5 m in width and 8–18 m apart, with a fill consisting of alternating mineral and organic horizons. As subparallel, linear depressions, they are likely to attract run-off and lateral, near-surface flow (Richards *et al.*, 1984; Freeze, 1987). Despite that, they are today vegetated and largely infilled, with only a minor scattering of angular debris at their upper extremities where bedrock is at, or close to, the surface.

The modern gully originates in a hill track drain, and cuts across the palaeogullies at a high oblique angle (Figure 4). It formed between the survey for the 1977 1:10 000 Ordnance Survey map and the Scottish Office Central Research Unit air photographic survey of 1988/89.

Methods

Investigative methods were designed to provide information about the timing and periodicity of organic and mineral horizon accumulation. Ten sections in the walls of the modern gully were logged in detail and those containing datable material are shown in Figures 5 and 6. Twenty-five radiocarbon dates derived from 1-cm-thick peat slices, wood and charcoal were calibrated as 2σ age ranges according to CALIB 3.0 (Stuiver and Reimer, 1993), using the intercept method (Table 2).

Table 1 Dated Holocene geomorphic events in Northern Scotland

Age range	Calibrated dates (2δ)	Processes	Location	Material dated	Authors
8500–7000	8500–6800	enhanced rates of slopewash	Coire Fee, SE Cairngorms	not known	Huntley (1981)
	8000–7000	enhanced rates of slopewash	Carn Dubh, east-central Grampians	intercalated peat	Tipping (1995)
7000–6000	7000–6000	large rotational landslips	Trotternish, Isle of Skye	cosmogenic Cl from slip surfaces	Ballantyne *et al.* (1998)
	6500–6200	hillslope debris flows	Glen Docherty, NW Highland	buried palaeosols	Curry (2000)
	< 6400–6010	gravel terrace on dissected fan	Glen Feshie, W Cairngorms	organic horizon	Bain *et al.* (1993)
6000–5000	6250–5800	enhanced rates of slopewash	Coire Fee, SE Cairngorms	(calibrated by Tipping, 1995)	Huntley (1981)
	< 5920–5310	high altitude solifluction	Cairngorms	buried peat	Sugden (1971)
	5900–5600	debris flows on talus slope	Trotternish, Isle of Skye	buried palaeosols	Hinchcliffe (1999)
	< 5890–4280	debris cone aggradation	Glen Etive, SW Highlands	soil horizon	Brazier *et al.* (1988)
	< 5600–5300	hillslope debris flows	Glen Docherty, NW Highlands	buried palaeosols	Curry (2000)
	5550–4850	enhanced rates of slopewash	Carn Dubh, east-central Grampians	intercalated peats	Tipping (1995)
5000–4000	4900–4600	hillslope debris flows	Glen Docherty, NW Highlands	buried palaeosols	Curry (2000)
	4750–4300	enhanced rates of slopewash	Coire Fee, SE Cairngorms	(calibrated by Tipping, 1995)	Huntley (1981)
	4700–4000	enhanced rates of slopewash	Carn Dubh, east-central Grampians	intercalated peats	Tipping (1995)
	< 4600–4260	high altitude solifluction	Arkle, NW Highlands	buried peat	Mottershead (1978)
4000–3000	< 4080–3720	gravel terrace emplacement	Glen Feshie, W Cairngorms	charcoal in buried podzol	Robertson-Rintoul (1986)
	< 3550–2890/ 2120–1730	enhanced catchment erosion	Braeroddach Loch, E Grampians	original lake sediment	Edwards and Rowntree (1980)
3000–2000	after 3140–2630	high altitude solifluction	Cairngorms	buried peat	Sugden (1971)
	< 2700–2400	floodplain incision	Edendon Valley, central Grampians	peat	Ballantyne and Whittington (1999)
	2200–2100	alluvial fan aggradation	Edendon Valley, central Grampians	peat	Ballantyne and Whittington (1999)
	2300–1700	debris flows on talus slope	Trotternish, Isle of Skye	buried palaeosols	Hinchcliffe (1999)
	< 2120–1830	debris cone aggradation	Glen Feshie, W Cairngorms	buried soil	Brazier and Ballantyne (1989)
	2120–1730	slopewash on talus slope	Trotternish, Isle of Skye	humified organic matter	Innes (1983b)
2000–1000	1900–1800	alluvial fan aggradation	Edendon Valley, central Grampians	peat	Ballantyne and Whittington (1999)
1000–500	1050–570/ 660–330	rapid high altitude solifluction	Fannich Mts, Northern Highlands	buried soil	Ballantyne (1986)
	900–700	alluvial fan aggradation	Edendon, central Grampians	peat	Ballantyne and Whittington (1999)
	< 710–560	slope wash on talus slope	Trotternish, Isle of Skye	humified organic matter	Innes (1983b)
	700 and 500	debris flows on talus slope	Trotternish, Isle of Skye	buried palaeosols	Hinchcliffe (1999)
	< 690	enhanced catchment erosion	Braeroddach Loch, E Grampians	organic lake sediment	Edwards and Rowntree (1980)
	< 650–510	cone incision and fan reworking	Glen Etive, SW Highlands	buried soil	Brazier *et al.* (1988)
500–modern	< 510	debris cone aggradation	Glen Feshie, W Cairngorms	buried soils; root	Brazier and Ballantyne (1989)
	< 450	hillslope debris flows	Glen Docherty, NW Highlands	buried palaeosols	Curry (2000)
	< 250	mainly unconfined debris flows	SW & NW Highlands & Cairngorms	lichenometry	Innes (1983a)

Pre-treatment of peat

Peat is composed of a number of chemically distinct fractions of which humic and fulvic acids in particular have high mobility in natural environments. Consequently, their ages may differ from those of more stable, acid- and alkali-insoluble humin by a factor of up to 10^3 years (Shore *et al.*, 1995; Bol *et al.*, 1996). A proportion of the samples, limited by funding for analysis, but including critical cases, were digested in 2M KOH to remove the humic, as well as fulvic, acids so that results could be evaluated. In detail:

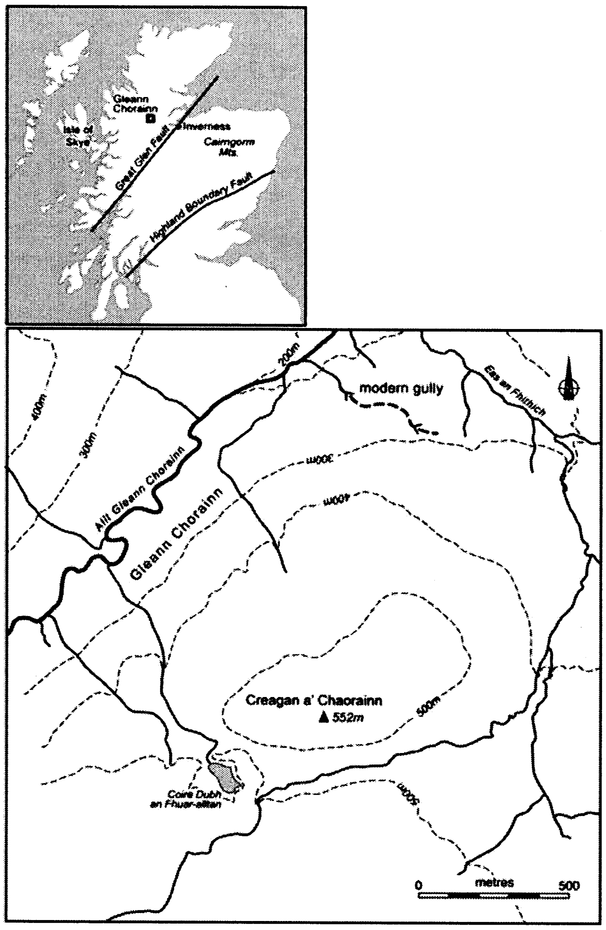

Figure 1 Location map for Gleann Chorainn and Creagan a' Chorainn, northern Scotland

(a) SRR-6052, 6064, 6067–6070 and 6072–6075 were digested in 2M HCl at 80°C for 24 h, washed, filtered and dried. The residue of humic + humin fractions was dated.

