Variation in glacial erosion near the southern margin of the Laurentide Ice Sheet, south-central Wisconsin, USA: Implications for cosmogenic dating of glacial terrains

Patrick M. Colgan[†]
Department of Geology, Northeastern University, 14 Holmes Hall, Boston, Massachusetts 02115, USA

Paul R. Bierman[‡]
Department of Geology and School of Natural Resources, University of Vermont, Burlington, Vermont 05405, USA

David M. Mickelson[§]
Department of Geology and Geophysics, University of Wisconsin, Weeks Hall, 1215 West Dayton Street, Madison, Wisconsin 53706, USA

Marc Caffee[#]
Purdue Rare Isotope Measurement Laboratory, Department of Physics, Purdue University, 1396 Physics Building, West Lafayette, Indiana 47907-1396, USA

ABSTRACT

We measured the abundance of cosmogenic ^{10}Be and ^{26}Al in 22 samples collected from five striated granite, metarhyolite, and quartzite outcrops in south-central Wisconsin that were covered by the late Wisconsin Laurentide Ice Sheet. In two outcrops, measured nuclide abundances are consistent with the existing radiocarbon chronology of ice retreat. In three outcrops, nuclide abundances were up to eight times higher than predicted by the radiocarbon chronology. At these three sites, several thousand years of ice flow eroded only centimeters to decimeters of rock, allowing a significant quantity of nuclides (10^5–10^6 atoms of ^{10}Be and ^{26}Al per gram of quartz), produced during prior periods of exposure, to remain. We calculate minimum-limiting glacial erosion rates of 0.01–0.25 mm·yr^{-1} for these rocks.

Rock properties, sample location on outcrops, and outcrop proximity to the former ice margin control the magnitude of cosmogenic nuclides inherited from periods of prior exposure. Four of five samples from very hard metarhyolite outcrops with widely spaced joints contain inherited nuclides; two samples carry the equivalent of >150,000 yr of surface exposure, even though they were covered by ice during the last-glacial-maximum advance. Samples highest on the landscape or in plucked areas have less inheritance than those from the lee sides of large hills or lower in the landscape. Three quartzite samples collected ~10 km up-ice from the margin contain three to four times the expected nuclide abundance (10^5 to 10^6 atoms per gram of quartz). In contrast, eight other quartzite and granite samples from two outcrops >50 km up-ice from the former margin contain only 10^5 atoms of ^{10}Be per gram of quartz, consistent with late Pleistocene exposure and little, if any, nuclide inheritance. This relationship between glacial erosion and distance from the former terminus is consistent with a marginal zone of minimal subglacial erosion; the ice was either frozen to its bed there, or the ice thickness and duration of ice cover were less near the terminus.

These data, together with simple modeling of nuclide production by deeply penetrating muons, suggest that many meters of rock must be removed to reduce inheritance to negligible levels (<1000 yr) in continental terrains with low long-term erosion rates. Our results indicate that cosmogenic dating of exposed bedrock surfaces near former ice margins or in areas where ice was frozen to the bed may be uncertain, and in some cases impossible, because nuclides are inherited from prior periods of cosmic-ray exposure. Unfortunately, striated glacial outcrops that can be used to demonstrate most easily the assumption of "no postglacial erosion" are also the most likely to have undergone little glacial erosion. This finding suggests that cosmogenic-nuclide production rates based on glacially striated surfaces may include cosmogenic nuclides inherited from prior exposure.

Keywords: cosmogenic isotopes, dating, exposure age, glacial erosion.

INTRODUCTION

In situ–produced cosmogenic nuclides are widely used for dating deglaciation. Since Phillips et al. (1990) assigned ages to moraines at Bloody Canyon in the Sierra Nevada by using measured ^{36}Cl activities, numerous studies have used ^{3}He, ^{10}Be, ^{26}Al, and ^{36}Cl to date glacial landforms in areas where organic matter is scarce (e.g., Brook and Kurz, 1993; Gosse et al., 1995a, 1995b; Phillips et al., 1996; Brook et al., 1996; Davis et al., 1999; Fabel and Harbor, 1999; Marsella et al., 2000). Interpretive models used to estimate exposure ages from measured nuclide abundances usually assume that the entire nuclide inventory was produced during the last period of surface

[†]E-mail: p.colgan@neu.edu.
[‡]E-mail: pbierman@zoo.uvm.edu.
[§]E-mail: davem@geology.wisc.edu.
[#]E-mail: mcaffee@physics.purdue.edu.

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exposure (Bierman, 1994). Nuclides carried over from prior periods of surface or near-surface exposure are termed inherited, and if not considered explicitly, result in inflated model age estimates.

Understanding the magnitude of inheritance is becoming increasingly important as isotopic measurements become more precise and the need for accurate age estimates becomes more pressing. For example, Gosse et al. (1995b) interpreted highly precise ^{10}Be abundances in morainal boulders as evidence of deposition during the Younger Dryas climatic oscillation, an interpretation requiring accuracy on the order of 1000 yr. In this and other cosmogenic studies, inheritance has been assumed to be inconsequential and is disregarded in model age calculations, despite early evidence that inheritance could be significant (Nishiizumi et al., 1989).

This paper demonstrates that inheritance of cosmogenic nuclides from periods of prior exposure occurs in midlatitude terrains affected by continental glaciation and that such inheritance can be significant, especially for hard rocks near former ice margins. We propose that one of the factors controlling the pattern of inheritance is the spatial location of erosion and deposition zones within the former ice lobe. This result suggests that caution is necessary when using glacially striated surfaces to estimate production rates of cosmogenic nuclides.

Cosmogenic Dating and the No-Inheritance Assumption

The potential for cosmogenic-nuclide inheritance arises from the ability of cosmic rays to penetrate rock to a considerable depth. Early on, it was realized that there are multiple nuclear pathways for the in situ production of cosmogenic nuclides (Lal, 1988). Fast-neutron spallation, which is responsible for the majority of nuclide production, diminishes rapidly with depth (Lal, 1988). Fast-neutron attenuation is modeled well by a decreasing exponential function with a characteristic attenuation length (Λ or $1/e$) of 160 g·cm^{-2}. This value corresponds to ~60 cm of rock (assuming density $\rho = 2.7$ g·cm^{-3}). If spallation were the only production pathway, then only ~2 m of rock must be removed to eliminate most inheritance (Fig. 1).

Although nuclide production by muons is usually a few percent of neutron production at the surface, it extends much deeper (Λ = 1300 g·cm^{-2}) and dominates nuclide production below 2–3 m depth (Fig. 1; Stone et al., 1998a; Brown et al., 1995; Granger and Smith, 2000). For some nuclides, such as ^{36}Cl produced from Ca, near-surface muon production is very important (>25%) and if disregarded will lead to significant changes in estimates of erosion rate and age (Stone et al., 1998a). The exact contribution of muons to production of ^{10}Be and ^{26}Al is less certain. Field and laboratory studies suggest that ~3% of ^{10}Be and ^{26}Al production on surfaces at sea level and high latitude is caused by muons (Brown et al., 1995).

The concentration of nuclides at depth is a function of the erosion and exposure history of an outcrop over a time frame equivalent to several effective half-lives. The effective half-life combines loss from both erosion and decay of nuclides (Lal, 1988; Gillespie and Bierman, 1995) and thus is shorter, by varying amounts, than the decay half-life. Therefore, it is difficult to generalize the specific depth of glacial erosion at which inheritance ceases to be a concern. Rather, each case must be evaluated individually, considering the total cosmic-ray exposure prior to glaciation, the depth of glacial erosion, rate of interglacial erosion, cover by unconsolidated sediments during ice-free periods, and the magnitude of cosmic-ray exposure since the last deglaciation.

In some studies, inheritance has been shown to be significant. In the Sierra Nevada, Nishiizumi et al. (1989) found excess ^{10}Be and ^{26}Al in ablation-zone granite outcrops polished and striated by the latest Pleistocene valley glacier at June Lake. On the highlands of Baffin Island, Bierman et al. (1999) found that glaciated gneissic bedrock surfaces had complex histories consistent with exposure, burial, and reexposure without significant erosion. Briner and Swanson (1998) found significant inheritance of ^{36}Cl in bedrock samples collected from a basalt roche moutonnée in the northern Puget Lowland from which they were able to set limits on glacial erosion rates.

