ELSEVIER

Earth and Planetary Science Letters 201 (2002) 397–406

EPSL

www.elsevier.com/locate/epsl

Landscape preservation under Fennoscandian ice sheets determined from in situ produced ^{10}Be and ^{26}Al

Derek Fabel [a,*], Arjen P. Stroeven [b], Jon Harbor [c], Johan Kleman [b], David Elmore [d], David Fink [e]

[a] *Research School of Earth Sciences, Australian National University, Canberra, ACT 0200, Australia*
[b] *Department of Physical Geography and Quaternary Geology, Stockholm University, S-106 91 Stockholm, Sweden*
[c] *Department of Earth and Atmospheric Sciences, Purdue University, West Lafayette, IN 47907, USA*
[d] *Purdue Rare Isotope Measurement Laboratory, Purdue University, West Lafayette, IN 47907, USA*
[e] *AMS-ANTARES, Physics Division, ANSTO, PMB1, Menai, NSW 2234, Australia*

Received 7 February 2002; received in revised form 2 May 2002; accepted 7 May 2002

Abstract

Some areas within ice sheet boundaries retain pre-existing landforms and thus either remained as ice free islands (nunataks) during glaciation, or were preserved under ice. Differentiating between these alternatives has significant implications for paleoenvironment, ice sheet surface elevation, and ice volume reconstructions. In the northern Swedish mountains, in situ cosmogenic ^{10}Be and ^{26}Al concentrations from glacial erratics on relict surfaces as well as glacially eroded bedrock adjacent to these surfaces, provide consistent last deglaciation exposure ages ($\sim$8–13 kyr), confirming ice sheet overriding as opposed to ice free conditions. However, these ages contrast with exposure ages of 34–61 kyr on bedrock surfaces in these same relict areas, demonstrating that relict areas were preserved with little erosion through multiple glacial cycles. Based on the difference in radioactive decay between ^{26}Al and ^{10}Be, the measured nuclide concentration in one of these bedrock surfaces suggests that it remained largely unmodified for a minimum period of 845^{+461}_{-418} kyr. These results indicate that relict areas need to be accounted for as frozen bed patches in basal boundary conditions for ice sheet models, and in landscape development models. Subglacial preservation also implies that source areas for glacial sediments in ocean cores are considerably smaller than the total area covered by ice sheets. These relict areas also have significance as potential long-term subglacial biologic refugia.

Keywords: cosmogenic elements; Sweden; ice sheets; erosion; glacial features

1. Introduction

After initial fierce debate over whether ice would preserve or erode underlying rocks and sediments [1,2], conventional wisdom since the early 1900s has been that glaciers and ice sheets are effective agents of erosion and landscape change [3]. Subsequent advances in understanding

* Corresponding author. Tel.: +61-2-6125-5518; Fax: +61-2-6125-5443.
E-mail addresses: derek.fabel@anu.edu.au (D. Fabel), arjen@geo.su.se (A.P. Stroeven), jharbor@purdue.edu (J. Harbor), kleman@natgeo.su.se (J. Kleman), elmore@purdue.edu (D. Elmore), fink@ansto.gov.au (D. Fink).

PII: S0012-821X(02)00714-8

ice sheet dynamics and glacial erosion processes have renewed interest in the thermal conditions beneath ice sheets, differentiating pressure melting conditions, when water at the bed facilitates erosion of the substrate, from conditions so cold that the ice is frozen to the substrate, resulting in preservation [4,5]. Identification of preserved areas can be complicated because alternate interpretations would be that these areas were ice free (nunataks), and thus that their edges were former ice margins. This is critical in all areas formerly covered by ice sheets, including the North Atlantic continental margin mountains where it is widely recognized that the geomorphology consists of a patchwork of ice scoured terrain and relict upland surfaces [6–9] (e.g. Fig. 1). Distinguishing between areas that were ice free from those covered by ice frozen to its bed can alter inferred ice surface elevations and volumes, and associated paleoclimatic reconstructions [5,10,11], and also has important implications for the basal boundary conditions of ice sheet models.

Recognition of patches of relict landforms in the northern Swedish mountains (Fig. 2) has been based on the presence of nonglacial features [8] within areas that are known to have been ice covered because of the presence of erratics (boulders with a lithology different from that locally available; Fig. 3), and from their positions relative to ice sheet terminal positions [11]. Well within the ice margins, and adjacent to relatively low-lying areas with clear evidence of extensive glacial erosion, are upland landscape patches with nonglacial features such as rounded symmetrical summits with weathering mantles and protruding bedrock outcrops (tors), V-shaped valleys, and abundant periglacial features (Figs. 1 and 3). However, this area has been covered repeatedly by ice sheets, as indicated by onshore mapping and dating of glacial landforms [12], ice surface reconstructions based on glaciological constraints [13], and patterns of isostatic rebound [14]. This patchwork of glacially scoured and relict surfaces (Fig. 2) led Kleman and Stroeven [8] to conclude that the average subglacial thermal regime of the ice sheets since the late Tertiary was frozen on the uplands and melting in the main valleys. The interpretation of these patches as relicts is based on qualitative analyses of landforms [8] and inferences from soil chemistry [15].