(b) SRR-6051, 6053–6063, 6065, 6066 and 6071 were digested in 2M KOH at 80°C until alkali soluble organics (the humic fraction) had been recovered. Excess KOH was removed using distilled water. The humin fractions were then digested in 2M HCl at 80° for 24 hs to remove acid soluble fractions, washed, filtered and dried. The resultant humin fraction (minus humic acids) was dated.

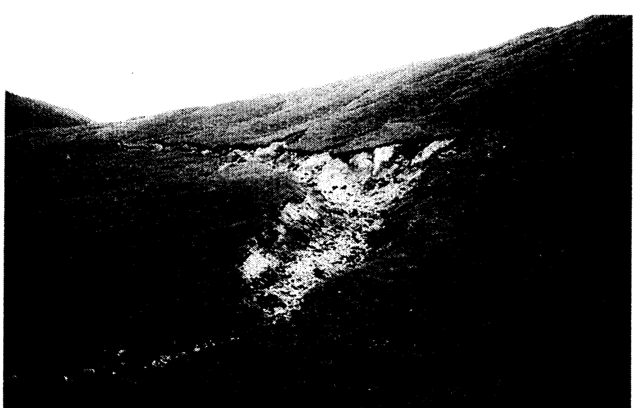

Figure 2 The modern gully crossing the downslope palaeogullies on Creagan a' Chorainn

Figure 3 Dissection into the palaeogully system from the modern gully (site of sections 7a,b and Figure 6)

Peat sampling and classification

Mobile carbon in vertically and laterally permeable drift may contaminate shallow samples. With one exception, where the soil was a poorly permeable sandy peat, peat surfaces immediately below modern soils were not, therefore, dated. The timing of the most recent slope movements, and the onset of critical conditions for the formation of modern soils, are consequently poorly constrained.

Peat humification is related to ground surface humidity (Clymo, 1984, 1991; Kemp, 1985, Fuchsman, 1986). Poor humification arises from rapid incorporation of dead vegetation into the water-logged catotelm, inhibiting oxidative degradation before compaction. A high degree of humification indicates lengthier aerobic degradation in the acrotelm. Peat was visually categorized in the field as (a) black, well-humified, (b) brown, fibrous and (c) poorly humified, with numerous macroscopic plant remains. Photomicrography (Reid, 2001) confirmed the presence in (a), (b) and (c), respectively, of dark, mid and light brown amorphous organic matter, which confirmed the field classification.

Palaeopodzol dating and analysis

Radiocarbon dating of podzols is complicated by long-term upward and downward translocation of organic matter (Matthews, 1984, 1993; Shore *et al.*, 1995; Fitzpatrick, 1986; Schaetzl and Isard, 1996). Palaeopodzol samples were taken from the basal centimetre of A horizons where carbon with the longest residence time has been found in age/depth profiles in alpine podzols, and offers the best approximation to the date of the earliest stages of profile differentiation (Matthews, 1984; Caseldine and Matthews, 1985).

The concentration of free iron and aluminium in podzol B horizons is related to the degree of weathering, leaching and illuviation during soil development (Fitzpatrick, 1986; Fanning and Fanning, 1989; Avery, 1990). High concentrations may indicate either long-lasting, or particularly rapid, soil development. Sampling was constrained by the small number of distinctly demarcated horizons where the basal 20 mm were sufficiently fine-grained to permit the collection of 200 g of particles <2 mm in diameter required for analysis. After abstraction in an 0.1M solution of $K_4P_2O_{17}$ (potassium pyrophosphate), concentrations of iron and aluminium were measured using conventional flame Atomic Absorbtion Spectrometry (Fe) and Graphite furnace AAS (Al). Results are presented as the mean of three assays.

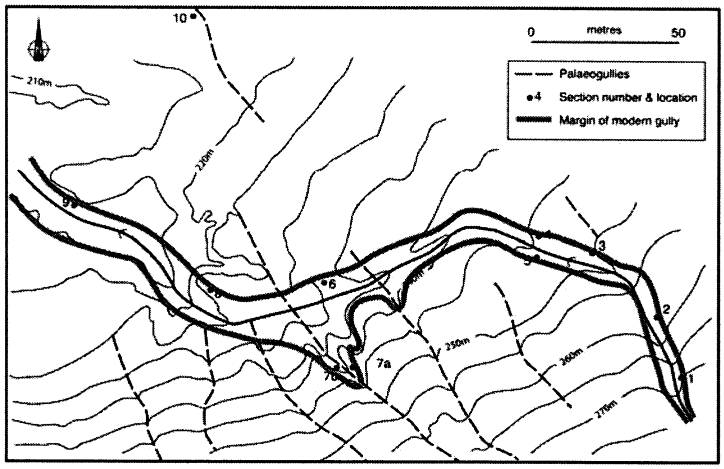

Figure 4 Ground survey map showing the palaeogullies transected by the modern gully, and location of sites 1–10

Mineral sediment analysis

Samples were taken from (a) fresh, interpalaeogully zone exposures in the walls of the modern gully and (b) palaeogully debris layers deposited over a period of 5000 years (Section 9). Clast size distribution, mineralogy, weathering, rounding and macroscopic organic content of till and interpeat debris in palaeogullies were compared. The mineralogy of sieved, weighed samples was determined from petrological microscopy. Grain size analysis was based on the Udden–Wentworth scale. Weathering was determined by identifying the presence of break-down products of the original rock-forming minerals, and their impact on changes in colour and rounding.

The silt and sand content of peat is a source of information about the sediment content of run-off (hence vegetation cover), and the extent of mixing of peat and mineral sediment at peat–debris flow boundaries. It was therefore assessed in hand specimens and microscope slides.

Microscopic investigation of mineral horizon/peat boundaries

The ages of upper peat surfaces in stratified peat–debris sequences have been used to date the emplacement of overlying debris on the basis that sharp contacts are non-erosional (eg, Innes, 1983a; Matthews *et al.*, 1997a; Curry 2000). Since this potentially incorporates large errors, Kubiena tin samples spanning peat–debris contacts (and one boundary between very well humified and very poorly humified peat) were collected where sediment coarseness permitted. Samples were dried over acetone vapour before being impregnated with resin thinned with acetone to accommodate the low permeability of peat. The extreme contrast between peat and unconsolidated sediment demanded very gentle outgassing to replace air bubbles with resin before 50 mm × 60 mm thin sections were produced. Conventional petrographic interpretation was applied to the description and interpretation of mineral sediment and the deformation and erosion of peat and fragile organic particles. Pedological criteria were used to assess peat humification and to identity some soil-forming processes (Kemp, 1985; Bertran, 1993; Elliott, 1996).

Results

Distribution of mineral and organic deposits

Palaeogully fill is variable. It largely consists of centimetre- to decimetre-thick peat alternating with layers of coarse diamic-