Other studies suggest that inheritance in some locations is insignificant. Cosmogenic ^{36}Cl was just above detection limits in granodiorite boulders sampled from roadcuts in Sierra Nevada moraines (Bierman, 1993; Bierman et al., 1995). In contrast to the highlands of Baffin Island, numerous ^{10}Be and ^{26}Al measurements in samples collected from recently deglaciated, fjord-bottom, gneiss surfaces show that Baffin Island valley glaciers have removed so much rock that most measurements were below detection limits (<1000 yr near sea level and high latitude; Davis et al., 1999). Within several kilometers of the hilly late Pleistocene margin of the Laurentide Ice Sheet in New Jersey, neither quartzite nor

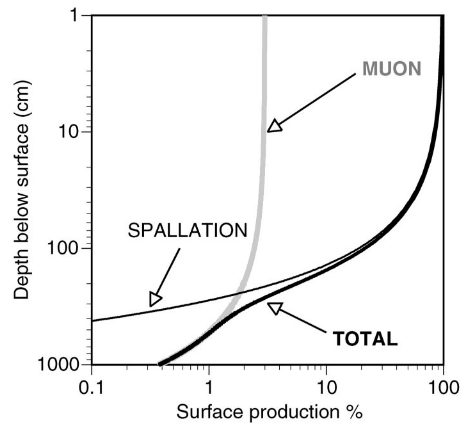

Figure 1. Relative production rates of ^{10}Be and ^{26}Al with depth in rock (ρ = 2.7 g·cm^{-3}). Heavy solid line is total production. Thin solid line is spallation production (Λ = 160 g·cm^{-2}). Light gray line is muon-induced production (Λ = 1300 g·cm^{-2}) modeled with muons producing 3% of nuclides at the rock surface.

gneiss bedrock surfaces contain significant inherited ^{10}Be (Clark et al., 1995; Larsen, 1996).

Uncertainty in production rates of cosmogenic nuclides as a function of time, altitude, and latitude add systematic errors to exposure ages and continue to confound rigorous comparison with other dating systems (Clark et al., 1995). What is most important, because production rates of cosmogenic nuclides used by most investigators have been calibrated from isotope abundance measured in samples collected from glaciated surfaces (Nishiizumi et al., 1989; Clark et al., 1995; Stone et al., 1998b), inheritance may also affect (inflate) production-rate estimates.

METHODS

Sample Site Selection

We analyzed 22 samples from 5 outcrops along a transect roughly parallel to flow of the Green Bay lobe at the last glacial maximum (Fig. 2). We chose these sites in order to test the current deglaciation chronology for the Green Bay lobe (Table 1 and Fig. 3). Deglaciation probably began sometime before 15,000 cal. yr B.P. (Clayton et al., 1992; Maher and Mickelson, 1996). By ~14,000 cal. yr B.P., ice had retreated enough to expose all of the sites sampled in this study.

Resistant Precambrian quartzite, metarhyolite, and granite crop out only in limited locations in the area once covered by the Green Bay lobe. Lower Paleozoic sandstone, dolo-

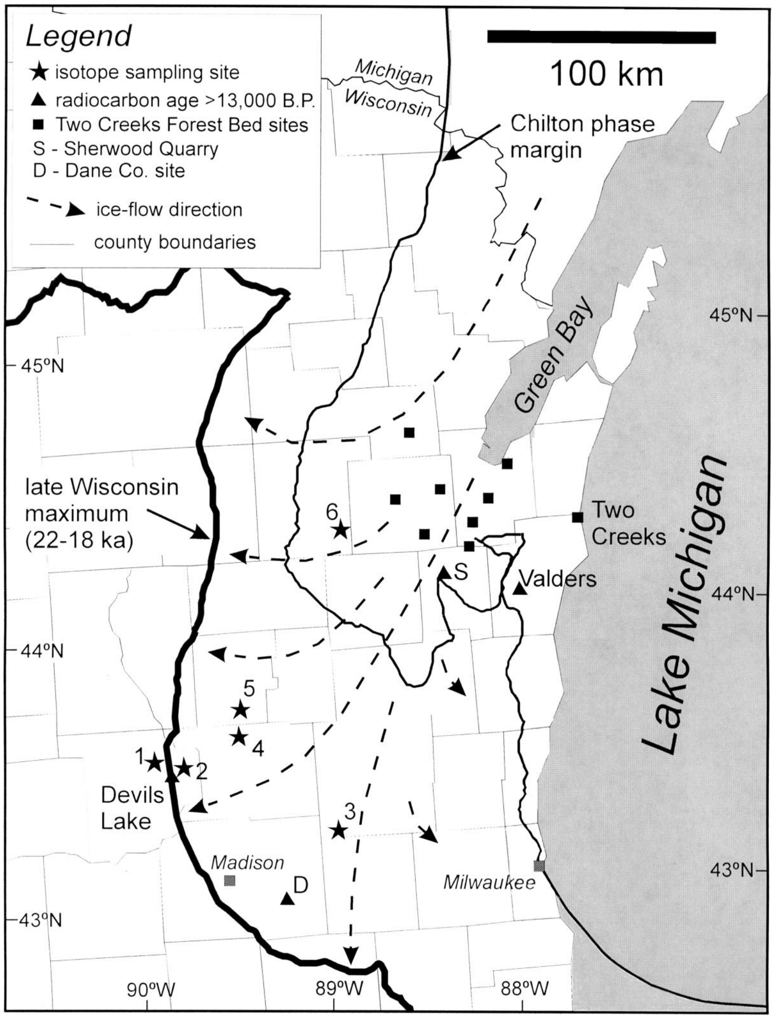

Figure 2. Sampling sites and important sections from which organic material has been radiocarbon dated in southeastern Wisconsin. 1—Rock Springs Quarry site (10 km outside margin); 2—Baraboo Hills site (10 km inside margin); 3—Waterloo site (55 km inside margin); 4—Bush and Katsma Rocks site (30 km inside margin); 5—Observatory Hill site (30 km inside margin); 6—Cactus Rock site (70 km inside margin); Squares are sites where an equivalent of the Two Creeks forest bed has been identified. Flow lines are based on drumlin orientations.

mite, and shale underlie most of the deglaciated surface. We sampled five surface outcrops and the wall of an abandoned quarry: Baraboo Hills (quartzite), Waterloo (quartzite), Bush and Katsma Rocks (metarhyolite), Observatory Hill (metarhyolite), Cactus Rock (granite), and Rock Springs Quarry (quartzite) (Figs. 1, 3, and 4; Table 2; and GSA Data Repository[1]). The outcrops were selected because they show evidence of glacial erosion, including streamlined and rounded outcrop form, evidence of glacial plucking (steep leeside slopes), and abrasion features such as striae, glacial polish, grooves, ventifacts, and chattermarks. The preservation of striae and other abrasion features demonstrates that there has been little postglacial erosion of these sites. The sites were also selected because they are topographically high in the landscape (10–60 m above the average terrain's elevation) and probably have had little loess, till, or snow cover in the time since deglaciation (detailed site description available [see Data Repository]). Some of the sites may have had a thin till cover just after deglaciation, but we found no evidence for this possibility in residual boulders or gravel on the surface.

[1]GSA Data Repository item 2002138, more information about sample preparation, analysis, and site descriptions, is available on the Web at http://www.geosociety.org/pubs/ft2002.htm. Requests may also be sent to editing@geosociety.org.

Sampling and Analytical Methods

Twenty-two samples, all >1 kg, were collected from the upper 5 cm of each outcrop by using a hammer and chisel (Fig. 4). Sample sites were photographed and located by using both topographic maps and the Global Positioning System. Samples were returned to the University of Vermont where they were crushed, ground, and sieved (detailed laboratory procedures are available [see Data Repository]). Eighteen of the samples contained sufficient quartz, and low enough stable Al, for accelerator mass spectrometric analysis (Table 3).

