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Relict non-glacial surfaces in formerly glaciated landscapes

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Abstract

Relict non-glacial surfaces occur within many formerly glaciated landscapes and contain important information on past surface processes and long-term landscape evolution. Relict non-glacial surfaces are distinguishable from glacial surfaces by large-scale morphologies, including rounded summits, fluvial valleys, and cryoplanation terraces and pediments, and the presence of tors, blockfields, and/or saprolites. Preservation during glaciation occurs either through coverage by non-erosive, cold-based, ice or as nunataks. Although surface morphologies and denudation rates indicate a continuous non-glacial surface history since preglacial times, relict non-glacial surfaces are dynamic features that have evolved during the Quaternary. Depending on spatial variables such as lithology, slope, regolith depth and the abundance of fine matrix and water some surfaces are denuding very slowly, while others display more rapid denudation. High spatial variability in denudation rates results in changing surface morphologies over time. Denudation rates also display high temporal variability, with much surface evolution having perhaps occurred soon after the initial onset of glaciation or during paraglacial phases. While some parts of non-glacial landscapes are currently active, others may be largely inactive relicts of past higher energy regimes. Although non-glacial surfaces are dynamic much remains to be determined regarding surface denudation rates and the magnitude of morphological changes over time.
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Keywords: relict surface; glaciated landscape; denudation; blockfield; cold-based ice; nunatak

1. Introduction

Relict surfaces in present landscapes were initiated under different environmental and geomorphological conditions than those prevailing in the landscapes today (Battiau-Queney, 1996; Migoñ and Goudie, 2001). The presence of relict surfaces has been subject to debate due to a number of factors. These include the association of relict surfaces with historical geomorphology and evolutionary theories of landscape development, a research emphasis over recent decades on process geomorphology and modern process rates, and problems in accurately establishing landform antiquity through independent dating techniques (Brunsden, 1993; Twidale, 1997; Widdowson, 1997; Twidale, 1998, 1999; Whalley et al., 2004). However, relict surfaces have been identified in many landscapes, including those that have been formerly glaciated (Reusch, 1901a,b; Wråk, 1908; Ahlmann, 1919; Ives, 1958a,b; Gjessing, 1967; Sugden and Watts, 1977; Hall, 1985; Lagerbäck, 1988a,b; Nesje et al., 1988; Kleman, 1994; Rea et al., 1996b; Ballantyne et al., 1997; Kleman and Stroeven, 1997; Bierman et al., 1999; Lidmar-Bergström et al., 2000; Fabel et al., 2002; Hättestrand and Stroeven, 2002; Briner et al., 2003; Marquette et al., 2004; Briner et al., 2005; Sugden et al., 2005; Fjellanger et al., in press).

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Formerly glaciated landscapes contain relict surfaces of glacial, and sometimes non-glacial, origins. Relict glacial surfaces date primarily from the last glaciation but potentially also from earlier glacial phases (Lagerbäck, 1988b; Kleman, 1994). Conversely, relict non-glacial surfaces may have preglacial origins, with further development under ice-free conditions during interglacials, interstadials, or on nunataks (Kleman and Borgström, 1990; Rea et al., 1996b; Ballantyne et al., 1997). Relict non-glacial surfaces are most readily identified as periglaciated, low-relief, plateau surfaces separated from lower, more recent, glacial landscapes by sharply-defined boundaries, such as trough walls (Brunsden, 1993; Ollier and Pain, 1997; Stroeven and Kleman, 1999; Migoñ and Goudie, 2001; Anderson, 2002; Munroe, 2006).

This paper reviews relict non-glacial surfaces within formerly glaciated landscapes. However, non-glacial relict surfaces have essentially identical counterparts in landscapes beyond former ice sheet margins. Periglacial landscapes in arid low-relief parts of the Canadian high-Arctic and Siberia (French, 2000), and formerly periglaciated landscapes in southern England (Ballantyne and Harris, 1994, p. 37; Migoñ and Goudie, 2001), central Europe (French, 2000), and parts of the Appalachian mountains in North America (Clark and Ciokosz, 1988), are morphologically and genetically analogous to the periglaciated non-glacial relict surfaces reviewed in this paper. The primary distinguishing feature of periglaciated relict surfaces in unglaciated landscapes beyond former ice sheet margins is that they have not been fragmented by glacial processes.

An issue with the term "relict" relates to the continuing modification of surfaces over time. Consequently, surfaces are the product of a number of dominant processes and climatic regimes. A polygenetic history blurs the definition of a relict surface and introduces a scale dependency. For example, while the gross morphology of a surface may show preglacial inheritance, the surface regolith displays Quaternary development, with imprints of Holocene geomorphology (Fig. 1). Landscape components of varying scale evolve at different rates and so have different "lifetimes" (Brunsden, 1993; Widdowson, 1997), with smaller-scale features, in particular, more likely to be short-lived and recent.

The continuing evolution of surfaces is also problematic for the term "preglacial," which is a frequently used label for relict non-glacial surfaces (Glasser and Hall, 1997; Kleman and Stroeven, 1997; Stroeven and Kleman, 1999). "Preglacial" is analogous to Migoñ and Goudie's (2001) concept of an "inherited" landscape and implies surface survival with minimal Quaternary modification (Fig. 2a). In some instances, however, relict non-glacial surfaces may exhibit dynamic equilibrium, where preglacial morphology is retained but with significant surface lowering (Fig. 2b; Gilbert, 1909; Hack, 1975; Brunsden, 1993; Small et al., 1999). Such surfaces can be considered "persistent" as preglacial morphology is maintained but current surface features are not necessarily old (Brunsden, 1993). In other instances, continuing evolution may result in morphological changes in addition to surface lowering (Fig. 2c; Anderson, 2002). Current surfaces perceived as preglacial do not then necessarily replicate the true preglacial morphology.

The difficulty in using "preglacial" to describe current surfaces is further highlighted by the possibility of Neogene glaciation of some formerly glaciated landscapes (Jansen and Sjøholm, 1991; Hölemann and Henrich,

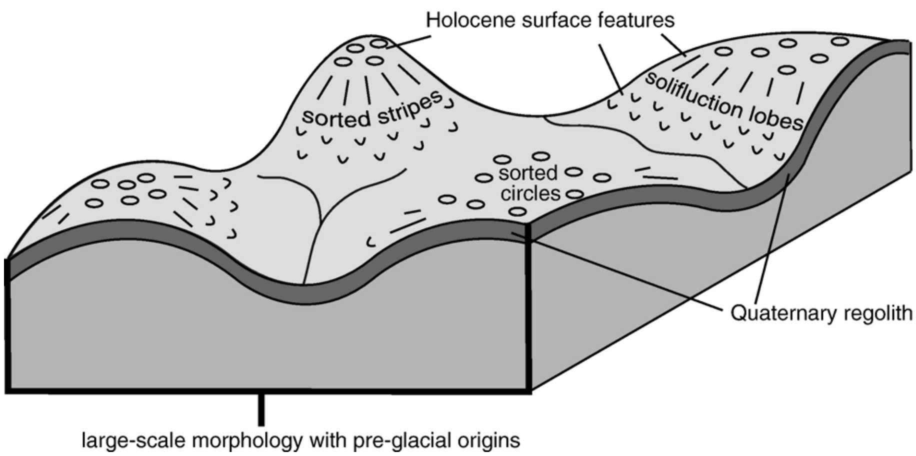

Fig. 1. Relict non-glacial surface with a polygenetic history. Large-scale morphology with preglacial inheritance is mantled by Quaternary regolith, with imprints of Holocene geomorphology.

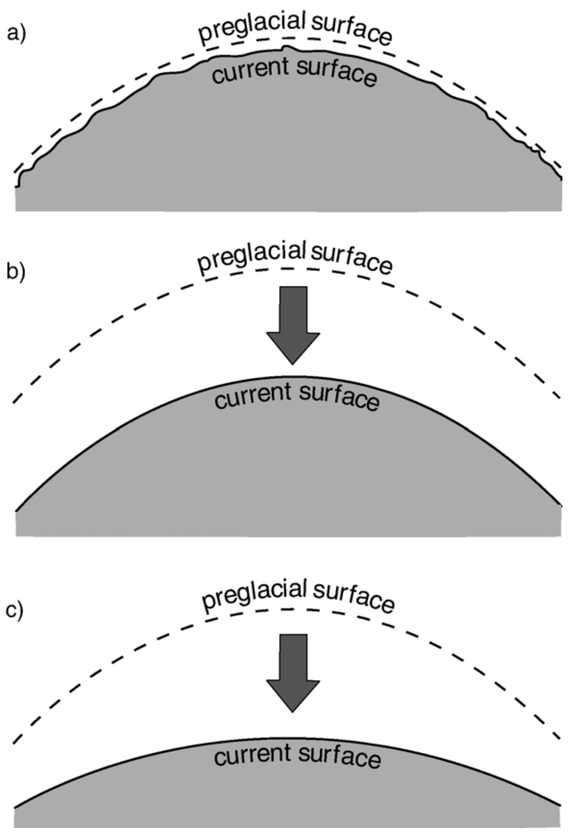

Fig. 2. Three possibilities for preglacial surface evolution through non-glacial processes: a) minimal denudational or morphological change, i.e. an "inherited" surface; b) dynamic evolution, where a surface is significantly denuded but morphology is maintained i.e. a "persistent" surface; c) significant denudation and morphological change.

1994; Mangerud et al., 1996). In such instances, a surface existing at the beginning of the Quaternary does not represent a preglacial surface, and the possibility of an extended history of major climatic fluctuations, including periodic glaciations, increases the likelihood of extensive surface modification. As true preglacial surfaces may not exist in the present landscape "non-glacial" is used here for relict surfaces that are clearly not of glacial origin.

Relict non-glacial surfaces are important for:

1. Inferring aspects of regional glacial history, such as basal thermal regimes, glacial erosion patterns, and thicknesses of former ice sheets (Hall and Sugden, 1987; Nesje et al., 1988; Sugden, 1989; Kleman, 1994; Ballantyne et al., 1997; Kleman and Stroeven, 1997; Bierman et al., 1999; Clarhäll and Kleman, 1999; Kleman et al., 1999; Briner et al., 2003; Marquette et al., 2004; Briner et al., 2005; Sugden et al., 2005; Linge et al., 2006; Phillips et al., 2006; Fjellanger et al., in press). The spatial distribution of relict non-glacial surfaces may also reveal information on glacial period climates (Munroe, 2006).
2. Determining processes and rates of surface evolution (Small et al., 1999; Anderson, 2002). Currently, ages of relict non-glacial surfaces are poorly constrained and little is known of their Quaternary evolution.
3. Reconstructing preglacial landscapes to provide a reference by which to quantify glacial erosion (Nesje and Whillans, 1994; Glasser and Hall, 1997; Kleman and Stroeven, 1997, Stroeven et al., 2002a). The Quaternary evolution of relict non-glacial surfaces should be accounted for in reconstructions of preglacial surfaces (Fig. 2).
4. Potentially determining interstadial, interglacial, and/or preglacial environmental conditions from, for example, fine matrix materials contained within relict non-glacial surface regolith (Hall, 1985; Rea et al., 1996b). Formerly glaciated landscapes currently occupy a range of climatic zones including polar (Stroeven and Kleman, 1999; Sugden et al., 2005), maritime (Whalley et al., 1981; Paasche et al., in press), continental (Small et al., 1999; Munroe, 2006), alpine (Dahl, 1966; Rea et al., 1996a,b), and cool temperate (Hall, 1986; Olvmo et al., 2005). Most continue to be subject to a periglacial regime (Kleman and Stroeven, 1997; Dredge, 2000; Marquette et al., 2004; Munroe, 2006). Knowledge of preglacial climates is limited with a number of studies speculating on warm humid conditions (Roaldset et al., 1982; Hall, 1986; Rea et al., 1996a,b; Marquette et al., 2004; Paasche et al., in press). The palaeoclimatic evidence maintained by relict non-glacial surfaces may vary between sites according to differences in denudation histories.