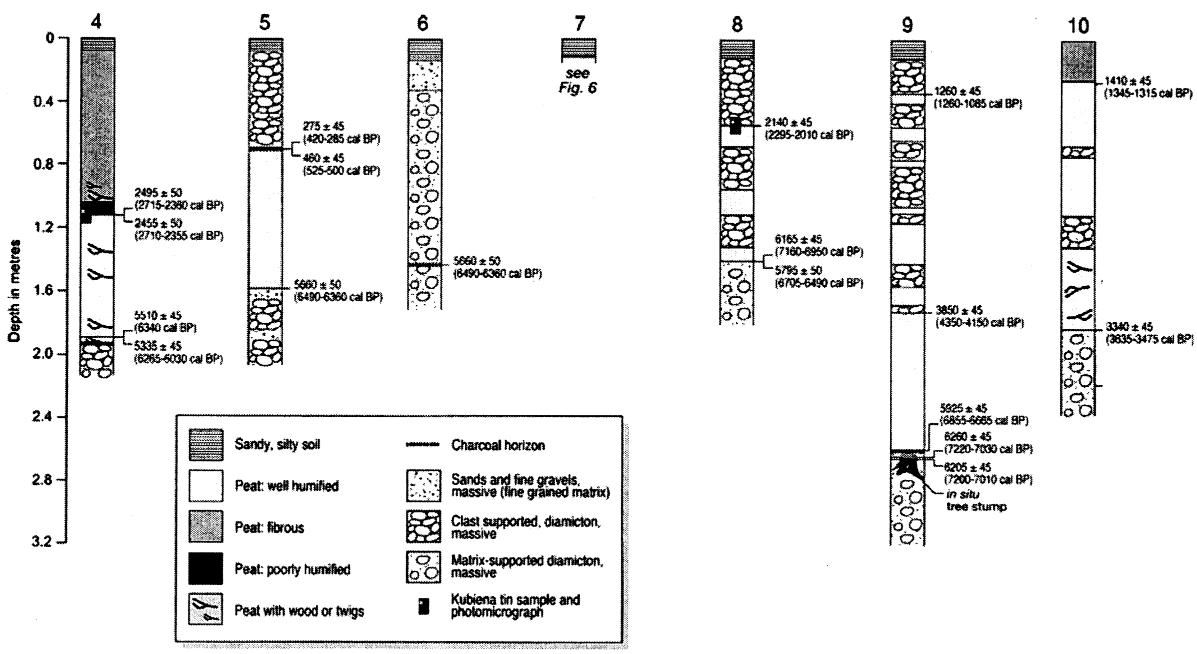

Figure 5 Stratigraphic logs for sections at sites 4–10 on Figure 4

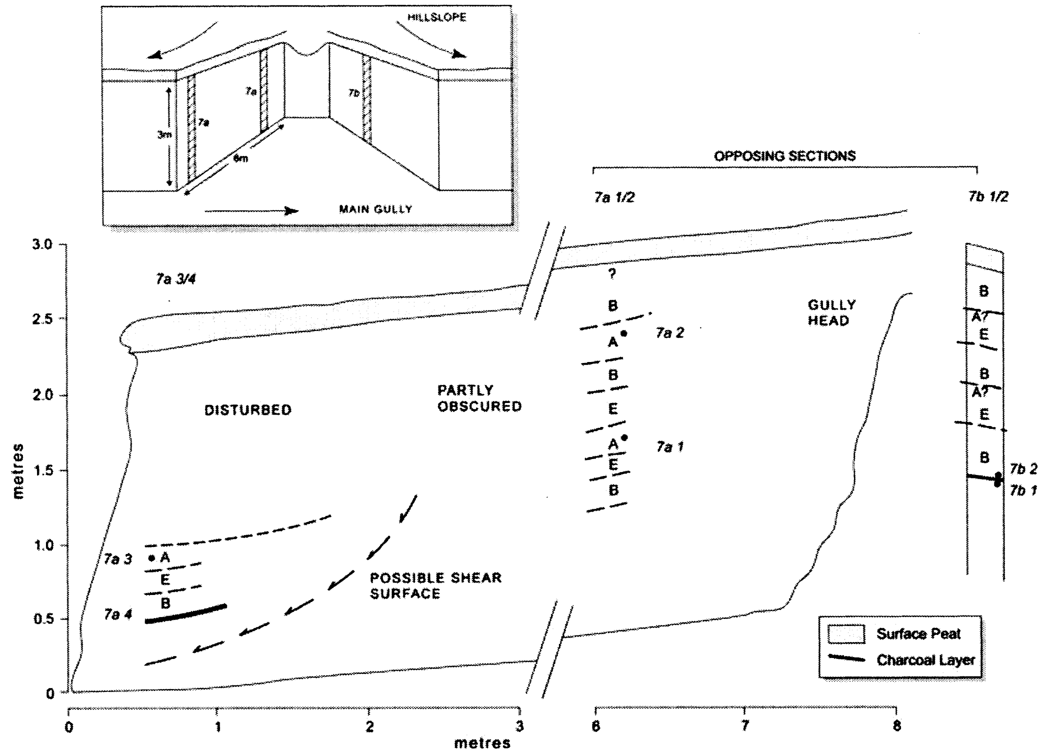

Figure 6 Sections in the side-gully (sites 7a,b) showing palaeopodzols and charcoal layers (Figure 3 also refers)

ton (eg, Sections 8, 9); thick peat (Sections 4 and 5); and stacked, eroded palaeopodzols developed in the coarse drift Sections 7a and 7b; and an *in situ* tree stump associated with thin immature soil (Section 9). The greatest number of peat/mineral couplets is found in sections which sample the lower slope. Macroscopic charcoal is associated with peat and palaeosols (eg, Sections 6 and 8). Organic matter in intergully zones is restricted to centimetre-thick layers of woody charcoal within diamicton, marking palaeosurfaces with a similar angle to the modern slope. Sections 6 and 7a sample such charcoal.

Reliability of radiocarbon dates

Materials dated (Table 2) are peat, charcoal, *in situ* wood and palaeopodzols. Dates are largely internally consistent within each logged section. Their reliability is illustrated by samples 9/1a and b where the age of scorched wood on the surface of an *in situ* tree stump (9/1a) is statistically indistinguishable from that of dispersed charcoal (presumably from the same fire) in the surrounding thin mineral soil (9/1b).

Charcoal dates from humin-only fractions show two date reversals (samples 7b/1 and 7b/2; 8/1 and 8/2, Table 2, Figures 5

Table 2 Radiocarbon and calibrated dates: Creagan a' Chaorainn

Lab code	Sample	Material	^{14}C age	cal. BP	δ^{13}C (%)
SRR-6065	7b/2	charcoal (humin only)	6615±45	7525–7395	−26.4
SRR-6063	7a/4	charcoal (humin only)	6405±45	7380–7300	−26.4
SRR-6069	9/1b	soil (humin+humic acids)	6260±45	7220–7030	−27.6
SRR-6070	9/1a	wood (humin only)	6205±45	7200–7010	−26.7
SRR-6067	8/2	peat (humin+humic acids)	6165±45	7160–6950	−28.5
SRR-6064	7b/1	peat (humin+humic acids)	5970±45	6890–6745	−27.2
SRR-6071	9/2	charcoal (humin only)	5925±55	6855–6665	−28.1
SRR-6066	8/1	charcoal (humin only)	5795±45	6705–6490	−27.3
SRR-6055	5/1	peat (humin only)	5660±50	6490–6360	−26.4
SRR-6052	4/2	peat (humin+humic acids)	5510±45	6395–6280	−27.6
SRR-6059	7a/1	podzol (humin only)	5450±45	6290–6200	−28.1
SRR-6051	4/1	charcoal (humin only)	5335±45	6265–6030	−28.2
SRR-6058	6/1	charcoal (humin only)	5290±45	6170–5945	−25.6
SRR-6061	7a/3	podzol (humin only)	4560±45	5440–5050	−28.6
SRR-6072	9/3	peat (humin+humic acids)	3850±45	4350–4150	−28.5
SRR-6074	10/1	peat (humin+humic acids)	3340±45	3635–3475	−28.3
SRR-6053	4/3	peat (humin only)	2495±50	2715–2360	−25
SRR-6054	4/4	peat (humin only)	2455±50	2710–2355	−25.3
SRR-6062	7a/3	podzol (humin+humic acids)	2435±45	2705–2355	−27.9
SRR-6068	8/3	peat (humin+humic acids)	2140±45	2295–2010	−28.6
SRR-6060	7a/2	podzol (humin only)	1820±45	1820–1630	−31.4
SRR-6075	10/2	peat (humin+humic acids)	1410±45	1345–1285	−27.7
SRR-6073	9/4	peat (humin+humic acids)	1260±50	1260–1085	−27.9
SRR-6056	5/2	peat (humin only)	460±45	525–500	−28.5
SRR-6057	5/3	charcoal (humin only)	275±45	420–285	−28.4

and 6). A thin peat at the base of Section 7b is at least 500 years younger than the charcoal on its upper surface. Similarly, sample 8/1 (charcoal) is several hundred years younger than the overlying peat (sample 8/2). Since the humin-only charcoal dates are likely to be more accurate than the mixed-fraction peat dates, both reversals are probably attributable to translocated carbon producing anomalously young peat ages. In Section 7b, an alternative explanation is that charcoal with a long residence time was washed onto the unvegetated diamicton long after its formation. Against this, preservation of fragile, centimetre-scale, fragments of wood charcoal for hundreds of years on a mobile slope seems less probable than abrasion and comminution.