Data Quality

Measured ratios for most paired blanks agree to within one standard deviation for both ^{10}Be and ^{26}Al (Data Repository Table DR1). Replicate analyses of WS-2 agree within one standard deviation for both ^{26}Al and ^{10}Be (Table 3). Three samples (WS-2, WS-2X, WS-6) for which duplicate Al targets were prepared from the same hydroxide give ratios that are indistinguishable at one standard deviation (Fig. 5). Grain-size specific samples of WS-1 agree somewhat less well, although all are similar within two standard deviations (Table 3). The average $^{26}Al/^{10}Be$ ratio of the data set, considering only samples with latest Pleistocene exposure ages ($n = 11$), is 6.04 ± 0.36 (one standard deviation), consistent with prior estimates (Nishiizumi et al., 1989). Linear-regression analysis of the same data set suggests a ratio of 5.99 ± 0.39 (one standard error) and an intercept not significantly different from zero (<5000 atoms ^{26}Al). The ^{26}Al and ^{10}Be measurements of samples with latest Pleistocene exposure ages are very well correlated ($r^2 = 0.96$).

Production Rates of Cosmogenic Nuclides

To calculate model ages for each sample, we use the production rates derived by Nishiizumi et al. (1989) from glacially eroded surfaces in the Sierra Nevada and the production rates of Stone et al. (1998b) derived from glaciated surfaces in Scotland. We have chosen these rates because they represent, respectively, the highest and lowest estimates currently in the literature. We apply corrections for the latitude, elevation, and sample thickness for each sample by using the scaling functions of Lal (1991) considering only neutrons. The uncertainties in the estimated model ages shown in Table 4 reflect only the analytical uncertainty (one standard deviation).

TABLE 1. RADIOCARBON AGES CALIBRATED BY USING CALIB VERSION 4.1

Site	Sample	Age (^{14}C yr B.P.)	Age (cal. yr B.P.)†
Northern Wisconsin‡	WIS-2022	26,060 ± 800	~32,000§§
Valders Quarry§	CAM-252810	14,500 ± 70	17,370 ± 255
	CAM-291200	14,210 ± 90	17,030 ± 260
	CAM-291210	13,980 ± 190	16,770 ± 320
	CAM-291190	13,150 ± 120	15,810 ± 320
	WIS-2293	12,965 ± 200	15,610 ± 420
Sherwood Quarry#	Beta-119360	13,370 ± 90	16,060 ± 270
Dane County††	WIS-431	13,120 ± 130	15,780 ± 330
Devils Lake‡‡	W-1004	12,880 ± 130	15,510 ± 340
	W-1075	12,520 ± 160	14,770 ± 310
	W-1073	12,260 ± 120	13,920 ± 180
Two Creeks‡‡	Multiple samples	11,850 ± 200	13,815 ± 260

†1σ error, calibrated with CALIB, Version 4.1 (Stulver and Reimer, 1993; and Stulver et al., 1998a, 1998b).
‡Attig et al. (1985).
§Maher and Mickelson (1996), and Maher et al. (1998).
#B.J. Socha (personal commun., 1998).
††Black (1976).
‡‡Maher and Mickelson (1996).
§§Bard et al. (1998).

RESULTS

Measured ^{10}Be and ^{26}Al abundances in our samples vary by over an order of magnitude (Table 3); many samples have far greater than expected in situ–produced cosmogenic nuclide activities. Measured ^{10}Be isotope abundances range from 0.95×10^5 to 9.68×10^5 atoms per gram of quartz, and ^{26}Al abundances range from 0.59×10^6 to 6.08×10^6 atoms per gram of quartz (Table 3). Because all of our samples were collected from outcrops that were covered by glaciers more recently than 30,000 yr ago, continuous surface exposure after deglaciation at this latitude and altitude should yield only $\sim 1 \times 10^5$ atoms of ^{10}Be and $\sim 6 \times 10^5$ atoms of ^{26}Al per gram of quartz. It is clear that many samples record integrated near-surface exposure histories much greater than the expected 15,000–20,000 yr (Table 1).

In general, the lowest and most spatially consistent nuclide abundances were measured in samples collected farthest from the former ice margin (>50 km along a flow line) at the Cactus Rock (1.03–1.09 $\times 10^5$ atoms of ^{10}Be per gram of quartz) and the Waterloo sites (0.95–1.20 $\times 10^5$ atoms of ^{10}Be per gram of quartz) (Fig. 6A). In contrast, samples collected from the outcrop nearest the former ice margin (~10 km), the Baraboo Quartzite near Devil's Lake, have two to five times the expected nuclide abundance (3.05–6.09 $\times 10^5$ atoms of ^{10}Be per gram of quartz) (Fig. 6A). Two samples of the Observatory Hill metarhyolite, collected ~30 km from the late Wisconsin ice margin, have the greatest nuclide abundance (1.69–9.68 $\times 10^5$ atoms of ^{10}Be per gram of quartz), up to eight times the expected abundance. Samples of the nearby Katsma metarhyolite contain only slightly higher than expected activities of ^{10}Be (1.25–2.18 $\times 10^5$ atoms per gram of quartz, Fig. 6A).

When the nuclide-abundance data are considered in terms of surface-exposure ages by using the interpretive model of Lal (1988) and the production rates of cosmogenic nuclides of either Nishiizumi et al. (1989) or Stone et al. (1998b), the results are quite unexpected for many samples (Table 4). Samples from Observatory Hill, 30 km up-ice from the late Wisconsin ice terminus, have exposure ages ranging from 23,000 to nearly 197,000 yr. Samples of Baraboo Quartzite collected near Devils Lake give model ages ranging from 35,000 to 95,000 yr. Samples from Katsma Rock yield more reasonable ages of 17,000–39,000 yr. Only two sites, the Cactus Rock granite and the Waterloo quartzite, have model age distributions (13,500–19,700 yr and 13,600–21,100 yr, respectively) that are consistent with the established radiocarbon chronology considering the current uncertainties both in production rates of cosmogenic nuclides and in radiocarbon ages (Table 4).

The data from Waterloo and Cactus Rock demonstrate the potential precision of cosmogenic age estimates made with multiple samples. Weighted average ^{10}Be and ^{26}Al model age estimates agree within one standard deviation, ~3%, for both sites. Samples collected adjacent to each other, such as WS-10 and WS-11, have average model ages that are indistinguishable at one standard deviation. The ^{10}Be and ^{26}Al abundances at the Cactus Rock and Waterloo site are within two standard deviations, making it unclear whether the younger model ages of the Cactus Rock site are significant. These samples are consistent

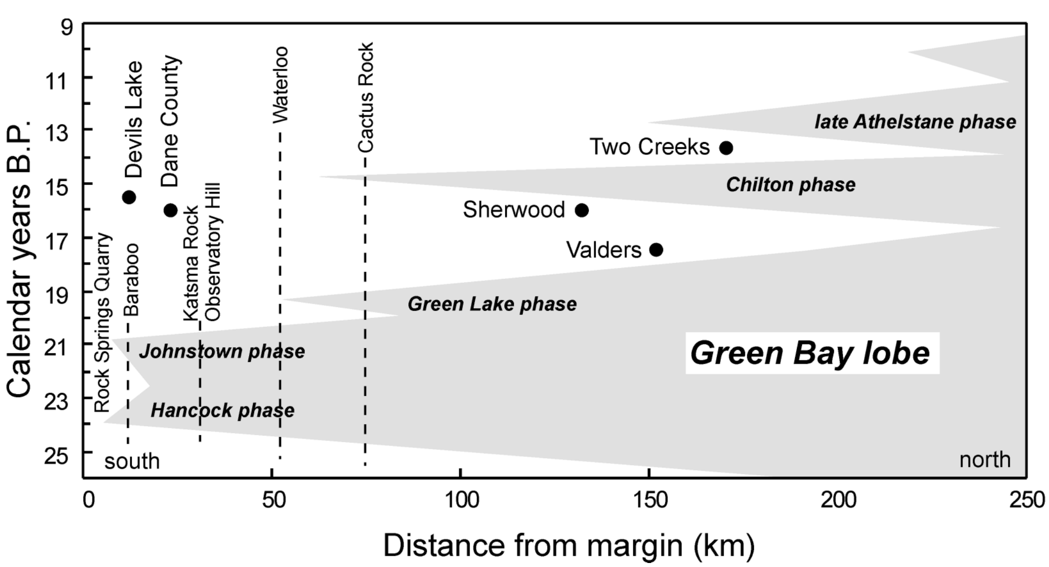

Figure 3. Interpretive time vs. distance diagram for the Green Bay lobe from 26,000 to 9000 cal. yr B.P. (after Clayton et al., 1992). Solid dots are sites of radiocarbon dates (Table 1). Only the major phases during which significant advance occurred are labeled. Cosmogenic isotope sample sites identified with dashed lines.