Relict non-glacial surfaces are reviewed here under five main topics. Firstly, relict non-glacial surfaces are shown to be identifiable through characteristic assemblages of morphological features. These features can also be used to interpret weathering histories, denudation rates, and ages of relict non-glacial surfaces. Secondly, preservation of relict non-glacial surfaces during glacial periods is shown to be a consequence of interactions between ice sheet and landscape factors that permit either formation of non-erosive, cold-based, ice or surface survival as nunataks. Thirdly, denudation rates of relict non-glacial surfaces are shown to be low, but with potentially significant spatial variation. Consequently, and discussed as topic four, some relict non-glacial surfaces may have exhibited considerable morphological evolution during the Quaternary. The review concludes with a discussion on how relict non-glacial surfaces can be used in reconstructing preglacial surfaces and quantifying glacial erosion. The central theme of this paper is that relict non-glacial surfaces are, both spatially and temporally, dynamic features.

2. Identifying and interpreting relict non-glacial surfaces in formerly glaciated landscapes

2.1. Large-scale morphology

Certain large-scale morphologies provide evidence of both the non-glacial and relict nature of surfaces. Typical non-glacial surface morphologies include symmetrical convex summits, formed by diffusive slope processes such as soil creep (Kaitanen, 1969; Carson and Kirkby, 1972; Fernandes and Dietrich, 1997), and V-shaped valleys of fluvial origin (Fig. 3; Kleman and Stroeven, 1997; Fabel et al., 2002; Stroeven et al., 2002a). Cryoplanation terraces and pediments may also feature, although their genesis and significance remain poorly understood (French and Harry, 1992; Czudek, 1995; Thorn and Hall, 2002). Due to their non-glacial appearance and presumed long periods of formation, landscapes containing the above features are interpreted as having preglacial origins (Glasser and Hall, 1997; Kleman and Stroeven, 1997, Stroeven and Kleman, 1999). However, non-glacial surfaces have continued to evolve during the Quaternary (Small et al., 1999; Anderson, 2002), although perhaps by different processes and at different rates to those that prevailed during their initiation. Furthermore, fluvial valleys can be carved into non-glacial surfaces during interstadial and/or interglacial periods to new base levels imposed by the formation of glacial troughs (Wråk, 1908; Ahlmann, 1919; Rudberg, 1992; Kleman and Stroeven, 1997; Stroeven et al., 2002a). Consequently, surfaces with non-glacial large-scale morphologies are not entirely preglacial features.

2.2. Tors

Tors are meso-scale morphologic features of relict non-glacial surfaces defined as bedrock outcrops exhibiting extensively weathered joints and protruding with vertical faces above their immediate surroundings (Hättestrand and Stroeven, 2002). Occurring most commonly at hill summits and at breaks in slope (Kaitanen, 1969; Eden and Green, 1971; Wang et al., 1981; Ballantyne, 1994; Anderson, 2002; Phillips et al., 2006), tors are a feature of virtually all climatic environments (Ballantyne, 1994). Tors form either through continuous processes of weathering and stripping (Palmer and Radley, 1961) or are a product of two-stage evolution, where stripping and exposure have followed a period of deep weathering (Linton, 1955). As tors evolve slowly and are easily altered by glacial erosion (Hättestrand and Stroeven, 2002) their presence within formerly glaciated landscapes indicates surfaces that have, at most, been subject to only partial glacial

Fig. 3. View looking south from Tarfala ridge, northern Sweden, displaying typical large-scale relict non-glacial surface morphology, with symmetrical rounded summits and a fluvially incised valley. Note also the blockfield mantle. (Photograph: Clas Hättestrand).

modification and maintain relict non-glacial components (Linton, 1959; Sugden, 1968; Kaitanen, 1969; Sugden and Watts, 1977; Söderman et al., 1983; Kaitanen, 1985; Ballantyne, 1994; Rapp, 1996; Kleman and Stroeven, 1997; Hättestrand and Stroeven, 2002; André, 2004). In many instances tors are considered as preglacial forms (Sugden, 1968; Kleman and Stroeven, 1997; Bierman et al., 1999; Hättestrand and Stroeven, 2002; André, 2004).

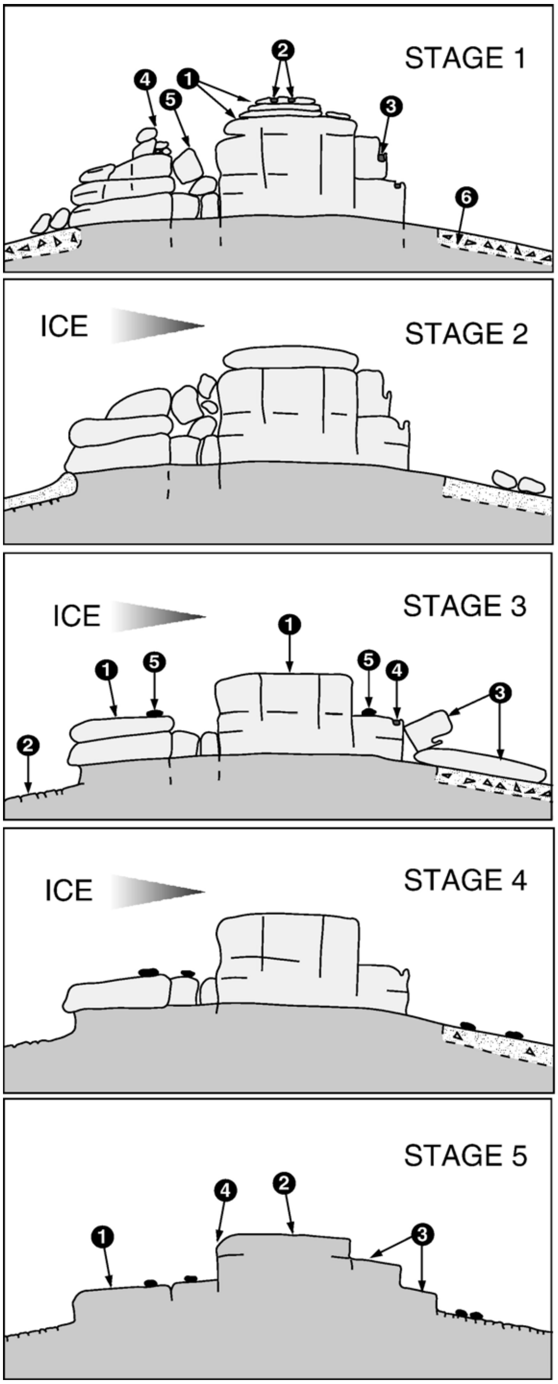

Fig. 4. Sequence of tor forms displaying progressively greater glacial modification. In stage 1, the tor is unmodified by glacial erosion, with features including: 1) widening of exfoliation joints and rounding of block edges; 2) weathering pits on horizontal surfaces; 3) weathering pit spillways on vertical surfaces; 4) perched blocks; 5) toppled blocks; 6) regolith adjacent to tor. In stage 2, the tor is inundated by cold-based ice. Although the ice sheet is frozen to the ground, internal ice deformation occurs, as indicated by the arrow, transporting detached blocks in a down-flow direction. In stage 3, flowing ice continues to modify the tor. Features include: 1) unweathered surfaces with no weathering pits; 2) removal of regolith; 3) boulder train of transported blocks; 4) local preservation of weathered surface with weathering pits; 5) erratics overlying surface with and without weathering pits. In stage 4, the tor is reduced to an upstanding plinth that resists further erosion by cold-based ice. In stage 5, wet-based, sliding basal ice further erodes the tor to a low bedrock slab. Features include: 1) erratics over the slab, which lacks open expansion joints, and a lack of lee-side regolith; 2) abraded surface lacking weathering pits; 3) lee-side plucking; 4) stoss-side abrasion (source: Fig. 2 in Phillips et al., 2006, with permission of Elsevier and the authors).

Although tors indicate the likely presence of relict non-glacial surfaces, the potential for modification by glacial erosion necessitates careful interpretation of their origins, ages, and possible significance as markers of relict surfaces and of site-specific glacial erosion depths (Hall and Phillips, in press). The presence of notches on tor-sides, representing former ground surfaces, indicates periods of stability followed by stripping, while the depth of joints and weathering pits can be used to infer both the antiquity of tors and the degree of glacial modification (Hall and Phillips, 2006, in press). Increasing depths of joints and weathering pits signify greater tor ages, with the absence of glacial modification also indicated by *in situ* crowning boulders. Where glacial modification has occurred, tors display varying forms according to the degree of modification, from slightly modified tors missing crowning blocks to significantly modified tor roots (Fig. 4; André, 2004; Phillips et al., 2006; Hall and Phillips, in press). Tor-like features may even occur in non-relict glacial landscapes with, for example, roches moutonnées possibly being glacially sculptured tors or precursor tor forms exhumed from the weathering front and modified by glacial processes (Sugden, 1968; Lindström, 1988; Lidmar-Bergström, 1989; André, 2004; Phillips et al., 2006). Consequently, inferring a non-glacial surface requires consideration of tor morphology in addition to other evidence, including over-all landscape morphology and/or blockfield covers.

Analysis of terrestrial cosmogenic nuclides (TCNs) has illustrated the antiquity of tors in formerly glaciated landscapes of North America (Bierman et al., 1999), Scotland (Phillips et al., 2006), and Sweden (Stroeven et al., 2002b). Due to complex exposure histories including possible periods of burial beneath ice sheets and the likelihood of episodic erosion (again possibly by ice sheets), interpreting tor lowering rates and ages requires careful consideration (Small et al., 1997; Bierman et al., 1999). Taking complex exposure histories into account, Bierman et al. (1999) calculated a minimum 450 ka of surface and near-surface exposure for Baffin Island tors, while Phillips et al. (2006) and Stroeven et al. (2002b) estimated minimum model histories of up to 626^{+102}_{-85} ka and 605 ka for tors on the Scottish Cairngorm Mountains and in northern Sweden, respectively. Phillips et al. (2006) summarised TCN analyses from a number of studies as indicating tor survival for periods of 200 to 845 ka, which confirms tor persistence over multiple glacial cycles but also illustrates a Quaternary, and not preglacial, history of tors as surface and near-surface features. Tors are dynamic features that continue to evolve during ice-free periods, with regolith stripping and bedrock erosion rates being too high to allow the extensive preservation of pre-Quaternary rock surfaces (Phillips et al., 2006).

2.3. Blockfields

The most common indicator of a relict non-glacial surface is a veneer of coarse regolith known as a blockfield. Sometimes termed felsenmeer or boulder fields, blockfields are most frequently associated with mountain summits and plateaus (Rea et al., 1996b; Ballantyne, 1998; Whalley et al., 2004) but can also occur at low elevations (Piirola, 1969; Hättestrand, 1994; Hättestrand and Stroeven, 2002). While blockfields are considered a key indicator of relict surfaces, there are various interpretations as to how they are formed, their rates of formation, and subsequent ages. Blockfields have been used as paleo-indicators of nunataks (Nesje, 1989; Ballantyne et al., 1998), non-erosive ice covers (Sugden and Watts, 1977; Kleman and Borgström, 1990; Hättestrand, 1994; Hättestrand and Stroeven, 2002), and local climate (Nesje, 1989; Ballantyne, 1998; Ballantyne et al., 1998).