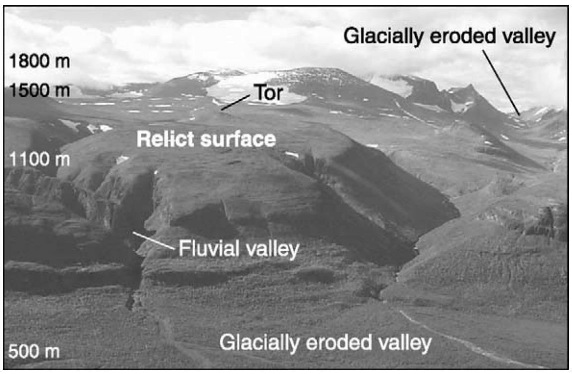

Fig. 1. Patchwork of landscape types typical of the northern Swedish mountains with glacially eroded valleys juxtaposed to rounded relict surfaces, tors, and V-shaped fluvial valleys. Elevations are approximate.

The goal of the work presented here was to provide an independent, quantitative test of the hypothesis that relict landform patches represent areas preserved under ice, rather than nunataks. A field campaign was mounted in northern Sweden to determine the surface exposure ages of relict surfaces, bedrock exposures where these relict surfaces have been glacially plucked, and erratics lying on relict surfaces, using cosmogenic nuclide techniques.

2. Methodology

The accumulation of in situ produced cosmogenic ^{10}Be (half-life = $1.51 \pm 0.05 \times 10^6$ yr [16]) and ^{26}Al (half-life = $7.1 \pm 0.2 \times 10^5$ yr [16]) in quartz exposed to cosmic radiation provides a means of determining the amount of time the rock has been at or near the ground surface [17–19]. Because nuclide production decreases with depth, removal of 2 m or more of irradiated rock during one glacial event will create a zero age surface [20]. In this context, areas known to have been ice covered should have exposure ages equivalent to deglaciation if they were significantly eroded by ice and older exposure ages if they suffered limited erosion or were completely protected. Further, nunataks should have exposure ages older than the last glaciation (unless there was significant subaerial erosion [18]). Ice cover of 10 m or more shields the underlying rock surface from most cosmic radiation, so areas that undergo multiple cycles of ice sheet overriding but no erosion should accumulate ^{10}Be and ^{26}Al concentrations equivalent to the sum of the ice free periods, minus decay during periods of ice shielding [19,21, 22].

Patches of relict landform assemblages, representing 20–25% of an area believed to have been covered by ice at the last glacial maximum, were identified in the northern Swedish mountains through extensive field and air photo mapping (Fig. 2). Quartz-rich samples for cosmogenic radionuclide analysis were collected from bedrock outcrops and erratics in mapped relict patches (Fig. 3) and from transverse lee-side scarps [23]. Erratics confirm that the sites were overridden by ice and were dated to determine whether they were deposited during the last glaciation. Bedrock in relict areas was sampled to determine whether these sites were in fact extensively eroded during the last glaciation (to give a deglacial exposure age) or whether they were moderately eroded or are relict (exposure ages reflect both postglacial time and cosmogenic nuclide inheritance from one or more previous ice free periods). Transverse lee-side scarps, a common landform at the transition between an up-ice frozen bed relict patch and a distal thawed zone, are thought to result from the entrainment of debris sheets in a zone of extending ice flow [24], and like erratics are indicators of ice overriding. The till scarps indicate that the full weathering mantle or till thickness and possibly some bedrock was removed.

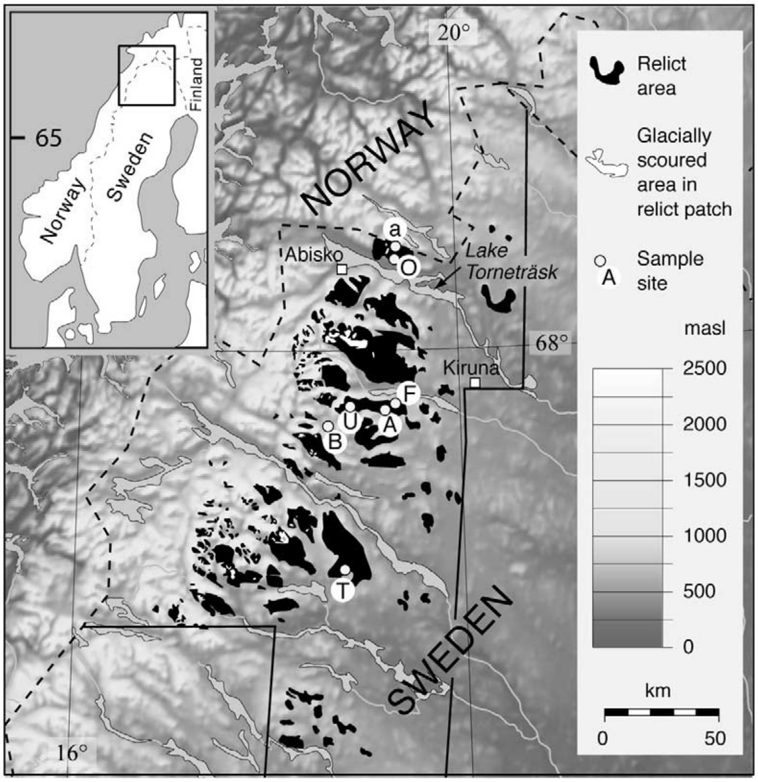

Fig. 2. Relict surfaces (black) in a portion of the northern Swedish mountains. The black line and dashed national boundaries bound the area where relict surfaces have been mapped in detail. Relict surfaces occupy 22% of this mapped area. Dots indicate the sample sites in Table 1: a, Alddasčorru; O, Olmáčohkka; F, Fávrratčohkka; A, Áilladis; U, Urttičohkka; T, Tjuolmma; B, Bálddavárri. The Aldasčorru site is shown in Fig. 3. The location of the map is shown by the black box in the inset.