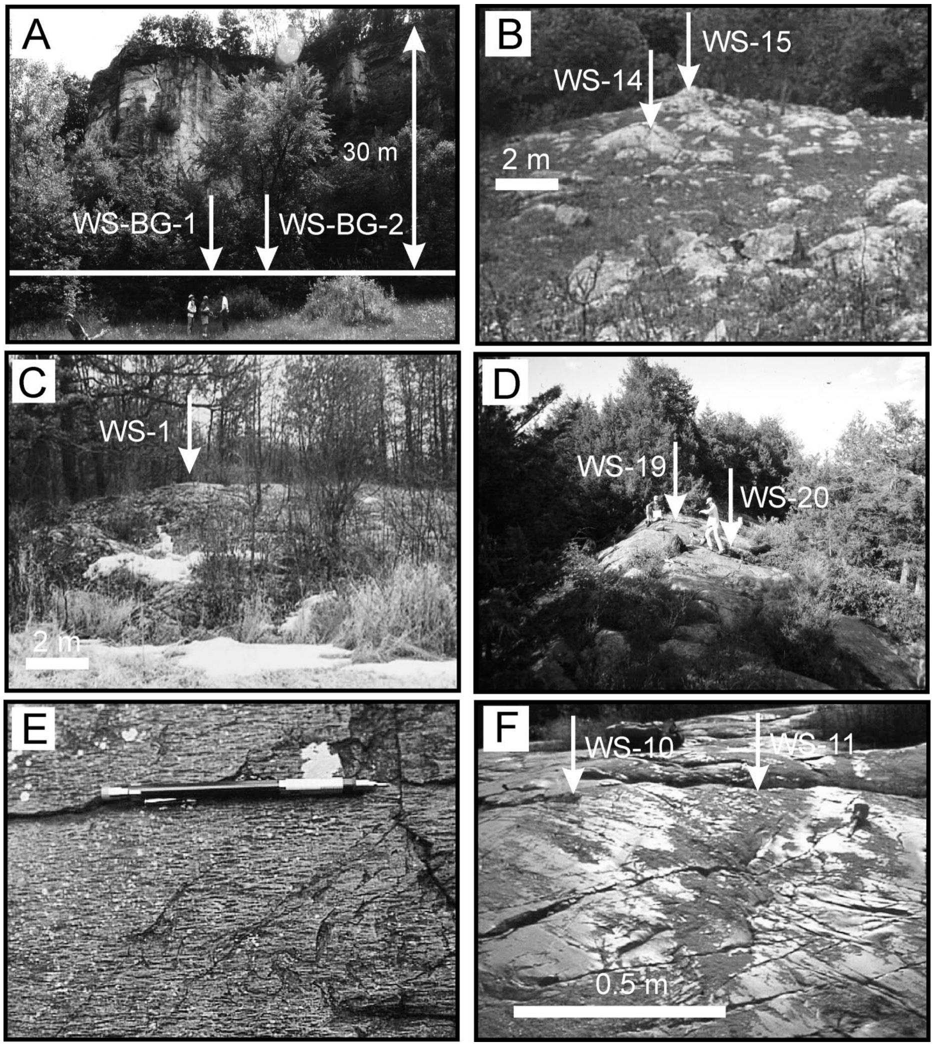

Figure 4. Photographs of sampling sites. Arrows show locations of where individual samples were taken. A—Rock Springs Quarry site (WS-BG-1 and WS-BG-2); B—Baraboo Hills site (WS-14 and WS-15); C—Waterloo site (WS-1); D—Katsma Rock site (WS-19 and WS-20); E—ventifact features on striae at Observatory Hill (WS-7); F—Cactus Rock site (WS-10 and WS-11) (note the glacial polish on this surface).

with the existing deglaciation chronology that suggests the ice retreated from its maximum position sometime before 15,000 cal. yr B.P. (Table 1 and Fig. 3).

We collected two samples from a depth of ~25–30 m at the Rock Springs Quarry (Fig. 1), ~10 km outside of the glacial margin in the Driftless area, assuming that they would serve as field blanks for this area. It is interesting that these samples have low, but detectable cosmogenic-nuclide abundances. Measured $^{10}Be/^{9}Be$ and $^{26}Al/^{27}Al$ ratios of these samples are only 2.2–3.6, and 5.1–18 times greater than the weighted average of pairs of batch blanks (Fig. 5). Low cosmogenic-nuclide abundances such as these in shielded samples could account for the equivalent of 1000–3000 yr of exposure at the surface in our field area. By using the muon versus depth relationship of Granger and Smith (2000), the nuclide activity we measured indicates very low long-term erosion rates of 4–5 m/m.y. (0.004–0.005 mm·yr^{-1}).

DISCUSSION

Considering the data and what is already known about the deglaciation history of Wisconsin, we interpret our results in terms of how nuclide inheritance is influenced by rock properties, by sampling position on the outcrop, and by the spatial distribution of glacial erosion. We suggest that an important control on inheritance is site location in relation to the amount of subglacial erosion.

What Controls Inheritance?

The significant inheritance we detected in nearly half our samples indicates that some bedrock surfaces exposed today were at or near Earth's surface prior to the last advance of the Laurentide Ice Sheet. Very high nuclide concentrations measured in samples from two metarhyolite outcrops (Observatory Hill and Katsma Rock) and a quartzite outcrop (Baraboo) suggest that overrunning by ice removed little material from surfaces that had previously been exposed to cosmic radiation. Conversely, the lack of significant inheritance at Waterloo and Cactus Rock is consistent with either significant glacial erosion, probably >5 m of rock, or significant cover by unconsolidated sediment during the previous interglacial. Because we have no way of knowing drift thickness prior to the last glaciation, we cannot distinguish between these explanations.

Rock properties such as strength and fracture spacing, as well as differences related to the subglacial setting, likely control the amount and distribution of glacial erosion. There are few established relationships between measures of rock strength and rates of glacial erosion; yet there is clearly a difference in glacial erosion rates between different rock types. All of the rocks in this study are quite strong and have widely spaced (>2 m) fractures (Haimson, 1978; Fig. 6B).

As has been found in other studies, sample location on particular outcrops also appears to influence the magnitude of inheritance (Briner and Swanson, 1998). Observatory Hill rises ~60 m above the surrounding landscape and is roughly circular and ~1 km in diameter at its base. A sample (WS-9) taken from a small isolated knob (with local relief of ~3 m, length ~10 m) near the summit of Observatory Hill had the lowest nuclide abundance at this site. Much greater nuclide abundances were measured in glacially polished and striated samples (WS-6 and WS-7) collected within several meters of each other on a broad (~10 m), flat outcrop just below (~5 m) the summit and on the down-ice side of Observatory Hill. Similarly, at Baraboo, the sample with the least inheritance (WS-15) was taken from the top of a small streamlined knob that had the greatest amount of local relief (~3 m, and length of 10 m) compared to the two other Baraboo samples taken within 100 m of it. All of the other samples from Cactus Rock, Bush and Katsma Rocks, and Waterloo were from the tops of small, rounded, glacially striated

