quiescent emission was the cumulative result of many ongoing events of class C or even lesser energy, but this issue of whether coronal heating is entirely the result of flare-like activity or has any steady component is an open issue even for the sun (*20*)

During solar maximum, the number of M flares per month is ~50, that is, ~1.7 per day. During this observation, Proxima Centauri was at least as active as the sun at maximum. It is important to put this in perspective for such a faint dwarf M star: The bolometric luminosity (mainly in the optical and infrared) of Proxima Centauri is only 6.7×10^{30} ergs s^{-1} (*21*), less than 1/500 that of the sun. The total x-ray luminosity at all wavelengths for the most energetic events on Proxima Centauri observed by ASCA and ROSAT is $L_x \sim 3 \times 10^{27}$ to 6×10^{27} ergs s^{-1}. This amounts to an instantaneous perturbation on the order of 0.1% of the total thermonuclear power of the star.

Observation of ordinary M-class flare events on Proxima Centauri indicates that the assumption that flares on the sun and on other stars are scaled versions of the same process is fundamentally sound. It remains a challenge to understand both how flares can, in an absolute sense, exceed those on the sun by up to four orders of magnitude on such extremely active stars as T Tauris (*22*) and RS CVn systems (*23*), and in a relative sense on such faint dwarf M stars as Proxima Centauri.

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17. The formal fits for the flare are $kT_1 = 0.63 \pm 0.01$ keV (*k* is Boltzmann's constant) with an emission measure $EM_1 = 5 \times 10^{49}$ cm^{-3} and $kT_2 = 3.83 \pm 0.67$ keV with $EM_2 = 3 \times 10^{49}$ cm^{-3}. These fits were calculated assuming a column density of hydrogen N_H conservatively fixed at 3×10^{19} cm^{-2}, resulting in a reduced $\chi^2 = 2.3$. For the quiescent intervals, $kT_1 = 0.53$ keV with $EM_1 = 10^{49}$ cm^{-3} and $kT_2 > 4$ keV with $EM_2 = 9 \times 10^{48}$ cm^{-3} (reduced $\chi^2 = 2.9$). In both cases, fitting was hindered by low count rates above 6 keV, and in particular, we do not consider the quiescent kT_2 fit to be credible. These temperatures are marginally statistically hotter than the Einstein Observatory determinations in (*13*).
18. A continuous distribution of temperature for the flaring plasma is likely to be more realistic than the two isolated temperatures T_1 and T_2. Moreover, the temperature of the DEM (differential emission measure) maximum and the highest temperature of the DEM are similar to T_1 and T_2 of the flare.
19. The average count rates for the flare intervals are $SIS_a = 0.331$, $SIS_b = 0.258$, $GIS_a = 0.136$, and $GIS_b = 0.168$ counts s^{-1}. The absolute values are quite dependent on the choice of flare interval; it is the ratio of 1.94 that is important. For the quiescent intervals, they are $SIS_a = 0.071$, $SIS_b = 0.063$, $GIS_a = 0.048$, and $GIS_b = 0.050$ counts s^{-1}.
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Beryllium-10 Dating of the Duration and Retreat of the Last Pinedale Glacial Sequence

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Accurate terrestrial glacial chronologies are needed for comparison with the marine record to establish the dynamics of global climate change during transitions from glacial to interglacial regimes. Cosmogenic beryllium-10 measurements in the Wind River Range indicate that the last glacial maximum (marine oxygen isotope stage 2) was achieved there by 21,700 ± 700 beryllium-10 years and lasted 5900 years. Ages of a sequence of recessional moraines and striated bedrock surfaces show that the initial deglaciation was rapid and that the entire glacial system retreated 33 kilometers to the cirque basin by 12,100 ± 500 beryllium-10 years.

Alpine glacial systems are more responsive than large continental ice sheets to changes in temperature and precipitation. Chronologies of alpine morainal sequences therefore offer the possibility of providing records of Pleistocene climate variations that have higher resolution than do global ice volume records (for example, marine isotopic data). Establishing tight time constraints on the age of morainal deposition has been difficult. Determining how long a glacier has occupied its terminal position (that is, the duration of the glacial maximum) has been even harder. Here, we use measurements of cosmogenic ^{10}Be in quartz exposed in boulders on alpine moraines to reconstruct the complete Pinedale glacial history of the Fremont Lake basin in Wyoming.

During the last major glaciation, a large ice cap on the Wind River Range fed outlet glaciers in the Fremont Lake basin located on the western piedmont of the Wind River Mountains in west-central Wyoming (Fig. 1). The Fremont Lake basin is the type locality for the Pinedale glaciation (*1*, *2*), corresponding to the late Wisconsinan–age (marine oxygen isotope stage 2) glaciation in the Rocky Mountains. The type-Pinedale glacial deposits have been extensively studied, and other deposits throughout North America and South America have been correlated to the type-Pinedale moraines. A precise chronology for the advance and retreat of Pinedale glaciers would provide the basis for addressing the synchronicity of glacial onset and termination in the Northern and Southern hemispheres.