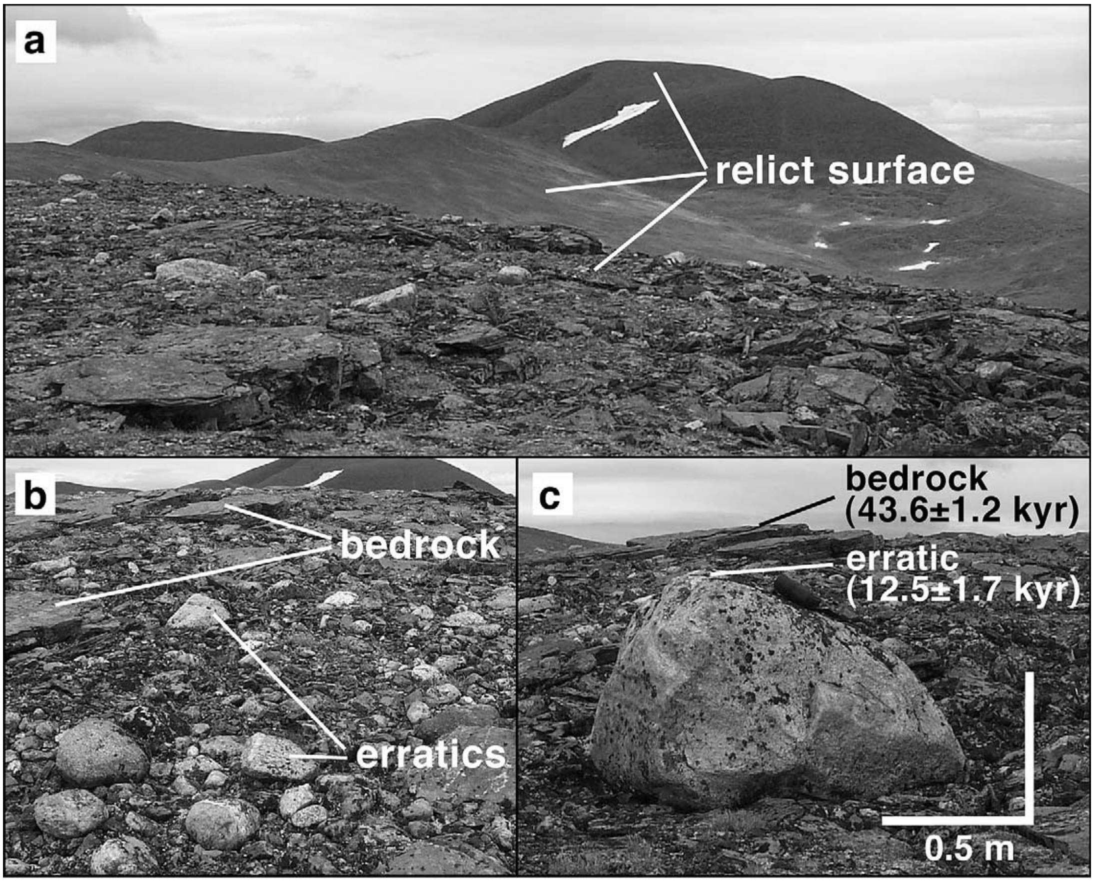

Fig. 3. Alddasorru field site. (a) Rounded symmetrical ridges and summits typical of a relict surface. (b) Light-colored, rounded quartzite erratics contrast with blocky, angular bedrock outcrops of fine-grained shist. (c) The apparent exposure age of the erratic (12.5 ± 1.7 ^{10}Be kyr) demonstrates deposition by the last Fennoscandian ice sheet. The inherited cosmogenic nuclide concentration from previous exposure periods results in a much greater apparent exposure age for the bedrock sample (43.6 ± 1.2 ^{10}Be kyr) indicating that it is a relict surface.

Samples from bedrock immediately downstream of the transverse lee-side scarps should provide deglaciation ages provided the combined weathering mantle, till, or bedrock thickness removed was in excess of 2 m and the transverse lee-side scarp was created during the last glaciation.

Approximately 30 g of pure quartz was separated from each sample using magnetic and heavy liquid separation followed by selective chemical dissolution [25]. The quartz was spiked with ~0.6 mg ^{9}Be carrier and dissolved. Total Al concentrations in aliquots of the dissolved quartz were quantified by ICP and assigned a 5% uncertainty. Al and Be were purified and separated by ion chromatography, selectively precipitated as hydroxides, and oxidized. A now corrected problem in the chemical extraction process resulted in unrecoverable loss of Al from 12 of the 15 samples processed and only ^{10}Be could be quantified for these.

To calculate cosmogenic surface exposure ages we used the standard models of [17] with sea level, high latitude ($>60°$) nuclide production rates scaled to altitude and latitude using [26] (Table 1). Adopting the scaling factors of [27] would increase calculated ages by no more than 5%. Uncertainties in single nuclide exposure ages (Table 1) are reported as one standard error of analytical uncertainty calculated from AMS counting statistics and uncertainty in total Al measurement by ICP (5%). The uncertainty shown in parentheses includes random (8%) and systematic uncertainties (10%) and should be used when comparing data to calendar ages [16].