TABLE 2. COSMOGENIC-ISOTOPE SAMPLES, SOUTHEASTERN WISCONSIN SITES

Sample	Site location name	UTM northing†	UTM easting†	Elevation‡ (m)	Rock type
WS-1	Waterloo	4785110	341230	247	Quartzite
WS-2	Waterloo	4784940	341350	259	Quartzite
WS-3	Waterloo	4785010	341330	265	Quartzite
WS-4	Waterloo	4785110	341180	244	Quartzite
WS-5	Waterloo	4785100	344550	245	Quartzite
WS-6	Observatory Hill	4841470	311230	338	Metarhyolite
WS-7	Observatory Hill	4841470	311230	338	Metarhyolite
WS-8	Observatory Hill	4841540	311240	338	Metarhyolite
WS-9	Observatory Hill	4841590	311260	338	Metarhyolite
WS-10	Cactus Rock	4912280	359860	265	Granite
WS-11	Cactus Rock	4912280	359870	265	Granite
WS-12	Cactus Rock	4912270	359820	262	Granite
WS-13	Cactus Rock	4912310	359810	256	Granite
WS-14	Baraboo	4815150	291690	408	Quartzite
WS-15	Baraboo	4815150	291640	405	Quartzite
WS-16	Baraboo	4815150	291630	405	Quartzite
WS-17	Bush Rock	4831960	310180	277	Metarhyolite
WS-18	Bush Rock	4822000	310330	279	Metarhyolite
WS-19	Katsma Rock	4831680	309880	277	Metarhyolite
WS-20	Katsma Rock	4831680	309880	275	Quartz vein
WS-BG-1	Rock Springs Quarry	481900	263000	280	Quartzite
WS-BG-2	Rock Springs Quarry	481900	263000	280	Quartzite

†UTM—universal transverse Mercator. Measurements determined by multiple readings of Garmin GPS 75, datum used WGS 84.
‡Determined from topographic maps.

knobs (<200 m in longest dimension) with <20 m of local relief. At these sites we did not see large variations in nuclide abundances over the outcrop, even though we sampled in several places up to several hundred meters apart.

The two sites that have the least inheritance and therefore were subjected to maximum glacial erosion, Waterloo and Cactus Rock, are farthest from the ice margin (>50 km up flow lines). These sites are also nearest to the axis of the former ice lobe. The sites with the most significant inheritance, Baraboo and Observatory Hill, are closer to the former terminus (10 and 30 km, respectively), suggesting that distance from the ice margin and proximity to the axis of the lobe are important influences on the magnitude of inheritance (Fig. 6A).

Geomorphic and sedimentologic evidence suggests that the area glaciated by the Green Bay lobe can be divided into two major landform-sediment zones, one of net glacial deposition (and low erosion), near the former terminus and one of net glacial erosion, farther up-ice and near the axis of the lobe (Whittecar and Mickelson, 1977, 1979; Mickelson et al., 1983; Colgan, 1996; Colgan and Mickelson, 1997). The marginal depositional zone is dominated by hummocky moraines, outwash fans, and ice-contact deposits, but the presence of a few glacially abraded sites like the ones we sampled at the Baraboo and Observatory Hill sites shows that abrasion did occur in this zone. Sample sites with the highest measured cosmogenic abundances are located within this zone of glacial deposition (and probably minimal glacial erosion). In contrast, the zone of glacial erosion up-ice is dominated by drumlins composed of eroded and deformed sediments and by streamlined bedrock hills of Paleozoic rock and a few rounded knobs (showing evidence of both abrasion and plucking) of Precambrian rocks like the ones we sampled. The sites with the lowest measured cosmogenic abundances are located in this zone of glacial erosion.

It is likely that the Green Bay lobe moved over permafrost (Mickelson et al., 1983; Cutler et al., 2000) during the initial advance to its glacial maximum position ~20,000 yr ago. Cutler et al. (2000) hypothesized, on the basis of numerical modeling, that the Green Bay lobe advanced over permafrost and that a frozen-bed zone up to 100 km wide was present for a few thousand years after the lobe reached its maximum position. Subglacial permafrost thawed at its up-ice edge first, and narrowed progressively after the lobe reached its maximum position. Because of this behavior, areas

TABLE 3. COSMOGENIC-ISOTOPE DATA, WISCONSIN GLACIAL SURFACES

Batch	Sample	Elevation (km)	Latitude	^{26}Al/^{10}Be	^{10}Be (SL, >60°)‡ (10^6 atom g^{-1})	^{26}Al (SL, >60°)‡ (10^6 atom g^{-1})	^{10}Be measured (10^6 atom g^{-1})	^{26}Al measured (10^6 atom g^{-1})	Sample-thickness correction
109	WS-1A	0.25	43.0	5.02 ± 0.49	0.105 ± 0.005	0.526 ± 0.045	0.122 ± 0.006	0.612 ± 0.053	1.04
89	WS-1B	0.25	43.0	6.65 ± 0.69	0.091 ± 0.005	0.602 ± 0.051	0.105 ± 0.006	0.700 ± 0.060	1.04
93	WS-1C	0.25	43.0	5.08 ± 0.54	0.115 ± 0.009	0.582 ± 0.044	0.133 ± 0.010	0.677 ± 0.051	1.04
89	WS-2†	0.27	43.0	5.50 ± 0.92	0.092 ± 0.014	0.483 ± 0.038	0.109 ± 0.017	0.598 ± 0.036	1.05
114	WS-2x†§	0.27	43.0	5.57 ± 0.37	0.089 ± 0.004	0.512 ± 0.033	0.105 ± 0.005	0.587 ± 0.029	1.05
89	WS-3	0.26	43.0	6.27 ± 0.53	0.096 ± 0.005	0.603 ± 0.042	0.116 ± 0.006	0.729 ± 0.051	1.01
89	WS-4	0.24	43.0	6.25 ± 0.65	0.081 ± 0.005	0.506 ± 0.040	0.095 ± 0.006	0.594 ± 0.047	1.02
114	WS-6†	0.32	43.8	6.28 ± 0.43	0.776 ± 0.031	5.049 ± 0.265	0.968 ± 0.038	6.080 ± 0.034	1.04
114	WS-7	0.32	43.8	5.99 ± 0.34	0.708 ± 0.021	4.240 ± 0.202	0.895 ± 0.027	5.355 ± 0.255	1.03
114	WS-9	0.32	43.8	6.34 ± 0.55	0.136 ± 0.008	0.863 ± 0.053	0.169 ± 0.010	1.068 ± 0.066	1.05
95	WS-10	0.28	44.3	6.55 ± 0.53	0.089 ± 0.004	0.585 ± 0.038	0.109 ± 0.005	0.714 ± 0.047	1.03
95	WS-11	0.28	44.3	6.02 ± 0.54	0.086 ± 0.005	0.515 ± 0.032	0.104 ± 0.007	0.629 ± 0.039	1.03
95	WS-12	0.28	44.3	5.86 ± 0.48	0.088 ± 0.005	0.514 ± 0.028	0.108 ± 0.007	0.632 ± 0.035	1.02
95	WS-13	0.27	44.3	5.81 ± 0.58	0.084 ± 0.007	0.486 ± 0.030	0.103 ± 0.008	0.599 ± 0.037	1.01
95	WS-14a	0.43	43.5	4.71 ± 0.27	0.438 ± 0.014	2.063 ± 0.099	0.609 ± 0.020	2.865 ± 0.137	1.02
89	WS-15	0.42	43.5	5.62 ± 0.40	0.222 ± 0.009	1.246 ± 0.074	0.305 ± 0.012	1.711 ± 0.101	1.03
89	WS-16	0.42	43.5	5.48 ± 0.30	0.432 ± 0.011	2.364 ± 0.112	0.595 ± 0.016	3.259 ± 0.154	1.03
114	WS-19	0.26	43.5	5.91 ± 0.57	0.183 ± 0.008	1.084 ± 0.094	0.218 ± 0.009	1.287 ± 0.112	1.03
109	WS-20	0.26	43.5	6.46 ± 0.42	0.103 ± 0.004	0.665 ± 0.034	0.125 ± 0.005	0.809 ± 0.041	1.01
95	WS-BG-1	0.28	43.5	4.23 ± 1.77	0.009 ± 0.005	0.039 ± 0.005	0.013 ± 0.004	0.048 ± 0.007	1.00
109	WS-BG-2	0.28	43.5	2.41 ± 0.68	0.018 ± 0.003	0.041 ± 0.017	0.023 ± 0.003	0.055 ± 0.014	1.00

†Two Al targets were measured for each of these samples; the averaged results are used to calculate nuclide abundance. The standard error of the mean is used for uncertainty.
‡Abundances normalized to sea level and high latitude (>60°) by using the formulation of Lal (1991), considering only neutron spallation.
§Indicates fully replicated sample.