Peat

Buried peat surface dates range from about 7.1 to 1.2 cal. ka BP (samples 7b/1, 9/4) although the former may be several hundred years too young (see reversed dates above). Early peats are uniformly well-humified. On palaeogully floors, sand-free, well-humified peat replaced diamicton between 7.2 and 6.3 cal. ka BP (Table 3). A change in peat type through time is evident in Sections 4 and 10, with fibrous peat replacing well-humified peat after about 2.4 cal. ka BP in the former, and after about 1.3 cal. ka BP in the latter. In Section 4, a sharp boundary separates black well-humified peat (sample 4/4) and poorly humified peat (sample 4/3) which itself gives way upwards to the fibrous variety. Statistically indistinguishable dates of about 2.5 cal. ka BP from either side of the sharp boundary indicate a rapid and persistent switch from one type to the other. The boundary is also marked by a sudden influx of silt and sand which includes some pitted (etched) fine sand and silt-sized quartz (Reid, 2001). No such trend in humification is apparent in successive thin peats (Sections 8, 9, 10).

Wood

Wood in the form of twigs is variably present in well-humified peat, and less frequently in fibrous peat. It is moderately to poorly preserved. Moderate preservation is defined here as soft, but recognizable lengths of small diameter round wood. Poor preservation is equated with fragments of wood fibre grading into amorphous peat. In poorly humified peat, wood is entirely absent.

One small *in situ* tree stump (Section 9) dated at *c.* 7.2 cal. ka BP, together with undated stumps in Sec. 2, and probable *in situ* tree roots (Section 5), were recorded. The dated stump is underlain by diamicton and overlain by thin immature soil and peat. In Section 5, the undated roots grew after 420 cal. BP. No *in situ* wood of intermediate age is preserved.

Macroscopic charcoal

Charcoal occurs on peat surfaces (eg, Section 5), on surface-parallel palaeosurfaces within coarse diamicton (Figure 6, sections 6, 7a, 8; Figure 7), in association with displaced palaeopodzols (Section 7a), on the burned surface of the *in situ* tree stump (sample 9/1a) and in association with thin, eroded mineral soils (Section 6, sample 9/1b). In diamicton, charcoal consists of thin layers of 2 mm–2 cm fragments retaining the form of wood fibres (Reid, 2001, Figure 5.16). On some peat surfaces and in eroded mineral soils it occurs as millimetre-scale, soft, plastic, often discontinuous layers, which may have formed from sparse, lignin-poor herbaceous plants. Its distribution in time is strikingly uneven (Table 4) since it is absent after about 6.0 cal. ka BP with the exception of one historical occurrence.

The association of macroscopic charcoal with slope instability is charted in Table 5. Burning of woody vegetation was typically followed, not by slope failure, but by the establishment of vegetation cover, accompanied by a rise in the water-table where drainage was poorer.

Table 3 The timing of basal peat growth on palaeogully floors

	Sample no.	^{14}C age	cal. BP
Increasing altitude ↑	4/2	5335±45	6395–6280
	5/1	5660±50	6490–6360
	7b/1	5970±45	6890–6745
	8/2	6165±45	7160–6950
	9/1	6260±45	7220–7030
	10/1	3340±45	3635–3475

Palaeopodzols

Palaeopodzols truncated to the level of either A or E horizons occur in Sections 7a and 7b on the opposite wall of the same, headward-eroding palaeogully (Figures 3, 8). Trace-only preservation of A horizons meant that no dates were obtained from 7b. Consequently, although each Section contains three stacked, eroded podzols, they could not be correlated.

In Section 7a, early profile differentiation occurred at about 6.2 cal. ka BP (sample 7a/1), 5.4–5.0 cal. ka BP (7a/3), and 1.8–1.6 cal. ka BP (7a/2). Disrupted podzol profiles below charcoal sample 7a/4 dated at 7.3 cal. ka BP in the lower part of Section 7a, suggest even earlier podzolization, while the youngest, poorly defined profile beneath the modern soil post-dates 1.6 cal. ka BP. Podzols therefore formed repeatedly over a period of at least 6000 years, and suffered disruption through repeated shallow, slope failure. Samples 7a/1 and 7a/2 appear to be sequential, but are separated by an age gap of some 4500 years. Since a disturbed profile of intermediate age (7a/3) appears downslope, they are probably separated by one or more cryptic erosion surfaces. The apparently sequential profiles in Section 7b may also therefore conceal erosional hiatuses.

For sample 7a/3, both humin-only and mixed carbon fraction dates were obtained. At least 2345, and up to 3085 calendar years separate them, illustrating the mobility of humic acids compared with humin, and the uncertainty attendant on mixed fraction podzol dates.

Oxalate extractable Fe and Al concentrations in podzol B horizons offer further information about the conditions of soil development. Table 6 compares Creagan a' Chaorainn palaeosol analyses with the few published figures for undated surface soils on slightly gentler slopes in the Scottish uplands. In the absence of modern baseline data, the figures are difficult to interpret, although iron is strongly enriched in the palaeosols compared with surface soils. The undated palaeopodzol from Section 7b has much the highest values of both Fe and Al. By definition, palaeosols cease to develop after burial. However, the continuation of iron complex translocation after soil burial on a humid permeable slope has not been excluded.

The lack of information from comparable settings means that an absolute standard for 'high' or 'low' concentrations is not available. Preliminary observations are that:

- Both metal concentrations increased from about 6.2 to 5.2 cal. ka BP, with a further increase in the late Holocene.
- Undated sample 7b/X is the most intensely leached.
- Repeated slope failures mean there is no linear relationship between depth below the surface and Fe and Al concentrations.
- Fe (but not Al) is greatly enriched in the palaeopodzols compared to their modern counterparts.

Figure 7 Organic layer with charcoal, in till at section 7a (Figure 6 refers)

Only maximum ages for the timing of the failures which truncated and/or translocated, podzols can be derived from soil dates, because there is no information on the time lapse between full profile development and slope failure. Profile differentiation itself may require between 350 years (Charman, 1992) and 1000 years (Fitzpatrick, 1986). Till failures that truncated podzol profiles probably therefore occurred at some point after 7.1 cal. ka BP (Sample 7b/2); 7.0 cal. ka BP (Sample 7a/4); and 5.9 cal. ka BP (Sample 7a/1).

Classification of mineral sediment

Analyses of samples of slope-blanketing diamicton and interpeat debris are summarized in Table 7. The mineralogy of the former reflects bedrock geology. The sediment is mineralogically and texturally immature, poorly sorted and lacks organic matter – characteristics consistent with a locally derived till. Water-saturated flow of till is indicated by frequent downslope-pointing elongate clasts. Crude water sorting is less common, while thicker interpeat debris horizons are reversely graded. Nearly 50% of debris particles are 50–150 mm in diameter; no till particles exceed 40 mm. Fine sand and silt form 12.5% of the till, but $<5\%$ of interpeat debris.

In summary, interpeat debris superficially resembles till but is distinguished from it by very local mineralogy, reverse grading in the thicker flows; larger maximum clast size; a much coarser skew in grain size distribution; more angular grains; and the presence of some weathered particles and organic matter. The similarity of the three samples deposited at intervals over a time span of 5000 years indicates a single, long-term source. Small exposures of shattered bedrock visible at the upper ends of some palaeogullies, and more extensively in the floor of the modern gully, are consistent with slope bedrock as that source. Weathered clasts, eroded peat and rootlets can be explained as surface material sparsely entrained in debris flows.

The third form of mineral deposit is slopewash. Slopewash takes several forms: occasional centimetre-scale layers derived from erosion of immature mineral soils (Sections 4, 6 and 9); undated washes of coarse, pebbly sand interrupting the development of thin peaty soils near the top of Section 6; sand and silt in peat at *c.* 2.7–2.4 cal. ka BP (above sample 4.3); and the matrix of modern soils.

Modern soils vary from sandy peat at the base of the slope, to friable, unlayered, immature, silty sands. Because of the difficulty of excluding modern carbon from samples, no date for the accumulation of their silt-sand matrix has been established, but Sample 5/3, and the sandy peat above Sample 10/2, suggest maximum ages younger than 420–285 BP and 1.3 cal. ka BP, respectively. No comparable thicknesses of silty sand are found in older deposits.