3. Results

The Alddasčorru site north of Lake Torneträsk (Figs. 2 and 3), a typical relict area, has blocky, angular bedrock outcrops and loose mantle, with no evidence of glacial erosion (e.g. striae or polish), but frequent, rounded erratics. The large erratic (Fig. 3c) has an apparent exposure age of 12.5 ± 1.7 ^{10}Be kyr (Table 1). In contrast, the bedrock apparent exposure ages of 43.6 ± 1.2 ^{10}Be kyr at Alddasčorru (Fig. 3c) and 60.7 ± 1.2 ^{10}Be kyr from a nearby relict patch at Olmáčohkka (Fig. 2) are considerably older than this. Results from three samples collected approximately 35 km further south at Fávrratčohkka (Fig. 2) provide almost identical results. A bedrock sample from the transverse lee-side scarp at the distal end of Fávrratčohkka hill yielded deglaciation ages of 11.0 ± 1.8 ^{10}Be kyr and 8.8 ± 1.5 ^{26}Al kyr. There was no bedrock exposed in the relict patch upstream of this transverse lee-side scarp, however, a bedrock sample from the relict patch at Áilladis (Fig. 2), a setting very similar and close to Fávrratčohkka, provided an apparent age of 51.1 ± 1.5 ^{10}Be kyr. Bedrock collected from a tor exposed in a relict patch at Urttičohkka, $\sim$450 m higher and 10 km west of Áilladis, yielded an apparent age of 40.3 ± 1.3 ^{10}Be kyr.

Similar results were obtained at Tjuolmma, a gently sloping relict surface rising about 100 m above the surrounding Ultevis Plateau. The relict surface at Tjuolmma is truncated on the distal side by a transverse lee-side scarp. Bedrock exposed at the transverse lee-side scarp and an erratic deposited on the relict surface at the summit of Tjuolmma gave deglaciation ages of 9.7 ± 1.5

Table 1
Cosmogenic nuclide data and apparent exposure ages

Sample location and elevation (masl)	Sample type	Correction factor	^{10}Be $\times 10^5$ atom g^{-1}	^{26}Al $\times 10^5$ atom g^{-1}	Apparent exposure age ^{10}Be kyr	^{26}Al kyr
Alddasčorru (1380)	erratic	0.98	0.63 ± 0.09		12.5 ± 1.7 (2.3)	
Alddasčorru (1380)	bedrock in relict patch	0.98	2.20 ± 0.06		43.6 ± 1.2 (5.8)	
Olmáčohkka (1355)	bedrock in relict patch	0.98	3.05 ± 0.06		60.7 ± 1.2 (8.0)	
Fávrratčohkka (850)	transverse lee-side scarp	0.96	0.56 ± 0.09	2.71 ± 0.46	11.0 ± 1.8 (2.3)	8.8 ± 1.5 (2.4)
Áilladis (910)	bedrock in relict patch	0.98	2.58 ± 0.08		51.1 ± 1.5 (6.8)	
Urttičohkka (1360)	bedrock in relict patch	0.95	2.04 ± 0.07		40.3 ± 1.3 (5.4)	
Tjuolmma (945)	erratic	0.99	0.39 ± 0.05		7.6 ± 1.0 (1.4)	
Tjuolmma (930)	transverse lee-side scarp	0.98	0.49 ± 0.07		9.7 ± 1.5 (1.9)	
Tjuolmma (920)	bedrock in relict patch	0.96	1.70 ± 0.07	7.64 ± 0.84	33.6 ± 1.5 (4.6)	24.9 ± 2.7 (5.9)
Tjuolmma-R (920)	bedrock in relict patch	0.96	1.83 ± 0.06		36.1 ± 1.3 (4.9)	
Bálddavárri (900)	erratic	0.97	0.61 ± 0.05		12.0 ± 0.9 (1.8)	
Bálddavárri (810)	erratic	0.94	0.66 ± 0.06^{A}		12.9 ± 1.2 (2.1)	
Bálddavárri (890)	bedrock	0.97	2.44 ± 0.06^{A}		48.4 ± 1.1 (6.4)	

All latitudes $>67°$N. Site locations as in Fig. 2. Correction factor includes geometric shielding [44] and sample thickness using an attenuation coefficient of 160 ± 10 g cm^{-2} and rock density of 2.8 g cm^{-3}. ^{26}Al/^{27}Al and ^{10}Be/^{9}Be measured by accelerator mass spectrometry at the Purdue Rare Isotope Measurement Laboratory (PRIME Lab) and the Australian National Tandem Accelerator for Applied Research (ANTARES) at the Australian Nuclear Science and Technology Organisation (ANSTO)^{A}. Procedural blanks were used to correct measured ratios. Total Al measured by ICP-AES and assigned 5% uncertainty. ^{10}Be and ^{26}Al concentrations normalized to sea level and high latitude ($>60°$) using [26] with standard 15°C sea level temperature. Apparent ages calculated assuming zero erosion and sea level and high latitude ^{10}Be and ^{26}Al production rates of 5.1 ± 0.3 atom g^{-1} yr^{-1} and 31.1 ± 1.9 atom g^{-1} yr^{-1} respectively [26]. Uncertainties represent one standard error measurement uncertainty, with random (8%) and systematic uncertainties (10%) added in quadrature and shown in parentheses [16].

^{10}Be kyr and 7.6 ± 1.0 ^{10}Be kyr. Two separate aliquots of the quartz derived from bedrock sampled in the relict area were processed and yielded apparent ages of 33.6 ± 1.5 ^{10}Be kyr and 36.1 ± 1.3 ^{10}Be kyr with a weighted mean of 35.1 ± 1.0 ^{10}Be kyr (Table 1). The ^{26}Al apparent age for the first aliquot is 24.9 ± 2.7 ^{26}Al kyr providing a ^{26}Al/^{10}Be ratio of 4.5 ± 0.5.