There are two broad classes of blockfields. Autochthonous blockfields are formed *in situ* from weathered underlying bedrock (Nesje et al., 1988; Ballantyne et al., 1998) and are of primary importance for identifying relict non-glacial surfaces. Allochthonous blockfields are formed either from upfreezing of clasts in till or by the downslope movement of weathered material (Potter and Moss, 1968; Nesje et al., 1988; Ballantyne et al., 1998; van Steijn et al., 2002; Boelhouwers, 2004; Sumner and Meiklejohn, 2004). The significance of allochthonous blockfields in identifying relict, non-glacial or glacial, surfaces remains uncertain although examples formed from till, termed “boulder depressions” and pre-dating the last glaciation, have been identified in central Sweden (Kleman and Borgström, 1990; Hättestrand, 1994). Resistant lithologies are essential to both forms of blockfields (Ballantyne, 1998; Boelhouwers, 2004; Fjellanger et al., in press) and distinguishing the two classes can be difficult (Rea et al., 1996b).

Autochthonous blockfield formation has traditionally been associated with frost shattering of bedrock under *severe* periglacial conditions (Svenonius, 1909; Lundqvist, 1944; Ives, 1958b; Dahl, 1966; Kleman and Borgström, 1990; Ballantyne, 1998; French, 2000). Frost shattering of porous materials (Walder and Hallet, 1985, 1986) and frost wedging of pre-fractured material has been demonstrated (Ballantyne, 1998; Boelhouwers, 2004). Substantially colder ground conditions may occur in coarse frost-wedged materials producing a positive feedback effect on further frost wedging (Harris and Pedersen, 1998; Sawada et al., 2003; Boelhouwers, 2004; Gorbunov et al., 2004). Some studies have downplayed the necessity of severe periglacial conditions for block formation from frost weathering by stating that it is the number of cycles around zero that is important rather than the intensity or duration of freezing (Wiman, 1963; Potts, 1970; Whalley and McGreevy, 1987). Sumner and Meiklejohn (2004) interpret limited blockfield formation at high altitudes on the sub-Antarctic Marion Island as a consequence of colder summit conditions.

The general role of frost shattering in blockfield formation has, however, been challenged (White, 1976; van Steijn et al., 2002; Boelhouwers, 2004). The presence of blockfields in arid, non-periglacial, environments, such as the Namibian desert (van Steijn et al., 2002; Boelhouwers, 2004) and the Sierra Nevadas of California (Friend et al., 2000), and the lack of evidence for the efficacy of frost to shatter intact crystalline bedrock (Lautridou and Seppälä, 1986; André, 2004; Whalley et al., 2004), indicate that processes other than frost shattering can be responsible for blockfield formation. Possible alternative mechanisms include thermal stress and thermal shock resulting from dramatic diurnal and seasonal variations in rock surface temperatures in polar, sub-polar, alpine, and desert environments (Hall, 1999; Lewkowicz, 2001; Hall et al., 2002; Hall and André, 2003; Sumner et al., 2004). However, the potential effects of thermal shock are restricted to a very shallow zone close to the rock surface and whether thermal stress can overcome the tensile strength of the rock remains uncertain (Boelhouwers, 2004). Regardless, autochthonous blockfields should not be viewed as necessarily periglacial features (French, 2000).

Assuming erosive ice sheets and/or high rates of frost shattering under periglacial conditions, a number of early studies speculated on a Holocene origin for autochthonous blockfields (Svenonius, 1909; Lundqvist, 1944; Rapp and Rudberg, 1960; Dahl, 1966). Recent studies of blockfields on well-jointed lithologies, such as quartzite and microgranite, indicate formation under severe periglacial conditions immediately following deglaciation (Ballantyne, 1998; Ballantyne et al., 1998), while blockfields on both frost-susceptible carbonate bedrock and frost-resistant granitoid and gneissic bedrock are currently active in the Canadian Arctic (Dredge, 1992, 2000).

In most instances, however, blockfield development has been associated with pre-Holocene frost weathering of bedrock and survival of at least one glaciation (Lautridou and Seppälä, 1986; Rea et al., 1996b; Sumner and Meiklejohn, 2004; Whalley et al., 2004). Some studies have speculated on blockfield development over several interglacials with survival beneath cold-based ice sheets during glaciations (Nesje et al., 1988; Kleman and Borgström, 1990). Alternatively, blockfields may have developed from frost shattering of bedrock on nunataks during glacial periods (Ballantyne and Harris, 1994, p. 186; Ballantyne, 1998; Ballantyne et al., 1998). Other studies have stated a much older, pre-Pleistocene, origin of autochthonous blockfields (Dahl, 1961; Caine, 1968; Ives, 1974; Clapperton, 1975; Roaldset et al., 1982; Dahl, 1987; Nesje et al., 1988; Rea et al., 1996b; Boelhouwers et al., 2002; André, 2004; Whalley et al., 2004; Fjellanger et al., in press).

Support for the antiquity of autochthonous blockfields has been found in the following evidence.

1. The current inactivity of most reported examples of autochthonous blockfields (Boelhouwers, 2004). Even at polar sites with available moisture and relatively warm rock temperatures blockfield development is currently very slow (Sumner and Meiklejohn, 2004).
2. The presence of rounded surface clasts and angular sub-surface clasts, indicating greater surface weathering and therefore stability over a long period (Ballantyne, 1998); an interpretation frequently supported by the presence of large lichen specimens on surface blocks (Nesje et al., 1988) or contrasts in weathering rinds between the upper and lower surfaces of blocks (Caine, 1968).
3. Very slow weathering rates of intact bedrock, as indicated by minimal post-glacial weathering of glacially modified bedrock and erratics (Lautridou and Seppälä, 1986; Nesje et al., 1988; Ballantyne, 1998; Bierman et al., 1999; Fabel et al., 2002; Briner et al., 2003; André, 2004; Li et al., 2005), and laboratory experiments of weathering on pre-Cambrian blocks (Lautridou and Seppälä, 1986).
4. The non-correlation of summit blockfields in formerly glaciated regions with modern temperature conditions (Nesje et al., 1988).
5. The presence of large quantities of silt- and clay-sized material in interstitial fines may indicate extended periods of weathering, involving also chemical processes (Rea et al., 1996b; Ballantyne, 1998; Allen et al., 2001; Marquette et al., 2004).
6. The association of blockfields with meso-scale and large-scale landscape features, which are also thought to signify relict non-glacial surfaces (Rea et al., 1996b; Kleman and Stroeven, 1997; Fabel et al., 2002).
7. Preservation beneath former ice sheets (Section 3.1).

Perhaps the most significant evidence relating to autochthonous blockfield development and potential ages is that, with the exception of carbonate blockfields in the Canadian Arctic (Dredge, 1992), there is no current formation of new blockfields from exposed bedrock (Boelhouwers, 2004). This includes environments, such as sub-Antarctic islands, that should be favourable to blockfield formation from either frost shattering or thermal effects (Sumner and Meiklejohn, 2004). Blockfield development is apparently dependent upon past weathering that likely occurred under different climatic conditions than those currently prevailing in formerly glaciated landscapes. Blockfield formation has been commonly attributed to break-up of bedrock that was chemically weakened during warmer preglacial conditions, followed by upfreezing of blocks to the surface under periglacial conditions (Caine, 1968; Rea et al., 1996b; Whalley et al., 2004).

Central to possible preglacial blockfield origins is the presence of large quantities of silt- and clay-sized material, containing secondary clay minerals. The extensive occurrence of clay minerals confirms chemical weathering, with assumed long periods and warm climates required for the accumulation of clays and depth of blockfields, typically around 1 m, implicating preglacial weathering (Rea et al., 1996b; Marquette et al., 2004; Whalley et al., 2004). For example, in a study of blockfields in northern Norway, Whalley et al. (1997, 2004) recorded substantial quantities of vermiculite, kaolinite, and gibbsite, with clay and silt composing 5–20% and 15–50% of fine matrix, respectively. Conversely, some laboratory experiments have shown that freeze-thaw weathering does not produce substantial quantities of fine material (Rea et al., 1996b; Whalley et al., 2004) and while current chemical action may be sufficient to produce weathering rinds and clay minerals, it is unlikely to account for whole blockfield thicknesses (Whalley et al., 1997).

Although the presence of substantial quantities of silt and clay forms the major argument for preglacial blockfield origins, the assertion that chemical weathering and fine matrix formation has been negligible in periglaciated landscapes during the Quaternary, and that blockfield development is therefore dependent upon preglacial weathering, remains questionable. According to laboratory experiments by Smith et al. (2002), frost weathering can produce substantial quantities of silt. Also, based on a maximum rate of fines production from frost weathering experiments by Lautridou and Seppälä (1986), a 0.75 m deep profile could be produced in 10 000 years (Rea et al., 1996b). Furthermore, chemical weathering, which results in the production of solutes, silt, and/or clay, is potentially significant in periglacial alpine and arctic environments (Reynolds and Johnson, 1972; Ugolini and Anderson, 1973; Darmody et al., 1987; Anderson et al., 2000; Smith et al., 2002; Dixon and Thorn, 2005). Microclimates and the micro-scale presence of water are important for chemical weathering, with water binding tightly to mineral surfaces to be ubiquitous in all terrestrial weathering surfaces (Pope et al., 1995). The potential then exists for chemical weathering to occur almost anywhere, including cold periglaciated environments. Chemical weathering rates are high in "weathering-limited" periglacial regions with thin soils (Bluth and Kump, 1994) and solute contents of streams studied in periglacial areas do not inherently differ from those in other climatic environments (Dixon and Thorn, 2005). It is also unlikely that frost weathering occurs in isolation from chemical weathering as both physical and chemical processes are dependent upon the presence of heat and water (Whalley et al., 2004; Dixon and Thorn, 2005), along with textural properties of bedrock and micro-scale or meso-scale structural weaknesses (Battiau-Queney, 1996; Whalley et al., 2004). Quaternary chemical rates may be high enough to make a significant contribution to fine matrix volumes and blockfield development, particularly during warm interglacials (Whalley et al., 2004). However, chemical weathering rates remain uncertain and much is still to be determined regarding the role of Quaternary chemical processes in blockfield development.

It is also questionable whether particular clay minerals are necessarily representative of a warmer preglacial climate (Bird and Chivas, 1988; Bouchard and Jolicoeur, 2000; Migoñ and Lidmar-Bergström, 2002). Although gibbsite has been used to indicate blockfield initiation during a warmer preglacial climate (Roaldset et al., 1982, Rea et al., 1996b; Marquette et al., 2004; Paasche et al., in press), its paleoclimatic implications are controversial (Ballantyne, 1998). Gibbsite may form in alpine climates (Reynolds, 1971) or during interglacials (Hall et al., 1989) and is currently forming on tors in the northern Appalachians (Wang et al., 1981). Gibbsite formation occurs in the first stages of weathering, and under present climatic conditions, in some formerly glaciated landscapes (Wang et al., 1981; Hall et al., 1989; Bouchard et al., 1995). Kaolinite has also been used to infer blockfield initiation during a warm preglacial climate (Roaldset et al., 1982; Paasche et al., in press). However, oxygen isotope studies indicate kaolinite formation in Australia under cold and cool temperate climates following the Permian glaciation (Bird and Chivas, 1988, 1989). Vermiculite, smectite, and kaolinite formation has been observed in regolith beneath snow patches in the Rocky Mountains (Thorn et al., 1989), with vermiculite formation also observed on the Jotunheimen plateau, Norway (Darmody et al., 1987). Clay minerals are not, therefore, necessarily indicative of particular climates, although quantities may support climatic and age inferences (Bouchard et al., 1995; Migoñ and Lidmar-Bergström, 2002; Whalley et al., 2004).

The apparently negligible process rates currently occurring on most blockfields have been attributed to negative feedback effects resulting from blockfield development. Boelhouwers (2004) considers that a coarse surface layer will protect underlying materials from thermal stress fatigue, with cold conditions helping to preserve surface detritus by limiting chemical weathering, rather than promoting its formation (Sumner and Meiklejohn, 2004). Blockfield thickness also exerts a negative feedback through limiting weathering as depth increases (Ahnert, 1977; Rea et al., 1996b). A lack of surface water and streams also protects blockfields from erosion, as observed in drylands (Twidale, 1997). The presence of blockfields therefore protects formerly glaciated landscapes from weathering and erosion, similar to the protection of surfaces by coarse debris on desert pavements (Twidale and Bourne, 2000).