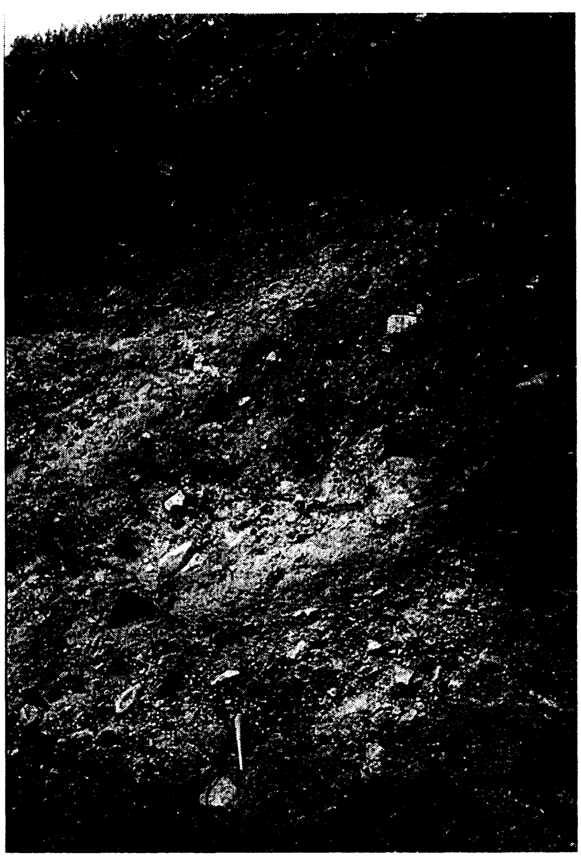

Figure 8 Palaeopodzols at section 7a (Figure 6 refers)

Table 4 Macroscopic charcoal dates

Sample	^{14}C age	cal. BP	depth (m)
7b/2	6615 ± 45	7525–7395	1.6
7a/4	6405 ± 45	7380–7300	1.5
9/2	5925 ± 55	6855–6665	2.6
8/1	5795 ± 45	6705–6490	1.4
4/1	5335 ± 45	6265–6030	1.9
6/1	5290 ± 45	6170–5945	1.3
5/3	275 ± 45	420–285	0.6

Table 5 The association of charcoal and slope instability on Creagan a' Chaorainn

Section	Stratum overlying charcoal	
	Mineral	Organic
4	2 cm slopewash	overlain by thick peat
5	debris flow	
6	debris flow	
7a		podzol
7b		peat
8		peat
9		peat

In summary, till may date from the end of the Younger Dryas Stade (Bennett and Boulton 1993), but episodic Holocene erosion of unstable bedrock in the gully system appears to be the main source of interpeat debris. Slopewash made a minor contribution to downslope transport until deposition of the 10–15 cm thick layer which forms the basis of modern soils.

Debris flow/peat boundaries

Reversely graded debris flows in palaeogullies brought mainly granule to silt-sized particles into contact with peat substrates, the fines acting as a conveyor belt on which large (up to boulder-sized) angular particles floated. Photomicrographs (Reid, 2001) show two principal effects of debris flows on underlying peat surfaces – little or no disturbance and small-scale mixing (Figure 9). Bouldery flows up to 0.5 m thick affected underlying peat surfaces to a maximum depth of <20 mm. Preservation and gentle deformation of occasional delicate plant fragments indicate low-energy mixing processes at the peat/debris interface. Any disturbance may be the product of sudden changes in hydrostatic pressures accompanied by briefly increased frictional forces in the soft peat substrate as debris flows defluidized on coming to rest, then reverted to a rigid state (Iverson, 1997; Major and Iverson, 1999). The low vertical, high lateral, permeability of peat (Tallis, 1965; Korhola, 1994) may restrict the depth of deformation.

The nature and chronology of Holocene slope evolution

The newly deglaciated slope may have been gullied and stabilized at a very early date (Ballantyne and Benn, 1996), but at least three millennia elapsed before the dated history begins at about 7.5 cal. ka BP (Sample 7b/2).

The oldest preserved organic materials developed on previously mobile inorganic sediment consist of charcoal (samples 7b/2, 7a/4), eroded thin mineral soil (9/1b), an *in situ* tree stump (9/1a) and peat (8/2), indicating that between about 7.5 and 7.0 cal. ka BP the slope was vegetated. Available dates for subsequent till failures, defined by buried charcoal surfaces and truncated podzol A horizons, are listed in Table 8. Repeated till failures buried the charcoal surface (sample 7a/3) to a depth of 2 m. Neither the level of podzol truncation nor the gentle (10–12°) slope angle match a theoretical failure model for slope podzols proposed by Brooks and Richards (1994), which suggested steep slopes and failure surfaces focused in B, rather than A horizons because of higher permeability in the former. Repeated, but mainly undated, podzol truncation and disrupted stratigraphy, indicate that dates in Table 8 merely sample the real number of events.

On palaeogully floors, sand-free, well-humified peat replaced mobile till between 7.2 and 6.3 cal. ka BP (Table 3), its growth episodically interrupted by flows of locally derived bedrock. Dated debris flows occurred at about 4.3 cal. ka BP (sample 9/3), 2.3 cal. ka BP (8/3), 1.3 cal. ka BP (9/4) and 0.5 cal. ka BP (5/2). However, the number of logged gully-constrained debris flows (15) greatly exceeds the number of dated peat surfaces from which their timing could be identified (4). In the absence of high-resolution dating, which could constrain peat growth rates, interpolated dates for the remaining surfaces are subject to unknown error. Bracketing dates indicate two flows between 6.9 and 2.9 cal. ka BP (Section 8); two flows in the period 6.6–4.4 cal. ka BP (Section 9); five flows between 4.0 and 1.0 cal. ka BP (Section 9), and three flows between 3.0 and 1.3 cal. ka BP (Section 10). These observations do not reflect the increasing late-Holocene event volume and frequency visible in Sections 8, 9 and 10. In Section 8, the debris deposited between 2.3 and 2.0 cal. ka BP is more than twice as thick as that which truncated the earliest phase of peat growth before 6.0 cal. ka BP. Section 9, which samples the main sediment sink on the slope and preserves nine peat/debris couplets, confirms this trend. Between 6.8 and 4.3 cal. ka BP the slow accumulation of peat was interrupted by only small volumes of debris on three occasions in the space of 2.5 ka, and mineral debris forms less than 20% by volume of the stratigraphic column – a debris:peat ratio of 1:5. Between 4.3 and 1.3 cal. ka BP, incursions of debris occurred on six occasions and the debris/peat thickness ratio is reversed to nearly 3:1, increasing to 4:1 above a depth of 1.1 m. Thus the late-Holocene ratio of debris to peat was up to 20 times greater than before 4250 cal. BP, with debris transported down the gully both more frequently, and in larger volumes.

The timing of changes in slope stability and hydrology are summarized in Figure 10.

Discussion

By analogy with patterns of rapid slope stabilization after recent deglaciation (Ballantyne and Benn, 1996), the north slope of Creagan a' Chaorainn, like many comparable vegetated slopes in the Scottish Highlands, probably rapidly stabilized in the earliest Holocene. The diversity of subsequent

Table 6 Fe/Al concentrations in palaeopodzol B horizons on Creagan a' Chaorainn (see Figure 5; slope 15–18°). Data for two surface podzols on slopes elsewhere in the UK are provided for comparison

Sample	^{14}C age (cal. ka BP)	Thickness (m)(Bh horizon)	Depth of sample (m)	Al (mg/l)	%Al	Fe (mg/l)	%Fe
7a/2	1820±45 (1725)	0.10–0.12	0.60–0.58	49.28	0.49	171.19	1.71
7a/3	4560±75 (5245)	0.25–0.30	0.95–0.92	47.08	0.45	143.71	1.36
7a/1	5450±45 (6245)	0.20–0.25	0.80–0.78	40.07	0.41	86.8	0.88
7b/X	not dated	0.10	1.0–0.98	69.25	0.68	185.01	1.83
Author	*Location*	*Material*	*Age*				
Avery 1990	Cairngorms	surface soil (slope 12°)	not reported		0.60		0.20
Avery 1990	Cumbria	surface soil (slope 11°)	not reported		0.20		0.30

Table 7 Characteristics of interpeat debris (shaded) and till (unshaded)