The final three samples were collected at Bálddavárri, an area previously described as having been extensively scoured by ice (see fig. 6 in [8]). Two erratics yielded deglaciation ages of 12.0 ± 0.9 ^{10}Be kyr and 12.9 ± 1.2 ^{10}Be kyr, while the apparent age of the bedrock sample was 48.4 ± 1.1 ^{10}Be kyr.

4. Discussion

Erratics and transverse lee-side scarps were sampled to determine deglaciation ages. The ages range from 7.6 ± 1.0 ^{10}Be kyr to 12.9 ± 1.2 ^{10}Be kyr with a weighted mean of 10.7 ± 0.5 ^{10}Be kyr. This is consistent with the accepted deglaciation age of ~10 ^{14}C kyr for this region [28], however, the age range suggests temporal variations in deglaciation between different sites. For example, at Tjuolmma and Bálddavárri the independent samples collected at each site are internally consistent (Table 1) with weighted mean ages 8.2 ± 0.8 ^{10}Be kyr and 12.3 ± 0.7 ^{10}Be kyr respectively, suggesting that Bálddavárri was ice free before Tjuolmma. Although determining the detailed pattern of deglaciation warrants further work, for the purpose of this paper the deglaciation ages clearly indicate that our sample sites were covered by ice during the last glacial cycle.

The apparent exposure ages from bedrock in relict patches are significantly older than the adjacent exposure ages from transverse lee-side scarps and overlying or nearby erratics. Given that the relict surfaces with ages of ~35–61 kyr have erratics on them with ages of ~8–13 kyr it is not possible that the apparent ages of the relict patches represent actual time elapsed since deglaciation. This therefore indicates that the ice sheet during last glaciation eroded an insufficient thickness of bedrock, weathering mantle, or till to remove the cosmogenic nuclide inventory accumulated during previous exposure periods. The duration of these previous exposure periods is difficult to ascertain because we do not know how much erosion and/or shielding the sample sites experienced. Erosion reduces the cosmogenic nuclide concentration through removal of material from the surface. Likewise, decay of ^{10}Be and ^{26}Al during periods of shielding by ice or till reduces the existing ^{10}Be and ^{26}Al inventory [19,21]. Although the range of apparent exposure ages from relict patches reflects site-specific differences in erosion or shielding, our data do not permit quantification of either. Therefore, the apparent ages, calculated assuming no erosion and/or shielding, are minimum estimates of the total exposure experienced by individual relict patches and clearly demonstrate that every sampled relict area has survived at least the last glacial maximum.

By adopting the reconstructed glacial history of the region [8,29–31] it is possible to broadly determine if relict patches have survived more than one glacial cycle. The ice occupation history of northern Sweden has been divided into periods of Mountain ice sheet and Fennoscandian ice sheet configurations [8,29], with the duration of each period broadly determined from the DSDP 607 marine benthic foraminifer oxygen isotope record of global ice volume [32–34]. All our sample sites are located in the region covered by ice during Mountain ice sheet periods. Therefore we adopt the δ^{18}O limits used by [8] for this style of ice sheet to divide the DSDP 607 record into periods of ice cover and ice free conditions for our sites (Fig. 4). According to this time scale, the youngest relict patch at Tjuolmma, with a weighted mean apparent age of 35.1 ± 1.0 ^{10}Be kyr, was first exposed at ~200 kyr during marine isotope stage (MIS) 7 (Fig. 4). Thus relict areas have survived at least two complete glacial cycles with minimal alteration (Fig. 4). Although the validity of slicing the DSDP 607 record in this fashion could be debated for periods extending beyond the last glacial cycle (the time for which there has been good temporal and spatial control on ice sheet extent), the measured cosmogenic nuclide concentrations for all the relict patches re-

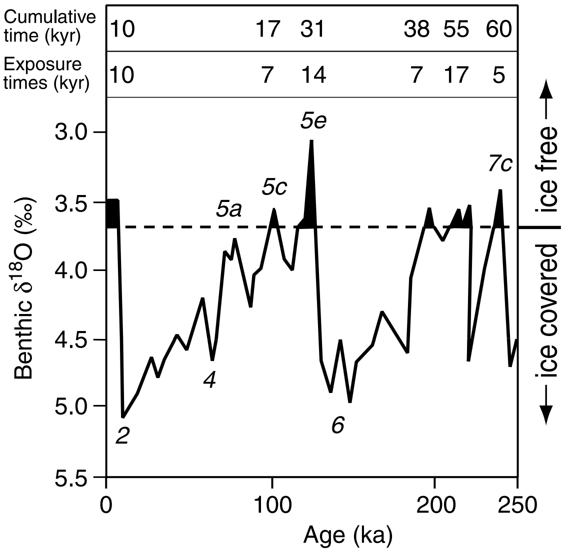

Fig. 4. Marine benthic oxygen isotope record for the last 250 kyr from DSDP 607 [34] with marine oxygen isotope stages (MIS) labelled in italics. In the reconstructed ice occupation history for the northern Swedish mountains ice sheets occur when $\delta^{18}O > 3.7$‰ [8]. Surface exposure only occurs during individual ice free periods (black-shaded areas above the dashed line). Durations of the ice free periods and cumulative exposure times are shown along the top of the figure. Exposure ages greater than 10.7 ± 0.5 kyr (weighted mean of youngest values in Table 1) represent cumulative exposure over multiple ice free periods. For example, to accumulate the measured minimum 43.6 ± 1.2 ^{10}Be kyr at the Alddasčorru site (Fig. 3) requires a minimum total history of ~ 210 kyr. The bedrock age indicates survival during the growth and decay of the Fennoscandian ice sheet even if this site was only covered during times of greatest ice sheet extent ($\delta^{18}O > 4.5$‰ [8]).