While current process rates on most blockfields appear to be very low and it is likely that, at least some, blockfields have preglacial origins, there remain many questions regarding blockfield formation. Rea et al.

(1996b) and Whalley et al. (2004) state that prolonged Tertiary weathering could have produced blockfield depths (~1 m) seen today on plateau surfaces in northern Norway. However, are erosion rates so low that the Quaternary has not provided enough time to erase shallow remnants of preglacial weathering profiles? If surfaces veneered with blockfields have lowered during the Quaternary, how and when does this occur? One possibility is that blockfields are periodically active, perhaps during paraglacial phases immediately following deglaciations (Caine, 1968; Ballantyne, 2002), when there is large supply of meltwater. Another possibility is that some blockfields may show current sub-surface activity. Although Whalley et al. (1997) state that sub-surface weathering would be unlikely due to insufficient temperature change, openwork blockfields may permit greater temperature fluctuations at depth than previously recognised and chemical weathering may be possible due to the ubiquitous presence, at least seasonally, of liquid water at or near blockfield bases. Moisture is considered a greater limitation on chemical weathering than temperature (Whalley et al., 2004) and to be effective it must be retained in contact with the weathering front (Ahnert, 1977; Anderson, 2002; Phillips, 2005). This occurs at blockfield bases, particularly where underlying permafrost forms an impermeable layer.

Stratigraphic studies of blockfields are few and, due to frost churning of profiles, have often assumed the absence of fine matrix zonation (Rea et al., 1996a; Whalley et al., 1997). Perhaps, where deep frost churning is relict, there is some zonation at depth within fine matrix materials, similar to that observed by Caine (1968), which may indicate Quaternary chemical weathering. One, or a few, cycles of frost wedging may also occur annually at blockfield bases, particularly in lithologies that are pre-cracked due to pressure release. Anderson (1998) suggested that under the proper annual temperature regime, frost crack growth might be more rapid in the sub-surface than at, or near, the surface. Chemical attack of the cracks then follows, with an associated feedback on the mechanical strength of rocks (Anderson, 2002). Where blockfield surfaces appear inactive, frost wedging may be occurring at depth with upfreezing occurring during colder, but ice-free, periods. The blocks upfrozen to the surface are remnants that are highly resistant to further weathering, giving the impression of inactive blockfields. Solutes and fines may be washed through the profile at depth (Caine, 1968; Potter and Moss, 1968; Strömquist, 1973), contributing to surface lowering and subsequently maintaining activity at the weathering front by keeping it near the surface (Phillips, 2005). Where pits have been excavated, frost wedging of bedrock has been indicated with no evidence of truncated, deep-weathered saprolite (Ballantyne, 1998).

There is a clear need for quantitative blockfield studies to determine the relative contributions of Quaternary and pre-Quaternary processes (van Steijn et al., 2002; Boelhouwers, 2004) and just how much present block-mantled landscapes replicate preglacial landscapes. There is an absence of field data on blockfield temperature and moisture conditions (Hall et al., 2002) and TCN analysis of vertical blockfield profiles, such as undertaken by Small et al. (1999), may provide information on blockfield erosion and regolith production rates over millennial timescales. Rates of chemical weathering also remain unclear (Whalley et al., 2004) and detailed studies of blockfields at different sites are required to test the dependence upon Neogene deep weathering. Blockfields are of varying age and significance (Ballantyne, 1998) and are likely to be the product of complex and diverse climatic histories with recent processes potentially strongly influenced by preglacial processes (Battiau-Queney, 1996; Boelhouwers, 2004; Whalley et al., 2004).

2.4. Saprolites and grusses

Saprolite is material derived from *in situ* isovolumetric weathering of rock where primary fabric and structure of the parent rock are retained (Bouchard et al., 1995; Battiau-Queney, 1996). As deep saprolites form over long periods and are widely considered to develop best in climates warmer than those that have prevailed in formerly glaciated landscapes during the Quaternary (Hall, 1985; Hillefors, 1985; Lundqvist, 1985; Hall et al., 1989; Wright, 1997; Islam et al., 2002), their presence can indicate surfaces that maintain preglacial inheritance (Bouchard, 1985; Lundqvist, 1985; Peulvast, 1985; Hall and Mellor, 1988; Lidmar-Bergström, 1989; Wright, 1997; Olvmo et al., 2005). Frequently, in formerly glaciated landscapes, saprolites may be the only way to identify relict surfaces (Bouchard, 1985; Hall, 1985; Lidmar-Bergström et al., 1997), with the depth and frequency of saprolite occurrence being inversely proportional to the intensity of glacial erosion (Hall, 1986; Hall and Sugden, 1987). Saprolites in formerly glaciated landscapes may be found at higher elevations, on plateaus (Peulvast, 1985) and interfluves (Battiau-Queney, 1984; Smith and McAlister, 1987; Bouchard and Pavich, 1989), and at lowland sites (Tynni, 1982; Hall, 1986; Hirvas et al., 1988; Hirvas, 1991; Lidmar-Bergström et al., 1997), particularly on the lee-sides of hills and scarps that have been protected from full glacial erosion (Bouchard, 1985; Hillefors, 1985; Kejonen, 1985; Lahti, 1985; LaSalle et al., 1985; Lehmuspelto, 1985; Lundqvist, 1985; Hall and Sugden, 1987; Olvmo et al., 2005).

Preglacial origins of saprolites can be confirmed in a number of ways. Geomorphic evidence, including an absence of signs of glacial erosion from saprolite sites (Hall and Sugden, 1987), the absence of saprolite in glacial troughs (Hall and Mellor, 1988) or on glaciated shield rocks (Bouchard et al., 1995), the presence of saprolites in sites protected from glacial erosion (Bouchard, 1985; Peulvast, 1985), and the incorporation of saprolite in till (Hall and Sugden, 1987; Wright, 1997; Lidmar-Bergström et al., 1999; Islam et al., 2002), indicates the existence of saprolite prior to glaciation. Mineralogy can also provide evidence for the preglacial existence of saprolites. Where saprolites are covered by till, they show different mineralogic composition and more advanced weathering than the overlying till. Gibbsite (Roaldset et al., 1982; LaSalle et al., 1985), halloysite (Hall and Mellor, 1988), kaolinite (Hall et al., 1989; Paasche et al., in press), and haematite (LaSalle et al., 1985; Hall et al., 1989; Lidmar-Bergström et al., 1997) have all been used to infer preglacial origins of saprolites. Hydromorphic iron and manganese minerals that were formed in poorly-drained conditions but now occur in freely-drained conditions indicate significant lowering of the groundwater table caused by glacial erosion and have preglacial origins (Hall, 1985; Hall et al., 1989). Furthermore, there is no evidence that saprolites 10–15 m deep can form during interglacial periods in formerly glaciated landscapes (Hall et al., 1989). Saprolites are generally reliable indicators of surfaces with preglacial origins and highlight limited, spatially variable, glacial erosion in many areas (Lundqvist, 1985; Peulvast, 1985; Söderman, 1985; Hall, 1986; Hall and Sugden, 1987; Rutherford, 1989; Sugden, 1989; Lidmar-Bergström, 1997). Non-glacial erosional processes operating during ice-free periods have also been insufficient to completely remove the preglacial regolith.

The characteristics of saprolites and associated grusses (disaggregated bedrock showing varying degrees of, but incomplete, chemical alteration; Dixon and Thorn, 2005) have been used as indicators of relief development (Hall, 1986; Pavich, 1989; Lidmar-Bergström, 1995; Battiau-Queney, 1996; Bouchard and Jolicoeur, 2000; Migoñ and Lidmar-Bergström, 2001; Migoñ and Thomas, 2002; Olvmo et al., 2005). Deep weathering varies greatly according to bedrock lithology, fabric, and jointing patterns, and as most surface erosion occurs on pre-weathered rock, both patterns of weathering and properties of saprolites are key influences on subsequent patterns and rates of denudation (Migoñ and Lidmar-Bergström, 2001). The role of deep weathering must therefore be considered in reconstructions of landform evolution (Gjessing, 1967; Migoñ and Lidmar-Bergström, 2001).

Although saprolites and grusses are also widely used as indicators of paleoclimate (Jahn, 1974; Wang et al., 1981; LaSalle et al., 1985; Peulvast, 1985; Bouchard et al., 1995) interpretations can be questionable. As gravely and sandy grusses form relatively rapidly under a range of climates and landscape settings (Migoñ and Lidmar-Bergström, 2002) they are only of Quaternary or late Neogene age (Williams et al., 1986; Migoñ and Lidmar-Bergström, 2001, 2002) and are poor climate indicators (Peulvast, 1985; Bouchard et al., 1995; Olvmo et al., 2005). Increasing clay content is interpreted as indicating longer periods of weathering and/or an increased reliance on warm conditions, with clayey grusses in formerly glaciated landscapes considered to date from, at least, the Miocene (Hall, 1985; Hall et al., 1989; Migoñ and Lidmar-Bergström, 2002; Islam et al., 2002). However, as with interpreting specific clay minerals in blockfields, the reliance of saprolite and clayey gruss development on a warm, sub-tropical or tropical, climate has likely been over-estimated (Ollier, 1988; Bird and Chivas, 1989; Bouchard et al., 1995; Crawford Elliott et al., 1997). Caution must continue to be extended to their paleoclimatic interpretation (Migoñ and Lidmar-Bergström, 2002) and, therefore, age.

The primary requirements for deep chemical weathering do not include a warm climate, but rather deep groundwater and time (Ollier, 1988). When basing climatic interpretations on saprolites; (i) the full range and quantity of clay minerals, (ii) the nature of the parent rock, (iii) drainage factors including vegetation, slope, and relief, and (iv) where they occur, the age of cover rocks, need to be considered (LaSalle and De Kimpe, 1989; Bouchard et al., 1995; Battiau-Queney, 1996; Lidmar-Bergström et al., 1997). Widespread saprolites with distinctive structure and mineralogy may be useful for paleoclimate interpretations but it is much more problematic if saprolites occur only in isolated patches, have unclear relationships with local geomorphic history, or are potentially truncated (Battiau-Queney, 1996; Migoñ and Lidmar-Bergström, 2002). Despite uncertainties regarding the role of climate in saprolite development, the age of a saprolite is frequently determined by relating its mineralogical and geochemical characteristics to a particular climate and then finding a period in which that climate was dominant (Migoñ and Lidmar-Bergström, 2002). There is a distinct need then for the dating of saprolites to be distinguished from their mineralogical and geochemical evolution (Bouchard and Jolicoeur, 2000) through, for example, stable isotope analysis (Bird and Chivas, 1988, 1989; Crawford Elliott et al., 1997; Gilg, 2000) and K–Ar and $^{40}Ar/^{39}Ar$ (Vasconcelos et al., 1992) dating.

3. The preservation of relict non-glacial surfaces during glacial periods

3.1. Preservation beneath cold-based ice

Relict non-glacial surfaces, which have been preserved beneath cold-based ice, have been extensively identified in Fennoscandia, along the eastern side of the Scandinavian mountain chain and in low lying northern areas (Kaitanen, 1969; Kleman and Borgström, 1990; Kleman and Stroeven, 1997; Hättestrand and Stroeven, 2002), in the Cairngorm Mountains of Scotland (Sugden, 1968; Hall and Sugden, 1987; Ballantyne, 1994; Glasser, 1995), on the Canadian Shield (Ives, 1958b; Aylesworth and Shilts, 1989a,b), and on Canadian Arctic islands (Sugden and Watts, 1977; Sugden, 1978; Dyke, 1983, 1993; Bierman et al., 1999; Briner et al., 2003). Other non-glacial patches preserved beneath cold-based ice have been identified on high coastal surfaces of Norway (Gellatly et al., 1988; Rea et al., 1996b; Whalley et al., 2004), Labrador (Ives, 1957; Marquette et al., 2004; Staiger et al., 2005), Newfoundland (Gosse et al., 1993), Greenland (Sugden, 1974), and Antarctica (Sugden et al., 2005). Beach ridges have been preserved beneath cold-based ice on Svalbard (Jonsson, 1983).