Clast size/(mm)	% of mass	Cumulative%	Mineralogy
50–150	49.3	49.3	Angular/very angular in all fractions except *c.* 5% of > 20 mm clasts
30–40	30.0	30.0	Angular to subangular in all fractions
30–50	16.9	66.2	
20–30	11.2	41.2	
20–30	9.5	75.7	Occasional weathered clasts
10–20	10.6	51.8	Mica schist particles (*c.* 70%); with pink (minor grey) gneiss & pegmatite
10–20	5.5	81.2	Pink (minor grey) gneiss & pegmatite with mica schist
1.5–10	19.1	70.9	
1.5–10	6.3	87.5	Quartz grains dominant in < 0.2 mm fraction
1.0–1.5	5.0	75.9	Quartz grains dominant in < 0.5 mm fractions
1.0–1.5	1.5	89.0	Fresh mica, feldspar & ferromagnesian particles in < 1.5 mm fraction
0.5–1.0	2.7	78.6	
0.5–1.0	1.4	90.4	Rare rootlets and amorphous organic matter
0.2–0.5	8.7	87.3	Fresh mica laths in fine fraction
0.2–0.5	5. 1	95.7	
< 0.2	12.7	100	
< 0.2	4.3	100	

palaeogully fill, the accumulation of erosion products towards the base of the slope, and the uncertain extent of preservation of intergully organic horizons in a coarse, mobile matrix, impedes stratigraphic correlation between Sections. The emphasis therefore, is on chronostratigraphic correlation based on indicators of stability and instability, surface wetness and the nature of vegetation cover. The history of episodic development spanning seven millennia, revealed by anthropogenic gullying in the last few decades, comprises several phases of development.

Before about 7.5 cal. ka BP preserved organic matter is absent. Palaeogullies were floored by mobile slope debris, and intergully zones preserve no evidence of stabilizing vegetation. Yet the slope had previously been deglaciated for over 3000 years and, by analogy with many well-studied sites in Scotland reviewed in Birks (1996), Huntley (1996), Lowe (1993), rapid postglacial herb, shrub and tree colonization of the fresh mineral substrate, accompanied by organic soil formation, is likely to have occurred. Since incision to bedrock in the modern gully means that lack of evidence for vegetation growth before 7.5 cal. ka BP is not an artefact of exposure, an alternative explanation is necessary. Similar evidence for slope mobility, combined with the absence of organic matter before about 7.5 cal. ka BP in a second valley in the northern Highlands, (Gleann Lichd 57°10′N, 0°W), indicates several millennia of net erosion before 7.5 cal. ka BP and suggests a common factor (Reid, 2001). This date provides a possible measure of the timescale of system adjustment to Holocene conditions. In the earliest Holocene, drier, warmer conditions may have resulted in more rapid oxidation of organic matter on mobile slopes. We deduce that net erosion during this phase was also controlled by continuing paraglacial sediment redistribution. Thereafter, a system change from net erosion to net accumulation may have been triggered by the climatic cooling at about 8.2–7.8 cal. ka BP in NW Europe, engendering (subject to other parameters discussed below) conditions for preservation of buried organic matter and growth of blanket peat. Average peat growth on the Scottish mountain slopes ranges from 1.4 to 4.0 cm/100 yr (Pears, 1975; Binney, 1997; Reid, 2001). Even 5–10 cm thick palaeogully peats therefore probably represent timespans of hundreds of years of uninterrupted vegetation on formerly mobile till in gullies that would have concentrated surface flow. Preservation of organic matter therefore implies that paraglacial sediment reworking had ceased to be the dominant slope process, since there was now a minimum of several hundred years between slope failures, despite a cooler, wetter climate. Once established, blanket peat, acting as an earth (rather than biological) material, probably established new system thresholds through its impact on slope hydrology and till mobility.

We therefore propose a timescale of some 3000 years for the bulk of slope system adjustment to postglacial conditions to be

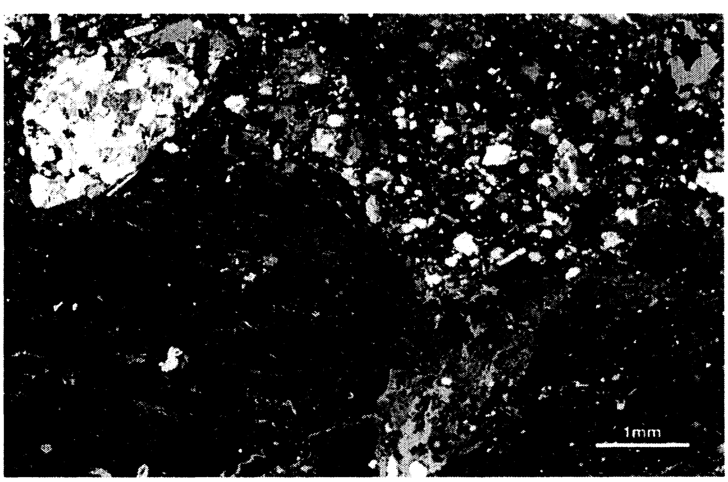

Figure 9 Photomicrograph from section 8 (Figure 5) showing mineral material from debris flow over a peat surface, indicating some impregnation but no disruption of the peat surface (× 20, XPL)

Table 8 Dated till failures on Creagan a' Chaorainn

Sample	Age range	Debris flow after[a]
7b/2	7525–7395[b]	7.5–7.4
7a/4	7380–7300[b]	7.4–7.3
6/1	6170–5945[b]	6.1–6.0
7a/1	6290–6200[c]	6.0–5.9
7a/3	5440–5050[c]	5.1–4.7
7a/2	1820–1630[c]	1.5–1.3

[a]The 'debris flow after' date allows for a minimum 350 yr soil development. Soil life spans before failure are unknown, so these dates are maxima.
[b]Age of charcoal surface buried by till.
[c]Age of (truncated) podzol A horizon buried by till, ie, the approximate timing of early organic accumulation.

completed on Creagan a' Chaorainn, and probably more widely in the Northern Highlands of Scotland. If this conclusion is valid, it suggests that the 10 ka timescale for paraglacial sediment redistribution determined in young mountain systems in North America (Church and Slaymaker, 1989) and New Zealand (eg, Church and Ryder, 1972), which has been applied to the Scottish uplands (eg, Brazier *et al.*, 1988; Ballantyne and Benn, 1996) may not be applicable in this setting. Alternative mechanisms for continued slope mobility must then be invoked. They are principally (and not mutually exclusively), climate change, weathering rates, evolving slope sensitivity, human impacts and vegetation changes.

The persistence of the palaeogully system throughout the Holocene indicates that it dominated mid-slope drainage during phases of both net erosion and net accumulation. Slopes with permeable drift and good hydraulic connectivity between drift and bedrock (conditions conducive to persistent leaching, and medium- to long-term groundwater flow paths and residence times) are likely to be sensitive to hydrological changes arising from climate variability. Shifts in the balance between interdependent system parameters such as weathering rates, shear strength of materials, water flow and vegetation cover, may arise from quite subtle adjustments in the relative rates and magnitudes of processes (eg, Thomas and Thorp, 1995; Phillips, 1999). Preservation of organic matter from about 7.5 cal. ka BP may thus have arisen when threshold conditions related to declining rates of paraglacial sediment redistribution, accompanied by a decline in erosive storms and snowmelt, permitted. Large or sudden changes are not necessary to trigger threshold conditions, and the evidence from Creagan a' Chaorainn tends to support marked system change arising from modest scale and/or gradual processes. Subsequent altered system states indicate further change leading to renewed breaching of stability thresholds. Details are considered below.