quire their preservation at least since before the development of the most recent Fennoscandian ice sheet at about 65 kyr (MIS 4). Additional evidence for preservation through multiple rather than just the last glaciation comes from lowland tor–summit surface samples collected further east in the Parkajoki area. These samples yielded $^{26}Al/^{10}Be$ ratios indicating minimum total exposure and shielding histories in excess of 600 kyr [22]. Preservation of this area is attributed primarily to cold-based conditions, though recognizing that its location was beneath or close to the last glacial maximum ice divide, preservation may also have been due to negligible horizontal ice flow, and hence erosive capacity, over the area during this and previous maximum positions of Fennoscandian ice sheets [35]. The results reported in this study come from samples collected in areas located beneath the ice divide during Mountain ice sheet glaciations and distant from the ice divide during Fennoscandian glaciation. Therefore it is likely that the sample sites experienced increased ice velocity regimes during the growth and decay phases of Fennoscandian ice sheets. Nevertheless, the single nuclide data from the relict patches indicate that glacial erosion was limited, suggesting frozen bed conditions. Furthermore, using the measured ^{10}Be and ^{26}Al concentrations for the Tjuolmma sample and a simple cosmogenic nuclide accumulation and decay model that assumes exposure prior to a single shielding period under thick ice with subsequent re-exposure since the last deglaciation, constrains the complex history for this site to 845 kyr (Table 2). This is a minimum total history since it does not take into account the multiple periods of exposure and shielding by ice which have affected this area throughout the Quaternary [22]. Our result corroborates the hypothesis that relict patches have survived multiple glaciations even in these areas.

Surprisingly, the bedrock result from Báldda-várri, where glacial scouring has obliterated any characteristic preglacial morphology and left behind rock basins and steep glacial facets (see fig. 6 in [8]), also shows cosmogenic nuclide inheritance. Thus even in an area that appears to have been severely scoured the last glaciations did not significantly modify the landscape. This has important implications for regional reconstructions of erosion patterns (e.g. Sugden, 1977) because the total area of glacial scouring in any one region may be the cumulative result of several cycles of glacial erosion, where each cycle only eroded spatially restricted areas. Hence it may be difficult to morphologically discern 'young' (last glacial cycle) scouring from scouring related to older glacial events, and care needs to be taken when trying to interpret subglacial temperature conditions and deglaciation patterns of the last Fennoscandian and other paleo ice sheets on the basis of landforms alone. Another implication of this pattern and extent of landscape preservation is a potentially significant reduction of effective sediment

Table 2
Complex history model results for bedrock from the relict area at Tjuolmma

	^{26}Al $\times10^5$ atom g^{-1}	^{10}Be $\times10^5$ atom g^{-1}	^{26}Al/^{10}Be	Exposure prior to shielding (kyr)	Shielding (kyr)	Exposure since deglaciation (kyr)	Total history (kyr)
Mean ratio	7.64	1.70	4.5	36.7	800	8.2	845
Minimum ratio	6.81	1.78	3.8	47.8	1250	8.2	1306
Maximum ratio	8.48	1.63	5.2	28.6	390	8.2	427

Exposure time prior to shielding (T) and shielding time (B) were inferred by solving:

$$\frac{N_{26}}{N_{10}} = \frac{\frac{P_{26}}{\lambda_{26}}\cdot(1-e^{-\lambda_{26}T})\cdot e^{-\lambda_{26}(B+t)} + \frac{P_{26}}{\lambda_{26}}\cdot(1-e^{-\lambda_{26}t})}{\frac{P_{10}}{\lambda_{10}}\cdot(1-e^{-\lambda_{10}T})\cdot e^{-\lambda_{10}(B+t)} + \frac{P_{10}}{\lambda_{10}}\cdot(1-e^{-\lambda_{10}t})} \quad (1)$$

where the subscripts denote ^{26}Al and ^{10}Be, N is the measured nuclide concentration and t is the exposure time since deglaciation. The production rates P_{26} and P_{10} are given in Table 1. The decay constants λ_{26} and λ_{10} are 4.56×10^{-7} yr^{-1} and 9.76×10^{-7} yr^{-1} respectively.

source areas used in estimates of landscape change from sediment delivery records in ocean and lake cores [36], and in interpreting offshore sediment records such as provenance studies related to Heinrich events [37,38]. Additionally, given previous work indicating that viable seeds might survive beneath an ice sheet [39–41], the presence of persistent preserved patches raises the possibility that some of these acted as subglacial refugia [41] and thus could have been source areas for postglacial biologic re-colonization.