The primary indicators of relict non-glacial surfaces having been over-ridden (perhaps repeatedly) by ice sheets are as follows:

1. The presence of erratics (Sugden, 1978).
2. The continuation of non-glacial surfaces beneath moraines (Kleman and Borgström, 1990; Kleman et al., 1992; Clarhäll and Kleman, 1999; Ebert and Kleman, 2004).
3. The emergence of non-glacial surfaces from beneath currently retreating ice caps (Whalley et al., 1981; Gellaty et al., 1988).
4. Boulder blankets, resulting from sub-glacial entrainment and dragging of blocks, leading from non-glacial surfaces and with scars indicating block sources (Kleman and Borgström, 1990). The present instability of blocks composing boulder blankets further illustrates the relict nature of surfaces mantled by stable blockfields (Kleman, pers. com. May 2005).
5. Incision of ice marginal channels into non-glacial surfaces (Lagerbäck, 1988a; Dyke, 1993; Dredge, 2000).

Landscape preservation beneath cold-based ice is being more frequently confirmed by TCN analysis of erratics, tors, and bedrock surfaces (Fabel et al., 2002; Stroeven et al., 2002b). TCN analysis of erratics on relict surfaces in the northern Swedish mountains has revealed exposure ages corresponding to the last deglaciation, thereby confirming glacial over-riding (Fabel et al., 2002). Conversely, exposure ages of bedrock surfaces are much older (34–61 ka), demonstrating that relict areas were preserved with little or no erosion during multiple glacial cycles, having been inundated by cold-based ice (Fabel et al., 2002). Similar results have been reported from Scotland (Phillips et al., 2006), Norway (Fjellanger et al., in press), and North America (Briner et al., 2003; Marquette et al., 2004; Staiger et al., 2005).

The preservation of landscapes beneath ice sheets is primarily a function of ice temperature (Sugden, 1974, 1978; Stroeven et al., 2002a). Basal sliding and landscape erosion occur where the ice sheet base is at pressure melting point. Where basal ice remains below pressure melting point (i.e. the ice is cold-based), there is no melting and the underlying landscape will likely be preserved. Cold-based ice can, however, be erosive where large quantities of debris are entrained and ice movement is occurring through internal deformation.

Landscape and climatic factors influence basal thermal regimes. The altitude of a land surface within the vertical ice profile affects basal melting (Sugden, 1974, 1978; Stroeven et al., 2002a). As thicker ice leads to higher ice temperatures and basal melting, low elevation surfaces, such as valleys, tend to be eroded. However, fluvial valleys eroded during interglacials and structurally aligned bedrock ridges at low elevations, may be preserved where they are non-aligned with the direction of ice flow (Kleman and Stroeven, 1997; Olvmo and Johansson, 2002). High elevation surfaces are covered by thinner ice, where ice temperatures are lower and basal melting is inhibited. High elevation surfaces are therefore more likely to be preserved beneath cold-based ice covers, particularly in high-relief locations where selective linear erosion occurs and/or in locations close to ice sheet margins where ice sheet thicknesses are declining (Sugden, 1974, 1978; Briner et al., 2003; Marquette et al., 2004; Whalley et al., 2004; Staiger et al., 2005).

Although land surface altitude regulates basal melting, relict non-glacial surfaces preserved by cold-based ice can be found at any elevation within the landscape. The most extensive relict areas are located at mid-elevations (Kleman and Stroeven, 1997), i.e. below the level of cirques, whose glaciers consume high-elevation relict surfaces during periods of restricted glaciation, and above the level of warm-based, erosive, ice that forms during more extensive glaciations. Relict surfaces at high altitudes can be found where cirque glaciers have not yet consumed the non-glacial terrain and at low-lying locations that were beneath former ice divides (Aylesworth and Shilts, 1989a,b; Kleman et al., 1997) or towards ice sheet margins, where there may have been divergent ice flow (Sugden, 1978; Kleman et al., 1999) and/or rough terrain (Hättestrand and Stroeven, 2002).

Climate influences basal melting and the likelihood of landscape preservation through its effect on glacier and ice sheet mass turnover (Sugden, 1974). A maritime climate increases precipitation and produces higher ice surface temperatures that penetrate into the glacier or ice sheet. The resulting increase in mass turnover enhances the likelihood of basal melting and landscape erosion. Conversely, increasing continentality reduces precipitation and ice temperatures, thereby contributing to lower ice sheet mass turnover and reducing the likelihood of basal melting and landscape erosion. Declining temperature and precipitation gradients from SW to NE are implicated in the preservation of relict non-glacial landscapes beneath cold-based ice in Greenland. An increase in continentality may also have contributed to the preservation of relict non-glacial surfaces during glacial periods in the Canadian high-Arctic (Sugden, 1978).

In some locations where basal melting has occurred, the presence of permeable regolith may have contributed towards landscape preservation beneath ice sheets. High permeability allows basal melt water to flow through the regolith rather than at the ice-bed interface, inhibiting ice flow and decreasing erosion (Sugden, 1978). Highly permeable blockfields, occurring frequently on low-gradient upland surfaces, therefore assist in surface preservation during periods of inundation by ice sheets (Staiger et al., 2005).

The preservation of relict non-glacial surfaces highlights both the persistence and self-sustaining nature of basal thermal regimes during the Quaternary glaciations, resulting from their strong coupling with landscape relief (Kleman, 1994; Kleman and Stroeven, 1997; Stroeven and Kleman, 1999). Due to the transformation of the landscape from fluvial to glacial and the subsequent increased efficiency of the landscape to discharge ice, the coupling of basal thermal regimes with the landscape has become progressively stronger during the Quaternary (Glasser and Hall, 1997). Many relict non-glacial surfaces have become increasingly isolated from glacial erosion and are likely to survive future glaciations.

3.2. Nunataks

While the surrounding terrain is inundated by ice, some high surfaces may escape over-riding glaciation by protruding through the ice sheet as nunataks. Typically, nunataks occur as arêtes. However, plateau surfaces may also form nunataks, particularly in high coastal locations near the periphery of ice sheets, where ice sheet thicknesses are declining and where fast moving outlet glaciers are flowing through valleys, further drawing down the ice sheet, leaving high surfaces exposed (Nesje et al., 1988; Ballantyne et al., 1998; Kaplan et al., 2001; Miller et al., 2002). In addition to providing important input into ice sheet and glacio-isostacy models through constraining ice thickness (Ives, 1958a,b; Sollid and Sørbel, 1979; McCarroll et al., 1995; Ballantyne et al., 1997, 1998; Clark et al., 2003), nunataks have been proposed as possible biological refugia, thereby helping to explain the presence of endemic plant and animal species in isolated high-altitude regions (Dahl, 1955; Ives, 1974; Dahl, 1987; Nesje et al., 1988). Distinguishing relict surfaces that were preserved as nunataks from those that persisted beneath cold-based ice can therefore be of considerable value.

Although nunataks existed within the Pleistocene ice sheets, the preservation of some relict non-glacial surfaces originally interpreted as nunataks is now better explained through coverage by cold-based ice (Ives, 1975; Marquette et al., 2004). Furthermore, where sharp-crested pinnacled arêtes have protruded through the ice sheet, high weathering and erosion rates may result in surfaces that are not relict (Ballantyne et al., 1998), at the regolith scale. It seems likely that the total area of relict non-glacial surfaces preserved as nunataks is less than the total relict area preserved beneath former cold-based ice, with most nunataks occurring as small patches on high coastal surfaces, such as those in Norway (Nesje et al., 1988; Winguth et al., 2005), western Scotland (Ballantyne et al., 1997, 1998; Ballantyne, 1999), Labrador (Clark et al., 2003), Baffin Island (Steig et al., 1998; Miller et al., 2002), and Svalbard (Landvik et al., 2003), which were near the peripheries of former ice sheets.

Where erratics are present, confirming preservation of a relict surface beneath former cold-based ice is relatively straightforward. However, confirming a former nunatak is generally much more difficult. The absence of erratics is not sufficient in itself and there are instances of presumed nunataks during the last glaciation containing erratics from an earlier glaciation (Nesje et al., 1988; Ballantyne et al., 1997, 1998). Rather, identifying potential nunataks is based on the presence of distinct weathering zones separated by trimlines (Ives, 1958b, 1974; Nesje et al., 1988; Locke, 1995; Ballantyne et al., 1997, 1998; Ballantyne, 1999).

Suspected former nunataks should contain evidence of greater summit weathering, such as; (i) blockfield mantels, (ii) reduced rebound on Schmidt hammer measurements, (iii) greater dilation joint depths, (iv) a higher abundance of clay minerals, and (v) significant contrasts in the mineral magnetic parameters of soils from above and below trimlines (Ballantyne et al., 1997, 1998; Ballantyne, 1999; Walden and Ballantyne, 2002). Whereas these parameters have been successfully applied in ruling out a number of alternative hypotheses of summit weathering (Ballantyne et al., 1997, 1998), nunataks are not distinguished from surfaces preserved beneath cold-based ice.

To exclude coverage by cold-based ice, further evidence for nunataks has been sought in TCN studies. In both northwest Scotland and in western Norway, $^{26}Al/^{10}Be$ ratios indicate possible former nunataks (Brook et al., 1996; Stone et al., 1998). However, the ability of TCN analysis to distinguish nunataks from surfaces formerly covered by cold-based ice is limited by the slight decrease in the $^{26}Al/^{10}Be$ ratio over a very long period. Radioactive decay alone would decrease the $^{26}Al/^{10}Be$ ratio from the production ratio by only about 10% in 250 ka (Brook et al., 1996). Detection is limited to long burial periods, with support for former nunataks through TCN studies remaining inconclusive. Perhaps, in some instances, complete burial occurred only for short periods at glacial maximums, with nunataks otherwise persisting.

Sharply defined trimlines, which decline regularly in altitude according to ice flow direction, have also been used to distinguish former nunataks from surfaces covered by cold-based ice (Nesje et al., 1988; Ballantyne et al., 1997, 1998; Ballantyne, 1999). However, a sharp trimline could represent the maximum altitudinal limit of warm-based ice within an ice sheet, with the long-term stability of the englacial thermal boundary being insignificant to trimline formation. Due to ice pressure differences on the stoss and lee-sides of topographic obstacles, the presumed necessity for a trimline attributable to an englacial thermal boundary to show an irregular altitudinal decline (Ballantyne et al., 1998; Stone et al., 1998) may be equally valid for a trimline indicating an ice sheet surface. Furthermore, although trimlines descending towards the Scottish and Norwegian coasts are consistent with former ice flow directions (Nesje et al., 1988; Ballantyne et al., 1997, 1998), they have not been corrected for isostatic rebound, which may be of particular significance to the Scottish examples showing more gradual declines in trimline altitudes (Ballantyne et al., 1997, 1998). Breaks in slope at the trimline elevation do not necessarily occur (see Fig. 6 in Ballantyne et al., 1998). Since many cited nunataks are located away from ice accumulation areas, glacial erosion below the trimline should be more pronounced (Sugden, 1978, Ballantyne et al., 1997) and produce breaks in slope. Recently, Stone and Ballantyne (2006) found bedrock exposure ages above an inferred trimline to be almost identical to bedrock samples below, indicating over-riding glaciation and an under-estimation of the former ice sheet surface elevation. As with weathering characteristics and TCN studies, the use of regularly declining sharply defined trimlines provides inconclusive support for former nunataks.