There are only a few dates that tightly constrain the timing or duration of the last Pinedale glacial maximum (LPGM). A minimum age of the LPGM at the type locality is limited by two radiocarbon dates of 6100 ± 100 and 9300 ± 80 ^{14}C years before present (B.P.) [6800 ± 100 and 10,000 ± 50 years, calibrated radiocarbon ages (*3*)] from the bottom of a bog (*4*). Approximately 33.8 km north-northwest of Fremont Lake, a vegetational change asso-

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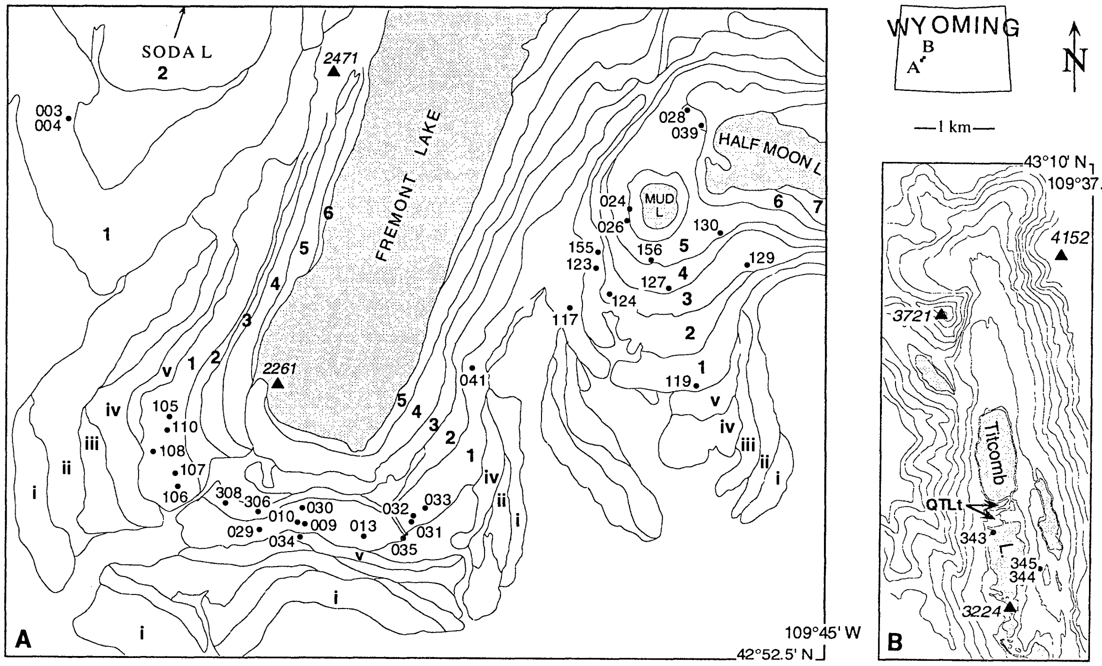

Fig. 1. Geologic map of the Fremont Lake basin (**A**) (2) and topographical map of the Titcomb Lakes basin (**B**) [contour interval, 61 m (200 feet)], showing the locations of the boulder and bedrock sample sites; 1 through 7, moraines of the Pinedale glaciation (QPt); i through v, moraines of the Bull Lake glaciation; (moraines are numbered in order of deposition, with 1 and i being the terminal moraines); 003, sample identification number (Table 1). Triangles provide spot elevations (in meters). Ice flow direction is southward on both maps. Soda Lake is less than 1 km north of the Fremont Lake basin map. Sample WY-93-346 is located 4 km south of the Titcomb Lakes basin map (B).

ciated with the Pinedale glaciation is bracketed between 20,800 and 10,200 ^{14}C years B.P. (23,800 and 11,200 years, calibrated) (*5*). Elsewhere in the Rocky Mountains, the timing and duration of the Pinedale glaciation appears to have been variable. Glacial advances began before 40,000 ^{14}C years B.P. (*6–8*) and continued to about 18,000 years (calibrated), with evidence for several advances in some locations. Some of the differences among glacial histories may reflect inconsistencies in dating (*9, 10*). It is generally believed (*8*) that the last time the Rocky Mountain Pinedale-age glaciers were at their maximum extent was between 20,000 and 15,000 ^{14}C years B.P. (23,000 and 17,500 years, calibrated).

The in situ cosmogenic method provides a means to date morainal materials directly (*11*). We measured concentrations of the radioactive nuclide ^{10}Be (half-life $T_{1/2} = 1.5 \times 10^6$ years) that is produced in quartz at Earth's surface by secondary cosmic-ray–induced spallation of oxygen and, to a lesser extent, silicon. The concentration of ^{10}Be (N_{10}) in a sample increases with exposure time (T). Under the conditions of little erosion and no predepositional exposure, we can solve for exposure age (in years):

$$T = -\frac{1}{\lambda_{10}} \ln\left[1 - \frac{N_{10}\lambda_{10}S\rho}{P_{10}\Lambda(1 - e^{-S\rho/\Lambda})}\right] \quad (1)$$

where P_{10} is the local production rate of ^{10}Be in quartz (calculated for the altitude and latitude from which the sample was collected), λ_{10} is the decay constant ($\lambda = \ln 2/T_{1/2}$) that for ^{10}Be is 0.46×10^{-6} year^{-1}, ρ is the average density of the sample (2.65 g cm^{-3} for the quartz-rich samples we used), Λ is the thickness of rock that reduces the cosmic-ray flux and hence the cosmogenic production rate to $1/e$ of its value at the surface ($\sim$150 g cm^{-2}), and S is the thickness (in centimeters) of a sample taken from the surface of a boulder.