Cosmogenic radionuclide exposure ages from palimpsest landscapes in the northern Swedish mountains indicate that ice sheet basal thermal conditions throughout two or more glacial cycles resulted in areas of extensive glacial erosion juxtaposed to areas with landscape preservation. The quantitative demonstration that relict landscapes were preserved subglacially, requires that boundaries between glacially sculpted and preserved landscapes in other areas should not automatically be interpreted as former ice limits in paleoclimatic and paleoglaciological reconstructions (e.g. [42] and references therein). The areas of relict landscapes that can be mapped today represent a minimum estimate of the extent of frozen patches [43]. Surprisingly, relict areas persisted through multiple ice sheet growth and decay phases despite expected changes in flow patterns and basal thermal regimes [11]. The persistence of cold-based conditions and related landscape preservation, which likely applies to other large paleo ice sheets such as the Laurentide ice sheet [6,42], indicates the need to incorporate frozen bed patches into basal boundary conditions for ice sheet models [5].

Acknowledgements

Field and analytical work were primarily supported by the Swedish Natural Science Research Council through grant G-AA/GU-12034-301 and the US National Science Foundation through grant OPP-9818162. We thank Dr. S. Porter, Dr. R. Oglesby, Dr. R. Randall, Dr. T. Swinehart, Dr. D. Shepardson, and Dr. W. Lawson for their constructive comments on an earlier version of the manuscript. Comments received by Dr. T. Dunai, Dr. D.Q. Bowen, and Dr. J. Karhu helped to improve the final manuscript.*[SK]*

References

[1] T.G. Bonney, On the formation of cirques and their bearing upon theories attributing the excavation of alpine valleys mainly to the action of glaciers, Q. J. Geol. Soc. 27 (1871) 312–324.
[2] E.J. Garwood, Features of alpine scenery due to glacial protection, Geogr. J. 36 (1910) 310–339.
[3] R.F. Flint, Glacial Geology and the Pleistocene Epoch, Wiley, New York, 1947, 589 pp.

[4] D.E. Sugden, B.S. John, Glaciers and Landscape: A Geomorphological Approach, Arnold Pub. Co., London, 1976, 376 pp.
[5] J. Kleman, C. Hättestrand, A. Clarhäll, Zooming in on frozen-bed patches: scale-dependent controls on Fennoscandian ice sheet basal thermal zonation, Ann. Glaciol. 28 (1999) 189–194.
[6] D.E. Sugden, S.H. Watts, Tors, felsenmeer, and glaciation in northern Cumberland Peninsula, Baffin Island, Can. J. Earth Sci. 14 (1977) 2817–2823.
[7] A.M. Hall, D.E. Sugden, Limited modification of mid-latitude landscapes by ice-sheets: the case of northeast Scotland, Earth Surf. Process. Landf. 12 (1987) 531–542.
[8] J. Kleman, A.P. Stroeven, Preglacial surface remnants and Quaternary glacial regimes in northwestern Sweden, Geomorphology 19 (1997) 35–54.
[9] A.P. Stroeven, D. Fabel, J. Harbor, C. Hättestrand, J. Kleman, Reconstructing the erosion history of glaciated passive margins: Applications of in-situ produced cosmogenic nuclide techniques, in: A.G. Doré, J. Cartwright, M.S. Stoker, J.P. Turner, N. White (Eds.), Exhumation of the North Atlantic Margin: Timing, Mechanisms and Implications for Petroleum Exploration, Geological Society, London, Special Publications 196, Geological Society, London, in press.
[10] E.J. Brook, A. Nesje, S.J. Lehman, G.M. Raisbeck, F. Yiou, Cosmogenic nuclide exposure ages along a vertical transect in western Norway: implications for the height of the Fennoscandian ice sheet, Geology 24 (1996) 207–210.
[11] J. Kleman, C. Hattestrand, I. Borgström, A. Stroeven, Fennoscandian palaeoglaciology reconstructed using a glacial geological inversion model, J. Glaciol. 43 (1997) 283–299.
[12] R. Lagerbäck, The Veiki moraines in northern Sweden - widespread evidence of an early Weichselian deglaciation, Boreas 17 (1988) 469–486.
[13] G.H. Denton, T.J. Hughes, The Last Great Ice Sheets, Wiley-Interscience, New York, 1981, 484 pp.
[14] K. Lambeck, J. Chappell, Sea level change through the last glacial cycle, Science 292 (2001) 679–686.
[15] C.E. Allen, R.G. Darmody, C.E. Thorn, J.C. Dixon, P. Schlyter, Clay mineralogy, chemical weathering and landscape evolution in Arctic-Alpine Sweden, Geoderma 99 (2001) 277–294.
[16] J.C. Gosse, F.M. Phillips, Terrestrial in situ cosmogenic nuclides: theory and application, Quat. Sci. Rev. 20 (2001) 1475–1560.
[17] D. Lal, Cosmic-ray labeling of erosion surfaces: in situ nuclide production rates and erosion models, Earth Planet. Sci. Lett. 104 (1991) 424–439.
[18] K. Nishiizumi, C.P. Kohl, J.R. Arnold, R. Dorn, J. Klein, D. Fink, R. Middleton, D. Lal, Role of in situ cosmogenic nuclides ^{10}Be and ^{26}Al in the study of diverse geomorphic processes, Earth Surf. Process. Landf. 18 (1993) 407–425.
[19] D. Fabel, J. Harbor, The use of in-situ produced cosmogenic radionuclides in glaciology and glacial geomorphology, Ann. Glaciol. 29 (1999) 103–110.
[20] D. Fabel, J. Stone, L.K. Fifield, R.G. Cresswell, Deglaciation of the Vestfold Hills, East Antarctica: Preliminary evidence from exposure dating of three subglacial erratics, in: C.A. Ricci (Ed.), The Antarctic Region: Geological Evolution and Processes, Terra Antarctica Publication, Siena, 1997, pp. 829–834.
[21] P.R. Bierman, K.A. Marsella, C. Patterson, P.T. Davis, M. Caffee, Mid-Pleistocene cosmogenic minimum-age limits for pre-Wisconsinan glacial surfaces in southwestern Minnesota and southern Baffin Island: a multiple nuclide approach, Geomorphology 27 (1999) 25–39.
[22] A.P. Stroeven, D. Fabel, C. Hättestrand, J. Harbor, A relict landscape in the centre of Fennoscandian glaciation: cosmogenic radionuclide evidence of tors preserved through multiple glacial cycles, Geomorphology 44 (2002) 145–154.
[23] A. Clarhäll, J. Kleman, Distribution and glaciological implications of relict surfaces on the Ultevis plateau, northwestern Sweden, Ann. Glaciol. 28 (1999) 202–208.
[24] J. Kleman, I. Borgström, Glacial land forms indicative of a partly frozen bed, J. Glaciol. 40 (1994) 255–264.
[25] C.P. Kohl, K. Nishiizumi, Chemical isolation of quartz for measurement of in situ-produced cosmogenic nuclides, Geochim. Cosmochim. Acta 56 (1992) 3586–3587.
[26] J.O. Stone, Air pressure and cosmogenic isotope production, J. Geophys. Res. 105 (2000) 23753–23759.
[27] T.J. Dunai, Scaling factors for the production rates of in-situ produced cosmogenic nuclides: a critical reevaluation, Earth Planet. Sci. Lett. 176 (2000) 159–171.
[28] B.E. Berglund, L. Barnekow, D. Hammarlund, P. Sandgren, I.F. Snowball, Holocene forest dynamics and climate changes in the Abisko area, northern Sweden - the Sonesson model of vegetation history reconsidered and confirmed, Ecol. Bull. 45 (1996) 15–30.
[29] C. Hättestrand, The glacial geomorphology of central and northern Sweden, Sver. Geol. Unders. Ca 85 (1998) 47.
[30] E. Jansen, T. Fronval, F. Rack, J.E.T. Channell, Pliocene–Pleistocene ice rafting history and cyclicity in the Nordic Seas during the last 3.5 Myr, Paleoceanography 15 (2000) 709–721.
[31] G.S. Boulton, P. Dongelmans, M. Punkari, M. Broadgate, Palaeoglaciology of an ice sheet through a glacial cycle: the European ice sheet through the Weichselian, Quat. Sci. Rev. 20 (2001) 591–625.
[32] M.E. Raymo, W.F. Ruddiman, J. Backman, B.M. Clement, D.G. Martinson, Late Pliocene variations in Northern Hemisphere ice sheets and North Atlantic Deep Water circulation, Palaeoceanography 4 (1989) 413–446.
[33] W.F. Ruddiman, M.E. Raymo, D.G. Martinson, B.M. Clement, J. Backman, Pleistocene evolution: Northern Hemisphere ice sheets and North Atlantic Ocean, Palaeoceanography 4 (1989) 353–412.
[34] D. Lazarus, C. Spencer-Cervato, M. Pika-Biolzi, J.P. Beckmann, K. von Salis, H. Hilbrecht, H. Thierstein, Revised chronology of Neogene DSDP Holes from World