Although confirming former nunataks through TCN studies and geomorphic evidence is difficult, support may be found in numerical ice sheet models. A two-dimensional ice-flow model applied to an east–west transect in western Norway indicated an ice thickness that was insufficient to completely inundate peaks along the transect (Winguth et al., 2005). Although the inferred ice sheet was thinner than most previous reconstructions, the ice surface elevation agreed with Näslund et al.'s (2003) reconstruction of the Fennoscandian ice sheet surface elevation. Further advances in ice sheet modelling may also provide support for former nunataks in other regions.

Perhaps the best demonstration of the difficulties in identifying former nunataks is provided by relict uplands along the east coast of Baffin Island. As Cumberland Sound and Hudson Strait were two major conduits for ice between the interior of the Laurentide ice sheet and the North Atlantic Ocean during the last glacial cycle, determining the extent and thickness of ice along the east coast of Baffin Island has significant implications for the history of the Laurentide ice sheet (Steig et al., 1998). Due partly to the difficulties in dating specific moraines (Miller et al., 2002), the question of whether these landscapes represent former nunataks or were preserved beneath cold-based ice has been the subject of considerable debate (Boyer and Pheasant, 1974; Sugden and Watts, 1977; Wolfe, 1996; Steig et al., 1998; Marsella et al., 2000; Miller et al., 2002).

Some studies conformed to the maximum-ice model of Laurentide glaciation where Baffin Island was completely inundated by ice (Flint, 1943; Denton and Hughes, 1981), while others supported a minimum-ice model where coastal uplands, featuring tors and blockfields, were located beyond the ice sheet limits (Coleman, 1920; Miller and Dyke, 1974). However, based on TCN analysis and introducing a moderate-ice paradigm, low-gradient, warm-based, glaciers moving across deformable sediments are now thought to have occupied fjords and sounds, while cold-based ice covered intervening upland surfaces (Kaplan et al., 2001; Miller et al., 2002; Briner et al., 2003) and ice-distal unscoured coastal lowlands (Briner et al., 2005). Due to steeper ice surface gradients, the zones of cold-based ice terminated inland leaving the coastal uplands partly unglaciated, with the preservation of relict upland surfaces therefore attributable to both nunataks and cold-based ice (Miller et al., 2002).

While non-glacial surfaces can be readily illustrated as relict, and some surfaces are easily shown as having been preserved beneath cold-based ice, the presence of former nunataks is inconclusive at most sites. This is not to say that they did not exist, it is just that the evidence to date is frequently insufficient to confirm their presence. Where there is such evidence, as occurs on Baffin Island, both nunataks and cold-based ice appear to have been present in close proximity, further complicating interpretations. Baffin Island also illustrates a climatic, as well as a topographic, control on the preservation of relict surfaces as nunataks, with both selective linear erosion and incomplete cold-based ice coverage partly attributable to precipitation starvation resulting from a severely cold Arctic environment. Although nunataks define maximum ice sheet thickness, the ice sheet may have attained maximum thickness at different times in different areas, with subsequent implications for ice sheet modelling (Ballantyne et al., 1998).

4. Denudation and rates of relict non-glacial surface lowering

Denudation is the removal of mass, by physical and chemical processes, from a land area such as a hillslope or drainage basin (Dixon and Thorn, 2005). Denudation rates display high spatial and temporal variability (Crickmay, 1975) according to pre-existing differences in weathering rates (Phillips, 2005). Although denudation results in surface lowering the two are not necessarily the same. Chemical denudation can remove mass without causing surface lowering (Bouchard and Jolicoeur, 2000; Dixon and Thorn, 2005), while vertical surfaces may be the major contributor to denudation calculated from streams (Phillips, 2005). Physical denudation can be used as a proxy for surface lowering in locations where it is the dominant lowering process, provided contributions from vertical surfaces are limited.

The preservation of *preglacial* surfaces is critically dependent upon low rates of denudation and surface lowering. Low denudation rates are partly a function of low relief and tectonic stability, with most relict non-glacial surfaces occurring in highly stable shield areas and old orogens (Battiau-Queney, 1996; Twidale, 1997). Low denudation rates may also be a consequence of; (i) lithology (Migoñ and Goudie, 2001), (ii) coverage by cold-based ice during glacial periods (Section 3.1), which results in lower average Quaternary rates, (iii) the presence of blockfields (Section 2.3), and (iv) a lack of surface drainage (Twidale, 1997). However, as mass wasting (defined by Flint and Skinner (1977, p. 122) as the “the movement of regolith downslope by gravity without the aid of a stream, a glacier, or wind”), is considered to reach its greatest intensity and efficacy under periglacial conditions (Ballantyne and Harris, 1994, p. 113; French, 1996, p. 149; Font et al., in press) it should not be assumed that relict non-glacial surfaces are characterised by uniformly low denudation rates. Confirming rates of denudation and surface lowering, and their spatial and temporal patterns, is crucial to determining the extent to which current non-glacial surfaces replicate preglacial surfaces.

Where the duration of erosion is known, physical denudation rates and surface lowering can be estimated from the volume and source area of eroded material. Direct point measurements within a catchment, recordings of sediment flux out of the catchment, and calculations of the volume of sediment in a downstream, or offshore, depositional basin can all be used to determine a catchment denudation rate (Fig. 5; Burbank and Anderson, 2001). Confirming low denudation rates of fragmented non-glacial surfaces in a formerly glaciated landscape is, however, difficult.

Although measurements of TCN concentrations in stream sediment can provide a long-term record of catchment sediment flux upstream of the sampling point (Granger et al., 1996), the potential for applying the strategy to determining the denudation rate of non-glacial surface fragments is highly restricted. Even where key conditions, such as a limited catchment altitudinal range and a uniform distribution of quartz throughout the catchment (Granger et al., 1996) can be met, it is extremely difficult in formerly glaciated landscapes to isolate a completely non-glacial catchment, even a very small one. Predominantly non-glacial surfaces may contain zones of partial glacial erosion (Kleman and Borgström, 1990; Glasser and Hall, 1997; Kleman and Stroeven, 1997; Clarhäll and Kleman, 1999; André, 2004; Phillips et al., 2006; Harbor et al., 2006), while very well-preserved non-glacial surfaces are frequently blanketed by thin till covers (Kleman and Stroeven, 1997; Bouchard and Jolicoeur, 2000). High periglaciated surfaces with no signs of former

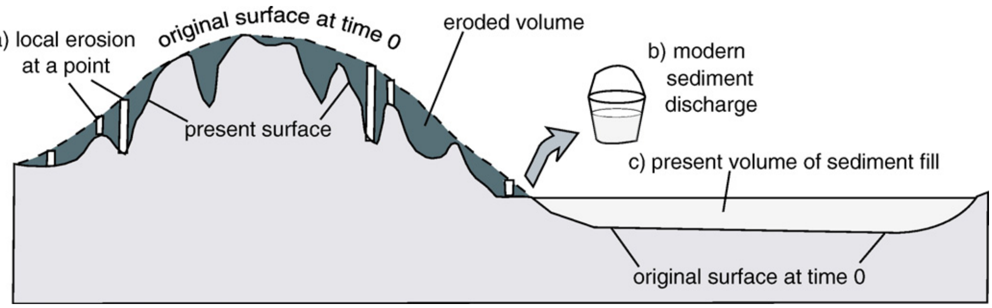

Fig. 5. Different methods for determining denudation of a catchment: a) point measurements; b) stream sediment flux from the catchment, and; c) sediment volume in a downstream depositional basin (source: Fig. 7.3 in Burbank and Anderson, 2001, p. 134, with permission of Blackwell Publishing and the authors).

glaciation often lack distinct surface drainage. If a very small catchment could be isolated, it would essentially be a point measurement that may have little relevance to the entire non-glacial surface fragment and to other non-glacial fragments.

The potential for using sediment volumes in depositional basins to determine physical denudation rates of relict non-glacial surfaces is also limited, due to contributions from both glacial and non-glacial erosion. Even assuming that all non-glacial surface erosion occurs during interglacial and interstadial periods in areas that have been inundated by cold-based ice, it is likely that there are substantial differences in denudation rates between these periods. Furthermore, if sediments from interglacial and interstadial periods could be distinguished from glacial period sediments, catchments often remain partly glacierised and/or sediments are sourced from both relict glacial and relict non-glacial parts of the landscape. It is not then possible to determine the sediment volume derived from erosion of relict non-glacial surfaces.

A further possible strategy using sediment volumes in depositional basins is to back-fill the landscape with sediments according to glacial erosion intensity (Glasser and Hall, 1997). Most sediments are back-filled into zones of high erosion intensity, such as glacial troughs and aerially-scoured surfaces, with smaller amounts back-filled into less intensely eroded zones, such as those containing roches moutonnées, and so on. In theory, remaining sediment volumes may then be attributed to relict non-glacial surfaces, permitting estimates of surface denudation. However, determining physical denudation rates of relict non-glacial surfaces by back-filling sediments will only be practicable where there are large remaining areas of relict non-glacial surfaces.

While point measurements are handicapped by the difficulties in integrating results across an entire surface (Cockburn and Summerfield, 2004), they form the best alternative for determining denudation rates of relict non-glacial surfaces. Maximum denudation rates, calculated from TCNs, of exposed bedrock surfaces, including tors, vary between 1.1 and 19 mm ka^{-1} (Small et al., 1997; Bierman et al., 1999; Stroeven et al., 2002b; Phillips et al., 2006). Studies of glacially modified bedrock provide useful information of Holocene denudation rates. For example, Holocene weathering of ice-polished bedrock surfaces in the northern Swedish mountains has been estimated at 0.2 to 5.0 mm ka^{-1}, depending on the rock type (André, 2002). In general, expected low rates of surface denudation have been confirmed, with denudation rates of relict non-glacial bedrock surfaces being similar in magnitude to bedrock denudation rates recorded in the southeastern Australian highlands (Heimsath et al., 2001a), Central Namib Desert and Namibian escarpment (Cockburn et al., 1999, 2000), the Drakensburg escarpment (Fleming et al., 1999), and the Eyre Peninsula and Northern Territory, Australia (Bierman and Caffee, 2001, 2002).

Denudation rates derived from bedrock relate, however, to the most resistant parts of the landscape, so will be lower than for regolith mantled surfaces (Cockburn and Summerfield, 2004; Phillips et al., 2006; Matsuoka et al., in press). In the Wind River Range, Wyoming, rates of bedrock lowering beneath 90–100 cm of regolith (mean 14.3±4.0 mm ka^{-1}) are slightly higher to almost double the rates on adjacent bare bedrock surfaces (mean 7.6±3.9 mm ka^{-1}; Small et al., 1999). In the Cairngorm Mountains, Scotland, a regolith erosion rate was calculated as 35±4.0 mm ka^{-1} (Phillips et al., 2006). These results are similar to saprolite production rates (maximum of 20 mm ka^{-1}) on the Appalachian Piedmont, beyond the area of former glaciation (Pavich, 1989). Differential weathering between granite and gneiss indicates that desert pavements in the Sør Rondane Mountains, East Antarctica, lower at rates exceeding 10 mm ka^{-1} compared with bedrock erosion rates of 0.1–0.7 mm ka^{-1} (Matsuoka et al., in press). Rates of regolith production and removal in other, non-glaciated, landscapes also indicate dynamic surface lowering (Heimsath et al., 1997, 1999, 2000, 2001a,b; Bierman and Caffee, 2001; Granger et al., 2001; Bierman and Caffee, 2002). As regolith mantled surfaces dominate relict non-glacial landscapes, measuring their denudation rates potentially provides a better impression of relict non-glacial landscape lowering than denudation rates of exposed bedrock. However, the possible removal or addition of regolith by former ice covers poses a challenge to measurements of denudation.