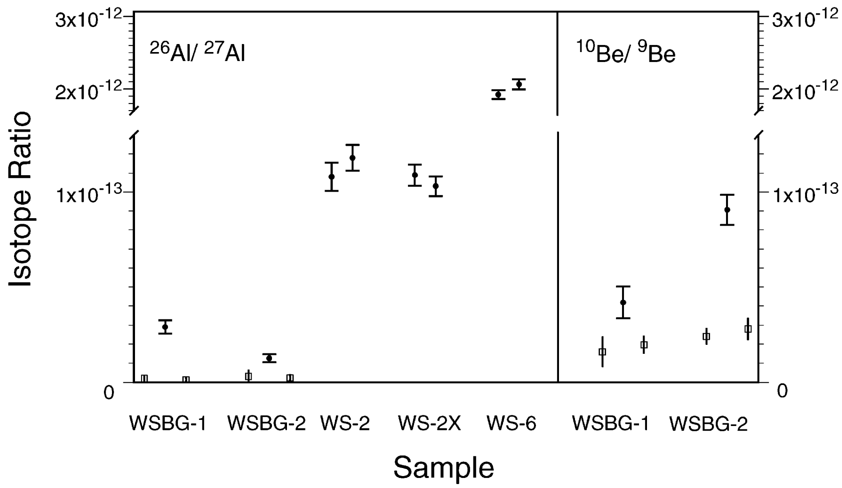

Figure 5. Isotopic ratios measured by AMS (accelerator mass spectrometry). For samples WS-BG-1 and WS-BG-2, batch blanks and samples are plotted along with 1σ uncertainties (bars), demonstrating that sample ratios (circles) are consistently above the ratios measured in blanks (squares). For WS-2, WS-2X, and WS-6, ratios measured on two targets prepared from the same hydroxide, but analyzed on different days, are plotted along with 1σ uncertainties showing that measured ratios reproduce well. Note split scale.

TABLE 4. COSMOGENIC ISOTOPE MODEL AGES, WISCONSIN GLACIAL SURFACES

Sample	Production rates of Stone et al. (1998b)[†]			Production rates of Nishiizumi et al. (1989)[‡]		
	^{10}Be model age (ka)	^{26}Al model age (ka)	Average age[§] (^{26}Al, ^{10}Be)	^{10}Be model age (ka)	^{26}Al model age (ka)	Average age[§] (^{26}Al, ^{10}Be)
Waterloo						
WS-1[#]	21.6 ± 0.9	20.4 ± 1.1	21.1 ± 0.9	16.8 ± 0.7	15.6 ± 0.9	16.3 ± 0.7
WS-2[#]	19.1 ± 0.9	17.9 ± 1.2	18.7 ± 0.9	14.9 ± 0.7	13.7 ± 0.9	14.5 ± 0.7
WS-3	20.6 ± 1.0	21.6 ± 1.5	20.9 ± 1.0	16.0 ± 0.8	16.5 ± 1.2	16.2 ± 0.8
WS-4	17.3 ± 1.2	18.1 ± 1.4	17.6 ± 1.2	13.5 ± 0.9	13.8 ± 1.1	13.6 ± 0.9
Average site age[§]	19.9 ± 0.6	19.4 ± 0.7	19.7 ± 0.6	15.5 ± 0.4	14.8 ± 0.5	15.2 ± 0.4
Observatory Hill						
WS-6	171.7 ± 7.1	197.1 ± 11.4		132.7 ± 5.4	147.5 ± 8.3	
WS-7	156.2 ± 4.9	162.8 ± 8.4		120.8 ± 3.8	122.4 ± 6.2	
WS-9	29.2 ± 1.8	31.1 ± 1.9		22.7 ± 1.4	23.7 ± 1.5	
Cactus Rock						
WS-10	19.1 ± 0.9	21.0 ± 1.4	19.7 ± 0.9	14.9 ± 0.7	16.0 ± 1.1	15.2 ± 0.7
WS-11	18.3 ± 1.2	18.4 ± 1.1	18.4 ± 1.2	14.2 ± 0.9	14.1 ± 0.9	14.2 ± 0.9
WS-12	18.7 ± 1.1	18.4 ± 1.0	18.6 ± 1.1	14.6 ± 0.9	14.1 ± 0.8	14.3 ± 0.9
WS-13	17.9 ± 1.4	17.4 ± 1.1	17.6 ± 1.4	13.9 ± 1.1	13.3 ± 0.8	13.5 ± 1.1
Average site age[§]	18.6 ± 0.6	18.6 ± 0.7	18.6 ± 0.6	14.5 ± 0.5	14.2 ± 0.5	14.4 ± 0.5
Baraboo						
WS-14a	95.3 ± 3.2	75.9 ± 3.8		73.9 ± 2.4	57.7 ± 2.8	
WS-15	47.7 ± 1.9	45.2 ± 2.7		37.1 ± 1.5	34.4 ± 2.1	
WS-16	93.8 ± 2.5	87.5 ± 4.3		72.8 ± 2.0	66.4 ± 3.2	
Katsma Rock						
WS-19	39.3 ± 1.7	39.2 ± 3.5		30.6 ± 1.3	29.9 ± 2.6	
WS-20	22.0 ± 0.9	23.9 ± 1.2		17.1 ± 0.7	18.2 ± 0.9	
Samples run in replicate						
WS-1A	22.4 ± 1.0	18.8 ± 1.6	21.4 ± 1.0	17.5 ± 0.8	14.4 ± 1.3	16.6 ± 0.8
WS-1B	19.3 ± 1.2	21.6 ± 1.9	20.0 ± 1.2	15.1 ± 0.9	16.5 ± 1.4	15.5 ± 0.9
WS-1C	24.5 ± 1.9	20.9 ± 1.6	22.4 ± 1.9	19.1 ± 1.5	15.9 ± 1.2	17.2 ± 1.4
Weighted average[§]	21.6 ± 0.9	20.4 ± 1.1	21.1 ± 0.9	16.8 ± 0.7	15.6 ± 0.9	16.3 ± 0.7
WS-2	19.7 ± 3.1	17.3 ± 1.4	17.7 ± 3.1	15.4 ± 2.4	13.2 ± 1.0	13.6 ± 2.4
WS-2	19.1 ± 0.9	18.3 ± 1.2	18.8 ± 0.9	14.9 ± 0.7	14.0 ± 0.9	14.6 ± 0.7
Weighted average[§]	19.1 ± 0.9	17.9 ± 1.2	18.7 ± 0.9	14.9 ± 0.7	13.7 ± 0.9	14.5 ± 0.7

[†] ^{10}Be production rate = 4.70 atoms of ^{10}Be per gram of quartz per year; ^{26}Al production rate = 28.2 atoms of ^{26}Al per gram of quartz per year.

[‡] ^{10}Be production rate = 6.03 atoms of ^{10}Be per gram of quartz per year; ^{26}Al production rate = 36.8 atoms of ^{26}Al per gram of quartz per year.

[§] Average (weighted) calculated considering uncertainty of each measurement.

[#] Weighted average of replicates used for calculation of average site age.

farthest up-ice would have experienced a wet, erosive, abrasive sliding bed with significant regelation longer than would areas nearest to the terminus. Our cosmogenic-nuclide data are consistent with these model results.

It is also possible that the duration of glacial cover and velocity of ice were as important as the basal thermal conditions. Areas nearest to the margin would have been covered with thinner, more slowly moving ice for much less time than would sites farther up-ice near the lobe axis. The Baraboo and Observatory Hill sites were probably covered with <500 m of ice for <5000 yr. The Waterloo and Cactus Rock sites could have been covered by ice >500 m thick for up to 10,000 yr (Clark, 1992; Colgan, 1999). Ice-flow velocities near the lobe axis may have been as high as several hundred meters per year (Colgan, 1999), but near the margin, they could have been very low or even zero (stagnant ice). It is likely that all of the factors mentioned would influence the magnitude of local glacial erosion (both abrasion and plucking) in any site covered by an ice lobe. Future studies will be necessary to determine the relative importance of these factors in erosion by continental ice lobes.