Ocean, Ocean Drilling Program, College Station, TX, 1995, 301 pp.

[35] C. Hättestrand, A.P. Stroeven, A relict landscape in the centre of Fennoscandian glaciation: Geomorphological evidence of minimal Quaternary glacial erosion, Geomorphology 44 (2002) 127–143.

[36] N.F. Glasser, A.M. Hall, Calculating Quaternary glacial erosion rates in northeast Scotland, Geomorphology 20 (1997) 29–48.

[37] R.H. Gwiazda, S.R. Hemming, W.S. Broecker, Provenance of icebergs during Heinrich event 3 and the contrast to their sources during other Heinrich episodes, Paleoceanography 11 (1996) 371–378.

[38] S.R. Hemming, T.O. Vorren, J. Kleman, Provinciality of ice rafting in the North Atlantic: application of $^{40}Ar/^{39}Ar$ dating of individual ice rafted hornblende grains, Quat. Int., in press, 2002.

[39] A.E. Porsild, C.R. Harington, G.A. Mulligan, *Lupinus arcticus* Wats, grown from seeds of Pleistocene age, Science 158 (1967) 113–114.

[40] B.M. Bergsma, J. Svoboda, B. Freedman, Entombed plant communities released by a retreating glacier at central Ellesmere Island, Canada, Arctic 37 (1984) 49–52.

[41] J. Kleman, Preservation of landforms under ice sheets and ice caps, Geomorphology 9 (1994) 19–32.

[42] G.H. Miller, A.P. Wolfe, E.J. Steig, P.E. Sauer, M.R. Kaplan, J.P. Briner, The Goldilocks dilemma: big ice, little ice, or 'just-right' ice in the Eastern Canadian Arctic, Quat. Sci. Rev. 21 (2002) 33–48.

[43] J. Kleman, C. Hättestrand, Frozen-bed Fennoscandian and Laurentide ice sheets during the Last Glacial Maximum, Nature 402 (1999) 63–66.

[44] J. Dunne, D. Elmore, P. Muzikar, Scaling factors for the rates of production of cosmogenic nuclides for geometric shielding and attenuation at depth on sloped surfaces, Geomorphology 27 (1999) 3–11.