Although chemical weathering features are limited in periglacial environments to rinds, coatings, and soils, chemical processes play a potentially significant role in the denudation of relict non-glacial surfaces (Dixon and Thorn, 2005). As water is always available at some scale chemical denudation operates continuously, but with high spatial and temporal variability (Pope et al., 1995; Dixon and Thorn, 2005). At the hillslope scale, spatial variations in chemical denudation reflect changes in drainage conditions (Battiau-Queney, 1996) that may be attributable to differences in bedrock lithology and depth of regolith cover, or to the presence of nivation hollows (Thorn and Hall, 2002; Dixon and Thorn, 2005). Temporal variations in chemical denudation rates may occur, for example, seasonally with melting or over a much longer period such as a glacial–paraglacial–periglacial cycle (Dixon and Thorn, 2005). The strong coupling of solute removal to the drainage system results in chemical denudation being more efficient than physical denudation, where sediments are frequently stored for long periods (Dixon and Thorn, 2005).

Chemical denudation has not been explicitly measured on relict non-glacial surfaces but has been recorded in alpine and polar periglacial environments (Rapp, 1960; Caine and Thurman, 1990; Drever and Zobrist, 1992; Campbell et al., 2002; Beylich et al., 2004). Alpine areas frequently display low rates of chemical denudation (Caine and Thurman, 1990; Drever and Zobrist, 1992; Riebe et al., 2004). In the Latnjavagge valley, northern Sweden, Beylich et al. (2004) recorded a chemical denudation rate of 5.4 t km^{-2} $year^{-1}$. However, in a nearby valley, Kärkevagge, Rapp (1960) and Campbell et al. (2002) recorded significantly greater chemical denudation rates of 27 t km^{-2} $year^{-1}$ and 46 t km^{-2} for an 82 day melt period, respectively. While rate differences in the Swedish examples have been attributed to varying lithology and measurement techniques (Campbell et al., 2001, 2002), it is likely that the presence of surface regolith is a crucial factor in chemical denudation rates (Dixon and Thorn, 2005). Large surface exposures of bedrock significantly restrict chemical denudation while ground water movement through regolith greatly enhances the potential for chemical weathering. As many relict non-glacial surfaces are regolith mantled, chemical denudation is potentially significant despite the rarity of chemically induced landforms. Although chemical denudation rates in the Latnjavagge valley are low they still slightly exceed rates of physical denudation (Beylich et al., 2006). For a more complete understanding of the role of chemical denudation in relict non-glacial landscape development the nature of chemical processes, including reactions that contribute dissolved loads to streams and their absolute and relative rates, need to be considered with physical denudational processes (Bouchard and Jolicoeur, 2000; Dixon and Thorn, 2005).

In addition to quantifying chemical denudation, further quantitative studies of regolith production and erosion are required to achieve a more accurate interpretation of non-glacial surface denudation rates, including spatial and temporal variations. Spatial variations, for example, in local climate, bedrock lithology, slope gradient, permafrost distribution and active layer depth, and regolith characteristics such as depth, fine matrix content, and water volume, can produce spatial variations in denudation rates. Temporal variations may contribute to landscape preservation with most denudation perhaps being achieved in the late Pliocene and early Pleistocene when landscapes were adjusting to a periglacial regime and/or during high-energy paraglacial periods when there was an abundance of available water and fresh regolith (Ballantyne, 2002; Dixon and Thorn, 2005). Landscapes, such as those mantled by blockfields, may then be relicts of higher energy periods remaining little altered during subsequent, or intervening, lower energy regimes (Brunsden, 2001; Granger et al., 2001; Dixon and Thorn, 2005).

5. Relict non-glacial surface evolution and numerical modelling

Degradation of relict non-glacial surfaces following cessation of tectonic activity can be explored through hillslope-scale numerical modelling. The hillslope scale is relevant to understanding the role of surface processes in relict surface degradation and numerical modelling allows their, often complex, linkages to be investigated (Burbank and Anderson, 2001, p. 231). Important processes, such as rates of regolith production and transport mechanisms, are being increasingly constrained through TCN studies at timescales relevant to hillslope evolution (McKean et al., 1993; Heimsath et al., 1997, 1999; Small et al., 1999; Heimsath et al., 2001a,b;

Anderson, 2002; Wilkinson et al., 2005). Typically, hill slopes and summits in temperate locations, possessing soil covers and distinct bedrock–soil interfaces, have been modelled (Dietrich et al., 1995; Braun et al., 2001; Roering et al., 2001), but comparatively little is known about surface evolution in periglaciated landscapes. Investigating non-glacial surface evolution through numerical modelling is of considerable value to understanding mechanisms and timescales of periglaciated relict surface evolution, providing a more definite basis for reconstructing preglacial landscapes and quantifying glacial erosion, and uniting "historical" landscape studies with physically-based "process" studies (Martin and Church, 2004).

Despite the potential of numerical hillslope modelling, there is considerable disagreement regarding fundamental slope processes. It remains unresolved whether maximum soil production conforms to an exponential decay model (Fig. 6a; Heimsath et al., 1997, 1999, 2000, 2001a,b) or to a finite-depth model (Fig. 6a; Dietrich et al., 1995; Small et al., 1999; Anderson, 2002). The importance of regolith mixing and subsequent downslope transport as a plug flow or shear flow is also unresolved (Fig. 6b). Simpler plug flow is usually assumed (McKean et al., 1993; Small et al., 1999; Braun et al., 2001) and Monaghan et al. (1992) found little change in predicted soil production rates evaluated through plug flow and pure shear flow. However, convex upward sediment velocity profiles, approximating shear flow, have been experimentally determined (Roering, 2004). Linear and non-linear transport mechanisms (where linear and non-linear refer to the relationship between sediment transport and slope; Fig. 6c) have each found support in different studies. Indications are that the transport mechanisms apply to different parts of the hillslope (Heimsath et al., 2000; Braun et al., 2001), with linear transport restricted to divergent parts of the landscape with low to moderate gradients (Schumm, 1967; McKean et al., 1993; Roering et al., 2001), such as convex summits (Gilbert, 1909; Small et al., 1999; Anderson, 2002) and ridge crests (Heimsath et al., 1999). Non-linear transport dominates steeper straight slopes away from summits and ridge crests (Howard, 1994; Martin and Church, 1997; Roering et al., 1999, 2001; Montgomery, 2003). There are also challenges in adapting numerical models to periglacial locations with coarse regolith of variable density, poorly defined bedrock–regolith interfaces, present or past permafrost, and/or past coverage by cold-based ice. A simplifying strategy would be to consider summits

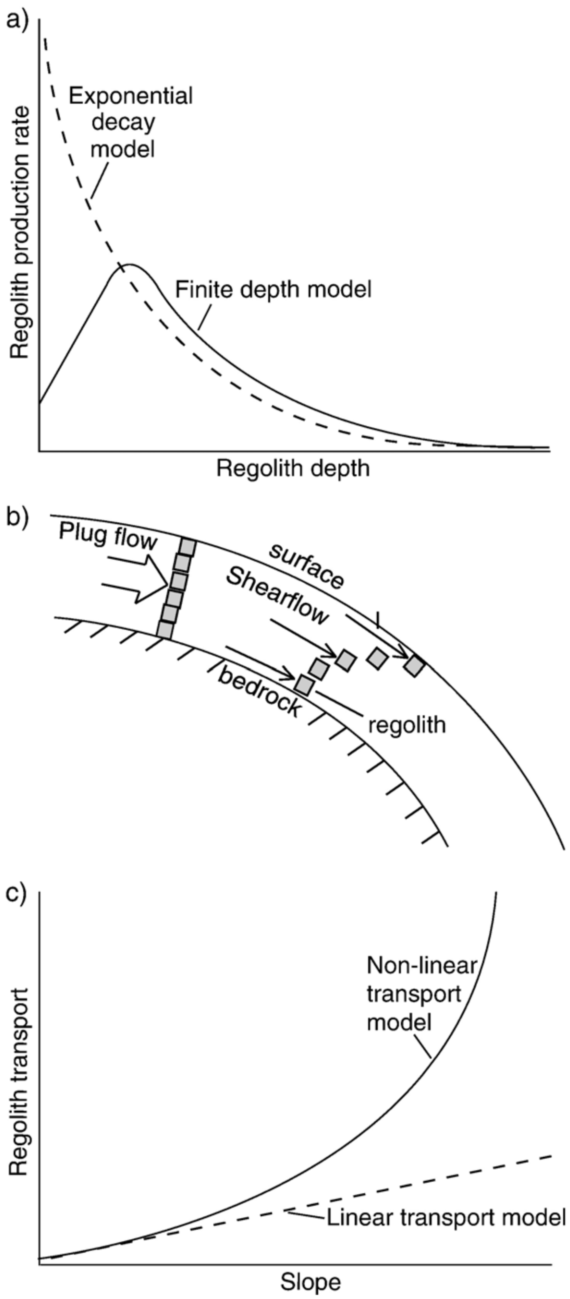

Fig. 6. A comparison of slope processes: a) finite depth and exponential decay models of regolith production. In the finite depth model, maximum regolith production occurs under a thin regolith cover, while according to the exponential decay model, maximum regolith production occurs where there is no regolith cover and decreases exponentially with increasing regolith depth. b) Plug flow versus shear flow. Plug flow occurs where the downslope regolith flux is constant with regolith depth. Conversely, shear flow describes an increase in regolith flux towards the surface. c) Linear and non-linear transport models. According to the linear transport model regolith flux increases linearly with slope, while in the non-linear transport model the increase in regolith flux is nearly linear at low slope gradients but then increases non-linearly, and very rapidly, at higher slope gradients (modified from Roering et al., 1999).

(Heimsath et al., 1999). Consisting of low-gradient convex slopes, summits could be modelled with linear transport, with rates of summit lowering pinning over-all non-glacial surface lowering.

One series of studies that combined numerical modelling with TCN-based analysis of denudation on periglaciated, non-glacial surfaces was that by Small et al. (1997, 1999) and Anderson (2002) on the high plateau surfaces of the Wind River Range, Wyoming. Based on ^{10}Be and ^{26}Al, Small et al. (1999) found that the surfaces were lowering at 14.3±4.0 m Ma^{-1} and 13.0±4.0 m Ma^{-1}, respectively, and were exhibiting dynamic equilibrium. Anderson (2002) concluded that the non-glacial surfaces are not static remnants of a once widespread erosion surface, but rather have evolved separately since fragmentation, with tors indicating initial conditions of thin regolith and steeper slopes. As neither the depth of erosion nor change in morphology was large, and the remnant non-glacial surfaces are strongly decoupled from glacial troughs, they may be used as geomorphic markers to quantify glacial erosion (Anderson, 2002).

Findings from the Wind River Range, relating particularly to the operation of dynamic equilibrium, are not necessarily widely applicable. The Wind River Range summit plateaus are relatively simple, composed of very low gradient, smooth convex slopes with bare bedrock edges (Small et al., 1999; Anderson, 2002). Other non-glacial surfaces may contain smaller convex summit areas merging into straight slopes, indicating the absence of dynamic equilibrium or its restriction to summits, the dominance of non-linear slope transport with potentially higher rates of evolution, and changing surface morphology over time (Roering et al., 1999, 2001). Hillslope studies in temperate locations indicate highly dynamic, non-steady state, evolution where lowering rates on different parts of the landscape differ by more than an order of magnitude depending on local soil thickness (Heimsath et al., 1997, 2001a,b). The evolution of periglaciated non-glacial surface remnants combining convex summits and straight slopes that are developed on resistant lithologies, mantled with coarse regolith, and lack surface drainage remains essentially unknown. While blockfields and bedrock exposures may modulate hillslope denudation rates (Granger et al., 2001), spatially variable denudation rates will result in evolving surface morphologies.