Erosion Rates

We cannot make finite estimates of either rates of glacial erosion or integrated rates of denudation because we cannot know the nuclide abundance of these outcrops before they were overrun by late Wisconsin ice. However, by using several simplifying assumptions we can estimate crudely limiting glacial erosion rates. For example, we can use the measured nuclide concentrations and the steady-state erosion models of Lal (1991) to calculate maximum limiting integrated erosion rates because such models assume constant surface exposure. Such limiting rates are quite low for some of our samples, ~0.004 mm·yr^{-1} for WS-6 and WS-7, the metarhyolite of Observatory Hill. This calculation assumes that the total time spent under ice is inconsequential compared to the exposure time.

We can also estimate minimum amounts of mass loss from the outcrops with inheritance and thus define minimum rates of glacial erosion (Davis et al., 1999). If we assume that the sample (WS-6) with the highest nuclide concentration was not eroded at all by the latest advance of Laurentide ice (provides N_0) and that the average abundance of ^{10}Be and ^{26}Al in Cactus Rock and Waterloo samples represents nuclide accumulation since deglaciation (N_{CW}, for Cactus Rock and Waterloo), we can calculate the amount of erosion (represented by z) that affected other sample sites by using the measured nuclide abundance (N_{CW}):

$$z = -\frac{\Lambda}{\rho}\ln\left(\frac{N_m - N_{CW}}{N_0}\right),$$

where z = depth of rock eroded (in centimeters), Λ = neutron-absorption coefficient (165 g·cm^{-2}), ρ = rock density (2.7 g·cm^{-3}), N_m = measured nuclide abundance (atoms per gram of quartz), N_0 = nuclide abundance at surface prior to last glacial erosion episode (atoms per gram of quartz), and N_{CW} = average measured nuclide abundance for Cactus Rock and Waterloo (atoms per gram of quartz).

By using this equation we calculate that sample sites such as WS-7 may have lost <10 cm to erosion, whereas sample site WS-20 may have been eroded ~2.5 m. Estimating rates of glacial erosion requires that we know the duration of ice cover. All of the sites were covered by both late Wisconsin ice and earlier advances of the Laurentide Ice Sheet. By using the limiting radiocarbon dates (Table 1), we surmise that ice covered outcrops for no more than ~10,000 yr (26,000 yr B.P. to 17,400 cal. yr B.P. for Valders); for all samples, but particularly the Baraboo samples, the duration of cover may have been much less.

By using the upper limit for the duration of ice cover and the minimum depths of erosion calculated as we have described, we estimate lower limits for glacial erosion rates of >0.01 to >0.25 mm·yr^{-1} (Table 5). Five of our estimates, those >0.1 mm·yr^{-1}, are consistent with rates measured by Davis et al. (1999) in Baffin Island (0.1 to >0.16 mm·yr^{-1}) and by Briner and Swanson (1998) in the Puget Lowland on a streamlined feature (0.09–0.35 mm·yr^{-1} for the stoss side, 0.63–0.72 mm·yr^{-1} for summit surfaces, and 0.13–0.28 mm·yr^{-1} for the lee side above the quarried zone). Our other three estimates, for samples WS-7, WS-14a, and WS-16 (>0.01, >0.04, >0.04 mm·yr^{-1}, respectively), are much lower than glacial erosion rates measured elsewhere. This difference likely reflects the exceptional competence of the quartzite and metarhyolite outcrops and the fact that the glacier bed was sliding for only a small part of the time the outcrops were ice covered. If, for instance, the glacier bed was sliding for only the last 1000 yr of ice cover, then rates of erosion, averaged over this short interval would be an order of magnitude higher.

With the exception of sample WS-14a, all samples have ^{26}Al/^{10}Be ratios near 6.0 and are thus consistent, considering analytical uncertainty, with continuous or nearly continuous surface exposure. Sample WS-14a, with ^{26}Al/^{10}Be = 4.71 ± 0.27, appears to have been either exposed, reburied, and reexposed or dosed under cover and recently exhumed. It is also possible for this one sample that stable Al was not fully recovered although we think this possibility unlikely because Be recovery was 99.8%. By using the methodology of

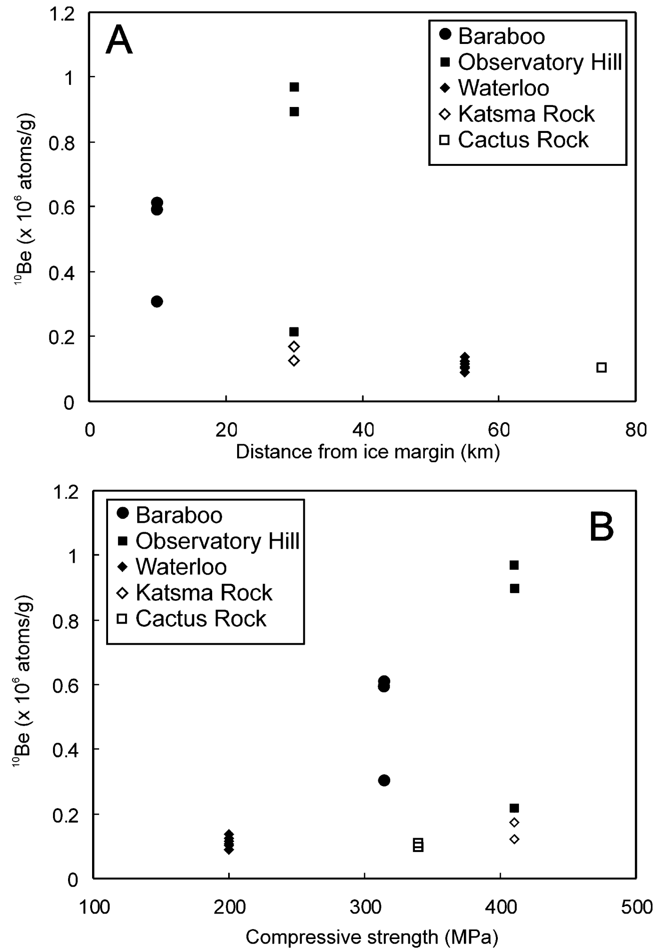

Figure 6. (A) Plot of ^{10}Be concentration vs. distance from the late Pleistocene ice margin. Sites closer to the margin have higher and more variable ^{10}Be concentrations. Sites >55 km from the ice margin have low concentrations and less variability in ^{10}Be concentration. (B) Plot showing ^{10}Be concentration vs. compressive strength of the rocks sampled (Haimson, 1978). Samples with high compressive strength have higher concentrations of ^{10}Be; weaker samples have lower concentrations of ^{10}Be. A nearly identical relationship exists between ^{26}Al concentration and compressive strength and between ^{26}Al concentration and distance from margin (not shown).

TABLE 5. MINIMUM EROSION RATES FOR WISCONSIN OUTCROPS

Sample	^{10}Be abundance, exposure corrected* (10^5 atoms per gram of rock)	Erosion† (m)	Minimum erosion rate§ (mm·yr^{-1})
WS-7	7.87	0.13	0.01
WS-9	0.61	1.69	0.17
WS-14a	5.01	0.40	0.04
WS-15	1.97	0.97	0.10
WS-16	4.87	0.42	0.04
WS-19	1.10	1.33	0.13
WS-20	0.17	2.47	0.25

*Calculated by subtracting average ^{10}Be abundance in samples from Cactus Rock and Waterloo.
†Calculated by using equation 1 and measurement of ^{10}Be abundance.
§Calculated by assuming 10,000 yr of burial and erosion by the Laurentide Ice Sheet.