Episodic Holocene slope evolution on Creagan a' Chaorainn can be inferred from: tree and herb growth on the formerly mobile substrate, on surfaces which were subsequently buried; a switch from erosion to deposition on palaeogully floors; changes in the magnitude and frequency of till failures and bedrock erosion through time; the development and truncation of successive podzol profiles; slopewash associated with

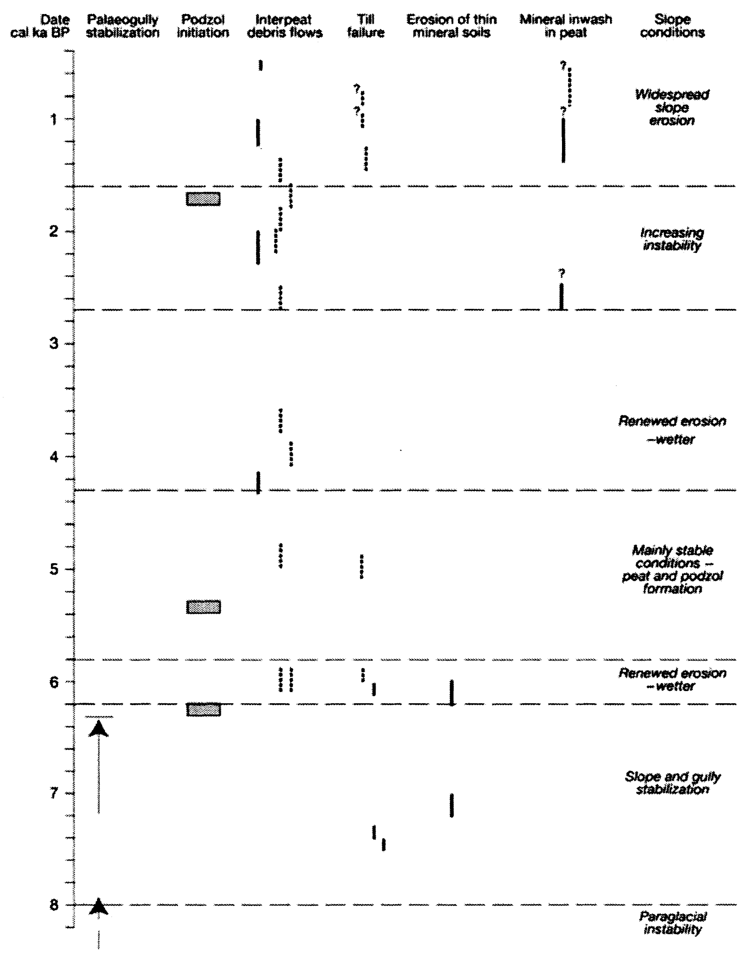

Figure 10 Chronology of slope evolution at Creagan a' Chorainn. Solid lines show dated events with error bars; dashed lines represent timespans of undated events

breached vegetation cover; changes in slope vegetation and hydrology; and charcoal stratigraphy. The sequence is summarized in Figure 10.

Erosion to bedrock and mobile debris in the modern gully suggest that even moderate seasonal flows prevent vegetation (hence peat) establishment. The appearance of peat on palaeogully floors after about 7.2 cal. ka BP therefore signals an environmental threshold. In the context of a reduced paraglacial sediment supply, and a warmer, drier climate, seasonally erosive conditions may have given way to a régime where intense storms and snow melt became rare. The virtual absence of sand and silt in palaeogully peats is informative. On average, less than 15% of vegetable matter contributing to peat survives leaching and bacterial degradation (Fuchsman, 1986). Fine sediment in precursor vegetation should therefore be concentrated at least six-fold during peat formation. Consequently, sediment-free palaeogully peats must have formed when the vegetation cover was closed and slope run-off almost sediment-free. This observation is consistent with a transition from characteristically erosive flow to steady seepage, which allowed establishment of a closed vegetation cover on the slope as well as palaeogully stabilization and infill.

Thereafter, the rate and volume of debris transport in palaeogullies varied systematically with time (see Results section final paragraph). In the zone of maximum sediment accumulation (Section 9) the ratio of debris to peat in the late Holocene is up to 20 times greater than before 4250 cal. BP, although dated events only sample the actual incidence of slope failures, and the enhanced late-Holocene frequency and magnitude of palaeogully debris flows is disproportionately under-represented. Estimated timings of events (dotted lines in Figure 10, based on average peat growth of 2.5–3.5 cm/100 yr, and 5.5–6.5 cm/100 yr in Section 10 at the base of the slope) indicate the extent of the gap. Further unmarked erosion surfaces are suggested by hiatuses in podzol and charcoal stratigraphy. Continuing destabilization is indicated by slopewash, which interrupted the evolution of thin peaty soil and, most recently, formed the relatively thick deposit in which modern soils are developing. These observations suggest that mid-Holocene stability gave way to renewed erosion, this time decoupled from postglacial relaxation.

Although we found no other reports of radiocarbon dated, iron-humus palaeopodzols on sub-alpine, wet temperate, mountain slopes in NW Europe, they provide additional insights into environmental change. The height of the water-table on Creagan a' Chaorainn is controlled by variations in the presence of basal inorganic clays, till thickness and induration, and the height of the bedrock surface. In some palaeogullies, a high water-table combined with mid-Holocene seepage and persistent leaching promoted peat growth, but where clay and bedrock (hence the water-table) were lower, podzolization occurred at the same time. Soil profile development is closely related to persistent humidity and leaching in a temperate climate (Catt, 1979; Ellis and Matthews, 1984; Caseldine and Matthews, 1985; Avery, 1990; Knight, 1990; Gerrard, 1992; Matthews, 1993; Bain, *et al*., 1993; Schaetzl and Isard, 1996; Bech, *et al*., 1997; Haynes *et al*., 1998). It also requires hundreds of years, so each truncated podzol indicates a period of slope stability followed by an erosive event. From Figure 10, a millennium of slope-stabilizing vegetation growth was interrupted by a variety of erosive events at around 6.0 cal. ka BP – towards the end of the mid-Holocene climatic optimum as defined elsewhere in NW Europe.

Repeated podzolization occurred from before 7.4 cal. ka BP until after 1.6 cal. ka BP, although the record is fragmentary. The 10^2–10^3 yr timescale of profile development is incompatible with both frequent, intense, slope-destabilizing storms and continuous, minor erosion. Instead, podzolization, like the formation of basal peats on gully floors, requires prolonged humidity combined with a stable slope. Relatively high concentrations of Fe and Al between 6.2 and 5.2 cal. ka BP, and again in the late Holocene, in the older and younger podzols imply relatively prolonged or intense leaching associated with climatic deterioration. The percentage of aluminium in the Creagan a' Chaorainn podzols is comparable with that in the modern Cairngorm soil, but much lower than in the Cumbrian example (Table 6). A possible explanation is recent acidification at the former locality.

The conditions in which bedrock weathering occurred, producing alternating peat/rock debris couplets in the palaeogullies, requires explanation. Intense channelled water flow in the gullies would have eroded the thin peats on which rock debris was emplaced. However, no such erosion occurred. Furthermore, twigs and roots of trees and shrubs at the upper surfaces of the sediment-free peats on gully floors immediately below some interpeat debris flows, suggest debris transport on a stable slope with a closed vegetation cover. Weathering of the bedrock, which was the principal source of gully debris, may rather have occurred in the conditions of persistent high humidity that produced steady seepage, the spread of blanket peat and a largely stable till slope. Consequent high porewater pressures are likely to be the immediate cause of many debris flows (eg, Iverson, 1997; Major, 1997; Sandersen, 1997; Matthews *et al.*, 1997a,b; Major and Iverson, 1999).

In these circumstances, the generation of unstable debris on gully floors, and at gully heads upslope, can be related to weathering of highly fractured, water-saturated bedrock. Holocene weathering rates in non-glaciated areas of NW Europe have been the subject of few studies and are seldom invoked to account for slope sensitivity and evolution. However, if postglacial relaxation had ceased to dominate a still demonstrably active slope, they require consideration. On Creagan a' Chaorainn, transport of fractured bedrock in gullies (possibly diluted by some gully wall till) appears to have been infrequent until about 4.3 cal. ka BP, gaining momentum only from about 2.6 cal. ka BP.

In a persistently wet climate, positive feedbacks between surface weathering (which increases the surface area of bedrock, hence weathering rates) and increased water flow (which increases rock:water ion exchange in open fractures, hence near-surface weathering) are likely to have destabilized fault-shattered bedrock via both granular weathering of exposed surfaces and reduced shear strength in near surface zones. The time lag between the onset of enhanced weathering and the generation of sufficient debris to produce water-saturated flow is unquantified. But judging by the gap between the end of the mid-Holocene climate optimum and the start of more frequent rock debris transport on Creagan a' Chaorainn after about 4.3 cal. ka BP, it may have been $\geq 10^3$ years. If overall rates of Holocene weathering were slow in relation to early rates of paraglacial sediment redistribution, so that their impact before the late mid Holocene was negligible, this accounts for persistent humidity in the late Holocene having led to both rapid podzolization and enhanced bedrock weathering and erosion, with a millennial-scale lag before the latter gained momentum.