6. Reconstructing preglacial surfaces and quantifying glacial erosion from non-glacial remnants

The potential use of non-glacial surface fragments in reconstructing local or regional preglacial surfaces is a key consideration. Where the preglacial landscape represents an uplifted planation or low-relief surface, a volume of erosion can be obtained by subtracting the current landscape from the reconstructed preglacial landscape. Where the time of uplift is known an average erosion rate can be calculated (Burbank and Anderson, 2001, p. 134). If significant erosion only commenced with the onset of glaciation then the erosion volume may approximate the glacial erosion volume and a glacial erosion rate may be calculated (Nesje et al., 1992). In some instances preglacial fluvial valleys can be reconstructed, providing more accurate estimates of glacial erosion quantities (Kleman and Stroeven, 1997; Stroeven et al., 2002a). From erosion volumes, it may be possible to distinguish the relative effects of glacial and fluvial erosion in landscape development and also determine isostatic landscape responses to erosional unloading (Glasser and Hall, 1997; Small and Anderson, 1998).

Local preglacial surfaces have been reconstructed by extending lines across glacial troughs from non-glacial surface remnants. The preglacial topography is extrapolated from the adjacent, existing slopes (Fig. 7a; Reusch, 1901a; Nesje et al., 1992; Nesje and Whillans, 1994; Kleman and Stroeven, 1997; Stroeven et al., 2002a; Bonow et al., 2003), permitting an estimate of glacial erosion. The method is valid if there is both negligible lowering of the non-glacial remnants, sometimes confirmed through TCN studies (Stroeven et al., 2002a), and the absence of change in relict surface form. If surface lowering occurs through dynamic equilibrium, quantities of both glacial and non-glacial erosion will be under-estimated (Fig. 7b; Glasser and Hall, 1997). Conversely, surface lowering involving changing non-glacial surface morphologies may result in over-estimations of glacial erosion and under-estimations of non-glacial erosion (Fig. 7c). The effect of trough incision and resultant cliff recession on relict surface form is also little known (Dixon and Thorn, 2005). Low gradient Wind River Range surfaces operate independently of both widening and lowering of glacial troughs (Anderson, 2002). However, possible changes in ground water flow and instabilities resulting from trough incision, particularly on steeper surfaces, could alter non-glacial surface edge morphology and result in an under-estimation of glacial erosion (Fig. 7d). An understanding of relict non-glacial surface denudation rates and morphological changes over time is required to increase the accuracy and precision of glacial erosion estimates.

Quantifying regional glacial erosion requires that the form of the preglacial landscape and the pattern of

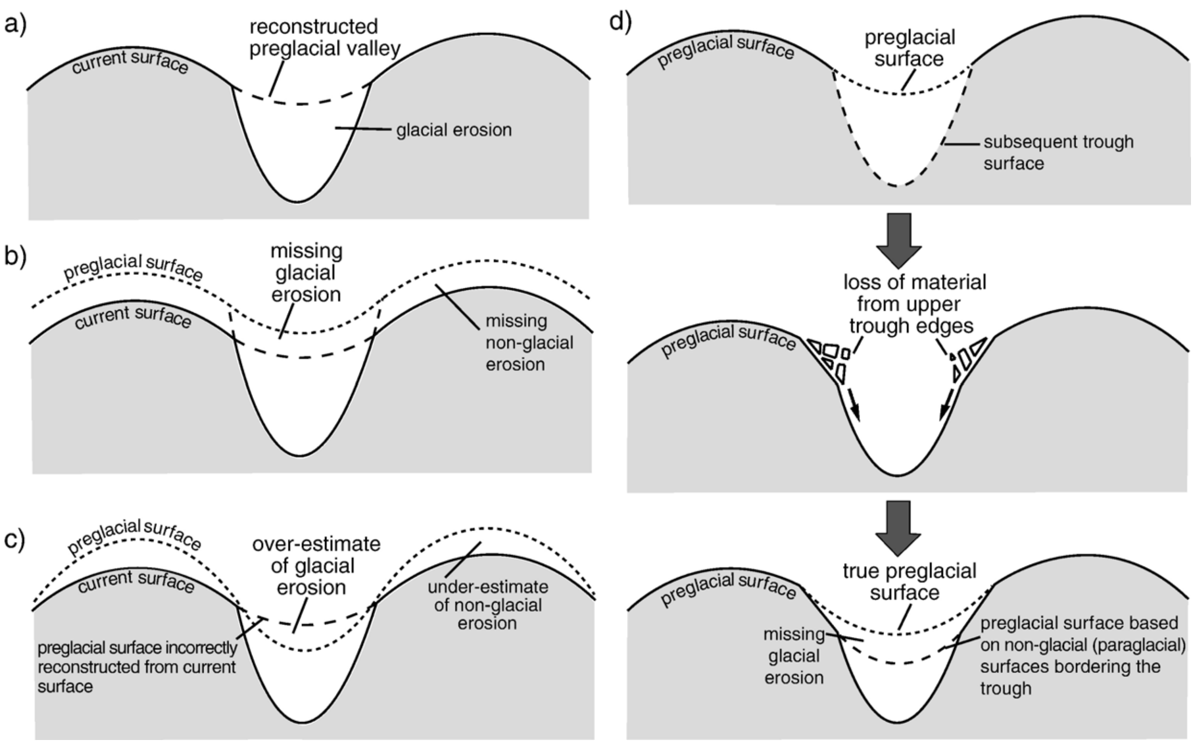

Fig. 7. a). Reconstruction of a local preglacial surface across a glacial trough based on remnant slopes. The cross-sectional area of glacial erosion is calculated by subtracting the current topography from the reconstructed topography. b) Under-estimation of glacial and non-glacial erosion based on current non-glacial surfaces and an assumption of static equilibrium. c) An over-estimate of glacial erosion and an under-estimate of non-glacial erosion resulting from evolution of remnant non-glacial surfaces. d) Preglacial surface remnants have undergone a paraglacial response to trough incision, leading to an incorrect preglacial valley reconstruction and an under-estimate of glacial erosion. For simplicity, static equilibrium of remnant preglacial surfaces has been assumed.

glacial erosion intensity are established from geomorphic evidence, such as landforms, *in situ* regolith, deposits, and denudation rates derived from TCNs (Sugden, 1976, 1978; Nesje et al., 1992; Glasser and Hall, 1997; Small and Anderson, 1998; Stroeven et al., 2002a; Li et al., 2005). Nesje et al. (1992) calculated the total Quaternary erosion of the Sognefjord drainage basin in Norway by subtracting the present topography from a reconstructed preglacial landscape. A basin erosion volume of 7610 km^3 and a glacial erosion rate of 2 ± 0.5 m ka^{-1} were derived. Glasser and Hall (1997) calculated glacial erosion in NE Scotland through a landscape reconstruction. By dividing the landscape into categories reflecting the extent of glacial modification and estimating erosion in each category an over-all erosion rate for the study area of 21 mm ka^{-1} was calculated for the entire Quaternary. The presence of spatial and temporal variations in rates of glacial erosion was apparent in the low erosion volumes compared with sediment volumes in the North Sea (Glasser and Hall, 1997). A better understanding of how non-glacial surfaces evolve should help to correct the discrepancy between estimates of glacial erosion based on offshore sediment volumes and those based on reconstructions of the preglacial landscape.

Quantifying erosion of glacial and non-glacial surfaces permits calculations of relief development in formerly glaciated landscapes. Small and Anderson (1998) calculated the average amount of erosion for the glacially-dissected Wind River Range, Wyoming, from a reference surface created by joining non-glacial surface remnants. Relict non-glacial summit surfaces were found in TCN studies to be eroding at $\sim$10 m Ma^{-1}, with erosion rates of intervening glacial valleys being an order of magnitude faster. About 50–100 m of isostatic summit uplift resulting from glacial trough incision was counteracted by non-glacial surface erosion, resulting in summit elevations that have remained essentially constant despite the increased relief (Small and Anderson, 1998). Similar relief generation has also been confirmed in the Torngat Mountains, Labrador (Staiger et al., 2005), with $<$1.4 m Ma^{-1} of bedrock erosion on summit plateaus compared with 50 m of glacial erosion in troughs.

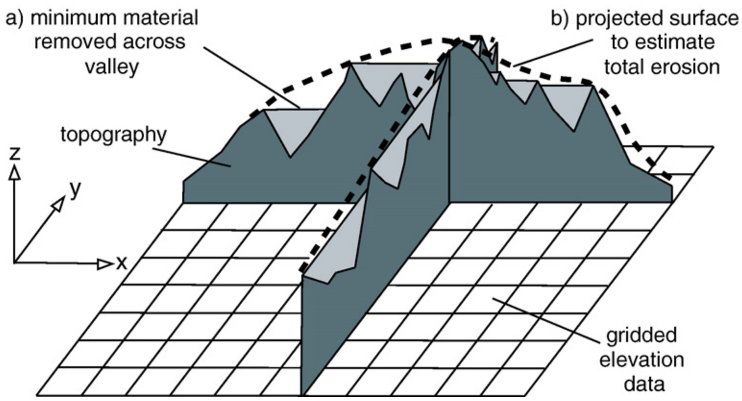

Fig. 8. a) Estimating minimum erosion volume by subtracting existing topography from dissected parts of the landscape that have been filled to the elevations of the lowest topographic boundaries surrounding them. b) Estimating erosion volume by subtracting existing topography from envelope surfaces connecting summits (source: Fig. 7.6A in Burbank and Anderson, 2001, p. 139, with permission of Blackwell Publishing and the authors).

If an over-all quantity of erosion is sought, with no requirement for the component of glacial erosion, then an initial, non-eroded, surface can be recreated from present summits. Minimum erosion volumes may be estimated by subtracting existing topography from dissected parts of the landscape that have been filled to the elevations of the lowest topographic boundaries surrounding them (Fig. 8a; Burbank and Anderson, 2001). More subjective estimates of erosion volumes can be determined by subtracting existing topography from envelope surfaces connecting summits (Fig. 8b; Burbank and Anderson, 2001). In both instances, erosion quantities will be under-estimated, as summit elevations have not been corrected for denudation over time, although actual errors are likely to be small in comparison to other uncertainties inherent in the technique.

Accurately establishing the form of the preglacial landscape and the pattern of glacial erosion intensity for calculations of glacial erosion requires detailed mapping of relict non-glacial surfaces, an understanding of denudation rates and changes in surface morphology, and an insight into ice sheet dynamics and basal thermal regimes. While there have been some efforts to map relict non-glacial surfaces in detail (Glasser and Hall, 1997; Kleman and Stroeven, 1997; Kleman et al., 1999), their temporal evolution has remained little studied. Particularly in regions of selective linear erosion, where large areas of non-glacial surfaces have been preserved, non-glacial processes may have contributed significantly to landscape denudation and resulting offshore sediment volumes.

7. Conclusion

Relict non-glacial surfaces within formerly glaciated landscapes are dynamic features. Although present morphologies and denudation rate evidence indicate a continuous non-glacial surface history since preglacial times, non-glacial surfaces have evolved during the Quaternary. Depending on spatial variables such as lithology, slope, regolith depth and the abundance of fine matrix and water, some surfaces are evolving slowly, while others display more rapid evolution. High spatial variability in denudation rates results in changing surface morphologies over time. Denudation rates also display high temporal variability, with much surface evolution having perhaps occurred soon after the initial onset of glaciation or during paraglacial phases. While some parts of non-glacial landscapes are currently active, others may be largely inactive relicts of past higher energy regimes. Although non-glacial surfaces are dynamic, much remains to be determined regarding surface denudation rates and the magnitude of morphological changes over time.

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