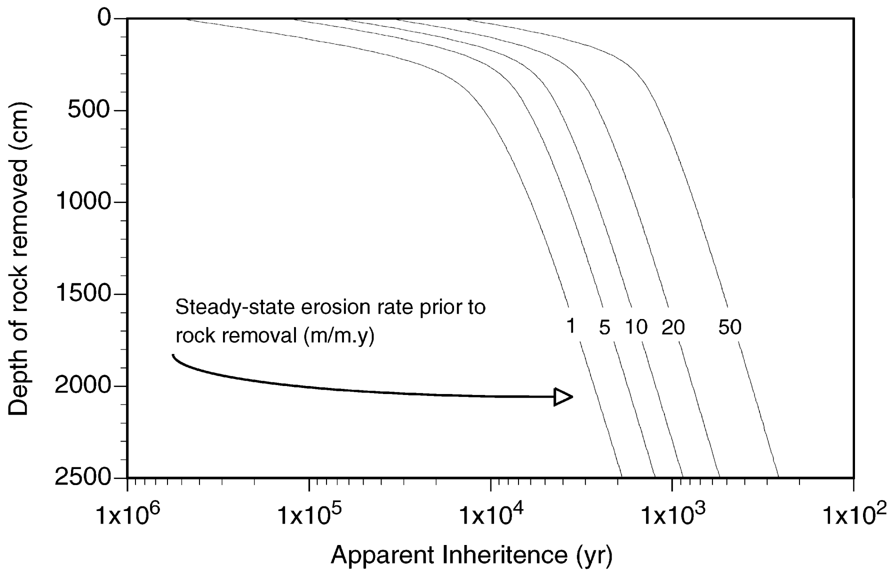

Figure 7. Apparent inheritance (in ^{10}Be years) resulting from instantaneous removal (glacial erosion) of rock (characteristic attenuation length $\Lambda = 2.7$) from a landscape that was in erosional steady state. Each line represents a different steady-state erosion rate (in m/m.y.). Increase in line slope reflects change from predominately neutron to predominately muon production at a depth of ~3 m. Calculations were made for lat 43.5°N and 250 m above sea level by using a neutron-attenuation factor of 165 g·cm^{-2}, the equations of Granger and Smith (2000), and the surface production-rate estimate of Nishiizumi et al. (1989).

Bierman et al. (1999), we calculate that if this sample had only been reexposed just before sampling, it would reveal a minimum total history of 525,000 yr including 126,000 yr of exposure followed by 399,000 yr of burial. Alternatively, if the sample had been exposed as long as the outcrops at Waterloo and Cactus Rock, we can subtract from measured nuclide abundances those nuclides produced during the latest period of irradiation (15,000 yr). We would then calculate a total exposure history of 490,000 yr including 95,000 yr of surface exposure and 395,000 yr of burial. It is virtually certain that this sample has been exposed to multiple periods of ice-sheet shielding, but has been subject to minimal erosion as ice advanced and retreated during the middle and late Pleistocene. Considering the ^{26}Al/^{10}Be of WS-14, which is 4.71 ± 0.27, suggests an extremely low long-term erosion rate at this site of 0.9 m/m.y. (Bierman et al., 1999).

Implications for Production-Rate Calibrations

Calculation of exposure ages requires estimates of production rates of cosmogenic nuclides. One method for estimating production rates involves measuring nuclide abundances in samples collected from surfaces for which the exposure age is known by other means. Several production-rate calibrations have been based on striated glaciated surfaces (Nishiizumi et al., 1989; Clark et al., 1995; Bierman et al., 1996; Stone et al., 1998b). These sites are chosen because the assumption of ''no postglacial erosion'' can be well established owing to the preservation of glacial striae and other abrasion features. In all cases, inheritance in these samples was assumed to be inconsequential. However, our data raise the possibility that some nuclides in samples used to estimate production rates were produced during prior periods of irradiation. If such inheritance did occur, production rates would be overestimated. Such overestimation could be tested by comparing production rates estimated in young volcanic rocks with those estimated from glaciated sites. Unfortunately, proving the ''no postdepositional erosion'' assumption in volcanic rocks may be as difficult as proving the no ''prior inheritance'' assumption in glaciated surfaces. Additionally, glacially striated outcrops may be preserved because they are very resistant to chemical and mechanical erosion, yet this quality makes it even more likely that they will contain some inherited cosmogenic nuclides. Thus, some of the best outcrops for proving the ''no postglacial erosion'' assumption may be the most likely to have been eroded the least during glaciation. This dilemma suggests caution in deriving production rates of cosmogenic nuclides from glacially striated outcrops.

Implications of the Unglaciated Quarry Samples

The quarry samples demonstrate that even deeply shielded rocks contain measurable ^{10}Be and ^{26}Al. The nuclides we detected must have been produced by muons, which penetrate far deeper than neutrons (Lal, 1988; Brown et al., 1995; Stone et al., 1998a; Granger and Smith, 2000). In order to limit ^{10}Be inheritance to 1000 yr or less, almost 7 m of rock must be removed between periods of surface exposure if the surface is eroding, on average, relatively quickly (50 m/m.y.). For surfaces eroding more slowly (5 m/m.y.), >25 m of rock must be removed for inheritance to be <1000 yr (Fig. 7). These calculations support our observations of significant inheritance and suggest that long-term subaerial erosion rates just outside the ice margin in Wisconsin are relatively low. They also suggest that even the Cactus Rock and Waterloo samples probably carry some inherited nuclides.

Implications for Dating Glacial Terrains

These data suggest that cosmogenic model ages may, in some instances, overestimate deglaciation ages because of inheritance of nuclides from prior periods of near-surface exposure. This finding is similar to that of Bierman et al. (1999), who collected and measured samples from the once-glaciated, high-elevation plateaus of Baffin Island and found that they had significant inheritance. Yet, in other areas, including the late Wisconsin Laurentide ice-sheet margin in New Jersey, Pleistocene moraines and glaciated bedrock in the Sierra Nevada in California, and a fiord bottom site in the Canadian Arctic, inheritance has been shown to be minimal (Bierman, 1993; Clark et al., 1995; Nishiizumi et al., 1989; Davis et al., 1999). These observations can best be explained by variation in glacial erosion due to basal ice conditions, velocity, and duration of ice cover.

Minimum inheritance in bedrock (and maximum glacial erosion) appears to occur in deep alpine glacial valleys and fjords where ice flowed rapidly and was not frozen to the bed (Davis et al., 1999). Conversely, thinly covered highlands in the arctic (Bierman et al., 1999) and low-relief margins such as parts of the southern Laurentide Ice Sheet were in general less deeply eroded by ice. In the Green

Bay lobe of Wisconsin, glacial erosion must have focused on the tops of isolated knobs, and areas farthest behind the margin and near the axis of the lobe where there was a sliding bed for a longer period of time.

Samples exposed at mountain elevations should have a lower percentage of inherited nuclides because neutron-induced production rises more quickly with elevation than does muon-induced production. Steep, high-altitude, temperate alpine glacial valleys are thus less likely to be affected by significant inheritance than cold, low-altitude, low-relief areas overrun by continental ice. Paradoxically, the very strong rocks that best preserve striations and ensure that the assumption of ''no postglacial erosion'' is valid appear to be the rocks most likely to violate the assumption of no-inheritance precisely because they are so resistant to erosion.

CONCLUSIONS

Our data suggest that caution is appropriate when cosmogenic-nuclide abundances in bedrock samples collected near former ice margins are interpreted as deglaciation ages, particularly if sampled outcrops are very resistant rocks such as quartzite and metarhyolite or if there is any evidence of a frozen glacial bed that may have minimized subglacial erosion. By using a suite of samples collected from very competent, well-preserved outcrops of Precambrian rocks in Wisconsin, we demonstrate the following:

1. At least 7 of 16 samples retain a significant amount of cosmogenic nuclides produced during a period of near-surface exposure prior to the period that we intended to date. Thus, model ages calculated for these samples greatly overestimate the time since late Wisconsin ice-sheet retreat.

2. Outcrops of strong rock near the former ice margin in zones that were subjected to little subglacial erosion (and outcrops in areas containing numerous depositional features such as moraines and hummocky topography) have the highest nuclide concentrations and, therefore, the most inheritance of nuclides from prior surface exposure. Conversely, outcrops of weaker rocks and outcrops farther from the ice margin (>50 km) in areas of high erosion (dominated by drumlins and streamlined bedrock knobs) appear to retain fewer nuclides from periods of prior exposure.

3. Deeply penetrating muons produce ^{10}Be and ^{26}Al at depth. In slowly eroding continental terrains, even 5–10 m of glacial erosion may leave significant nuclides produced during prior interglacial intervals. These add >1000 yr of inheritance to model exposure ages and inflate—to an unknown degree—any production-rate estimates derived from glacial terrains.

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