If the above arguments are correct, the case for anthropogenic influences on slope evolution on Creagan a' Chaorainn – and comparable natural slopes – is weak. The restricted time distribution of macroscopic charcoal-forming events (Table 4), and the weak association of vegetation burning with subsequent slope failure (Table 5), tend to support this view. From Table 5, debris flows overlie charcoal in only two out of seven sequences. Moreover, slope mass movement was frequent in periods when macroscopic charcoal was absent. Evidence for Creagan a' Chaorainn does not therefore include removal of woody vegetation (anthropogenic or otherwise) as an independent driver of slope instability. In the main, burned areas were rapidly recolonized and restabilized. An alternative explanation for the restriction of macroscopic charcoal to the mid Holocene is the coeval 'pine decline' in N Scotland reviewed by Bennett (1995) and Birks (1996) and attributed to widespread soil paludification. Speculatively, evidence for that pine decline on Creagan a' Chaorainn lies in the inverse correlation between the presence of the trees that produced macroscopic charcoal between 7.5 and 6.5 cal. ka BP, and the development of blanket peat in the cooler wetter conditions that followed.

The lack of substantial thicknesses of slopewash (fines derived from exposed till) before the time when modern soils began to establish, casts further doubt on early anthropogenic impacts. The 10–15 cm thick layer which forms the present-day soil matrix is unique in slope history, and points to unprecedented erosion of peat and other stabilizing vegetation only within the last few hundred years.

The variations in slope wetness through time (summarized in Table 8) indicate a persistent switch from drier to wetter surface conditions after about 2.7 cal. ka BP. Peat humification changes in Section 4 (samples 4/3, 4/4), and statistically indistinguishable ages from either side of the peat boundary, indicate a rapid rise in the water-table during which very poorly humified peat, devoid of woody fragments washed in from upslope, formed. The contemporary influx of fine sediment suggests enhanced run-off, breached vegetation cover, and erosion, as well as water-logging inimical to shrubs and trees. Pitting (etching) of some of the mineral grains in Sample 4/4 (Reid, 2001) invites explanation, although the solubility of quartz in the slope environment is unclear. At up to 25°C, quartz solubility in sedimentary rock is negligible at pH <9 (eg, Deer *et al.*, 1966, Tucker, 1986). However, a laboratory study by Bennett *et al.* (1988) found an 8–10 fold increase in quartz solubility in the presence of salicylic, oxalic and humic acids compared with its solubility in water of pH 7. Etching may therefore have occurred either during the Holocene in acid conditions beneath peat, or in a high pH environment during subglacial weathering of calcium-rich minerals in crystalline bedrock. In either event, etched grains confirm exposure of the mineral substrate.

This return to net erosion on Creagan a' Chaorainn after about 2.7 cal. ka BP marked a further system change characterized by till failures and debris flows in gullies. At some point after 1.3 cal. ka BP erosion appears to have intensified. Fines were released from exposed till, forming the matrix for modern soils, while run-off washed sand and silt into sumps where the raised water-table promoted continuous peat growth. The lack of dated samples immediately below the modern soils provides little basis for judging whether precipitation-driven erosion has persisted into modern times. However, minor slopewash at the base of the cliff section where till is exposed beneath thin, degrading peat, indicates that some till erosion is continuing. Heavy grazing and trampling by deer and sheep over recent decades has disrupted the peat cover and the modern gully, dating from the 1980s, and resulting from track drainage, is unambiguously anthropogenic. Gully widening over the last five years, together with headward erosion in some of the palaeogullies (Figures 3, 7), suggest that positive feedbacks are perpetuating the erosion. It remains unclear therefore, whether recent human impacts have simply exacerbated pre-existing erosive conditions, or whether this constitutes a phase of renewed, principally anthropogenic, erosion.

When paraglacial sediment redistribution, slope hydrology, bedrock weathering and the impacts of blanket peat on exposure of mobile sediment are considered together, phases of slope stability and mobility on Creagan a' Chaorainn can be related to climate shifts known to have affected the whole of NW Europe, in particular millennial-frequency incursions of cold, Arctic surface waters. Response times reflect increased precipitation of the different elements of the slope system (till failure, peat erosion, peat humification, bedrock weathering) are highly variable so that slope evolution is, overall, an imprecise proxy for climate change. It does, however, lend itself to the definition of several broad phases when evolving system sensitivity, combined with climate forcing, drove the system to find new equilibria. These are: a mobile slope, dominated by paraglacial sediment redistribution before 7.5 cal. ka BP; overall stability during the Holocene climatic optimum; precipitation- and weathering-driven renewal of instability after about 4.3 cal. ka BP, which gathered pace with intensifying climate deterioration after about 2.7 cal. ka BP; and anthropogenic erosion in the last few hundred years.

Conclusions

Early Holocene slope development at Creagan a' Chaorainn probably reflected a paraglacial adjustment to postglacial conditions (Church and Ryder, 1972; Ballantyne, 2002a,b). But throughout the Holocene, the north slope of Creagan a' Chaorainn has also evolved in response to changing climate and intrinsic controls on slope sensitivity. We deduce that paraglacial adjustment effectively terminated before 7.5 cal. ka BP, when peat cover began to establish. There has been no event comparable with the rock avalanche (dated to 3950 ± 320 ka) at Beinn Allegin, attributed by Ballantyne and Stone (2004) to persistent paraglacial stress release. Both before and since this time, slope stability and instability have shown a broad synchrony with fluctuations in Holocene climate documented from other sources in NW Europe. The long period of stability, during which peat progressively colonized palaeogully floors, ended with a number of debris flows at around 6.0 cal. ka BP. This is close in time to a southward incursion of the polar front in the eastern N Atlantic as logged by Bond *et al.* (1997). The absence of dated erosional events thereafter suggests stabilization of the slope during the warm, dry conditions of the Holocene climatic optimum. Conditions deteriorated after *c.* 4.5 cal. ka BP and there are signs of slope instability during the following millennium, starting at about 4.4 cal. ka BP when reactivation of slope failures occurred in Norway, as well as at a second site in the Scottish Highlands (Gleann Lichd, Reid, 2001). Recent tephra dating in Scottish peats has also confirmed a transition to wetter conditions post 4310 ka, Hekla-4 isochron (Langdon and Barber, 2004). From *c.* 4.0 to 2.7 cal. ka BP there is a lack of recorded activity, but after the latter date the slope appears to have become active again, perhaps reflecting the lag time inherent in producing unstable debris from weathered bedrock. This, too, corresponds with cooling of N Atlantic sea surface temperatures. It is unclear from available data whether mass movement on Creagan a' Chaorainn then diminished, only to resume in response to further sea surface cooling accompanied by deteriorating climate at about 1.3 cal. ka BP, or whether conditions remained sufficiently wet to induce relatively frequent slope failures throughout this period of time. After *c.* 1.0. cal. ka BP activity may have tailed off, but the data are limited. Further studies are needed to establish whether the extreme oceanicity of the climate affecting the mountains of N

Scotland has made for closer time-matching of slope response with sea temperature variations than in the more continental, alpine environments of Fennoscandia.

Holocene weathering and blanket peat appear to have played a significant role in changing slope sensitivity at this site. It can be argued that the slope was not as sensitive at the time of climatic deterioration around 4.5 cal. ka BP as it was in the late Holocene, and we therefore have fewer events recorded at this time. The spread of blanket peat may have amplified this effect.

Anthropogenic influence on these events remains unquantified. Although Davies and Tipping (2004) have found evidence for small-scale human activity after 4400 cal. yr BP in nearby Glen Affric, there is no strong argument or clear evidence for invoking land-use impacts prior to modern gullying and peat erosion at Creagan a' Chaorainn.

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Notes

This paper is published posthumously on behalf of Dr Elspeth Reid who was tragically killed in a road accident in 2003. All the data come from her PhD thesis (Reid, 2001) and most of the text revisions were agreed with Dr Reid before her death. Final versions of the figures have been produced with the help of Bill Jamieson, and brief references to some recent papers have been included in support of Dr Reid's own conclusions. It is hoped that the published paper fairly represents her scientific data and interpretations.

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