With few exceptions, we measured ^{10}Be in samples collected from the center of flat horizontal (within 15°) surfaces on boulders resting >1.5 m above the crests of moraines or bedrock knobs that protrude >2 m above ground level. None of the 44 boulders sampled in the Fremont Lake basin experienced significant shielding from cosmic radiation by snow, ice, or sediment, and so no shielding corrections were necessary. Similarly, no corrections were made for the effects of erosion, although correcting for an erosion rate of 1.5 μm year^{-1} [an upper limit for what is likely for the Pinedale samples (*12, 13*)] would increase all the ages by $\sim$3%.

The 3% uncertainty reported for individual ^{10}Be dates is the total analytical error in our ^{10}Be measurements (*14, 15*). Reproducibility for identical samples is even better (*16*). We determined mean ages for moraine and bedrock surfaces from the average of all samples. Uncertainties for moraine ages are the sum in quadrature of two components: (i) the uncertainty of the mean, calculated from either the internal error (the variance of individual measurements around their mean) or the analytical error, with the larger of the two divided by the square root of the number of measurements; and (ii) a 4% systematic uncertainty that includes uncertainties in the standard, carrier, and so forth (Table 1).

The correspondence between ^{10}Be years and calendar years, and thus the comparability of these results with results obtained by other dating methods, depends on knowledge of the local production rate of ^{10}Be during the past 20,000 years. The uncertainty in production rates for a ^{10}Be exposure age of 20,000 years at Pinedale is $\sim$15% ($\sim$3000 years) (*17*).

The Pinedale-age terminal moraines of the three ice lobes—Soda Lake lobe (QPt1-SL), Fremont Lake lobe (QPt1-FL), and Half Moon Lake lobe (QPt1-HL)—are assumed to correspond to the LPGM. Fifteen boulders (Table 1) from the QPt1-FL moraine provide a mean age of 18,200 ± 900 ^{10}Be years (*18*) (Fig. 2). The low ^{10}Be concentration in sample WY-91-030 is difficult to explain; the boulder was not resampled or reanalyzed. The range of ages obtained on the other 14 boulders from QPt1-FL is 5900 ± 800 years (*19*). This range is larger than expected from estimates of measurement uncertainty (see below), and we conclude that it represents the time interval during which boulders were delivered to the moraine. We rule out differential erodibility as a factor that would affect Pinedale ages because the presence of glacial polish on some surfaces suggests they have not eroded even 1 or 2 mm in $\sim$20,000 years. The limit on erosion rate imposed by the oldest date for a boulder (*20*) (WY-91-034, Table 1) places the maximum erosion rate at considerably less than 4 μm year^{-1}. A rock erosion rate of 13 μm year^{-1} would be required to reduce a Pinedale exposure age by 5900 years, the range of the ^{10}Be dates on QPt1-FL. The tight age grouping of boulders from recessional moraines (see below) and the

Table 1. Sample data. The elevation given is elevation above sea level; the thickness given is the thickness of the sample. The ^{10}Be concentration given is the measured concentration (not corrected for sample thickness, elevation, or latitude). The sample age given in thousands of ^{10}Be years, has been corrected for sample thickness but not for erosion [numbers in parentheses represent 1σ uncertainties, in 10^3 years, estimated to be $\pm 3\%$ (*14*)]. Bedrock surfaces are indicated by "-b" after the sample number. Landform age is the mean of all sample ages. The uncertainty for the landform ages includes a 4% systematic uncertainty. When comparing landform ages measured in this work internally (with other measurements within this work), landform age uncertainties should be reduced by subtracting in quadrature the 4% systematic error. Ranges, not means, of the ages of terminal moraines are reported because of the long time (~5900 ± 800 years) required to build them (*19*). The ranges are minimum durations: We cannot be sure that we have sampled either the first or the last boulder deposited on the moraine. However, the end of the LPGM is also constrained by the age of the oldest recessional moraine; best estimate = 15.8×10^3 years ago. The approximate north latitudes for the sample sites are as follows: QBt5-FL, 42.89°; QPt1-FL, 42.90°; QPt1-HL, 42.91°; QPt1-SL, 42.93°; QPt2 to QPt5, 42.93°; HL bedrock, 42.94°; Seneca Lake, 43.05°; and >QTLt, 43.11°.

Sample	Elevation (km)	Thickness (cm)	^{10}Be (10^5 atoms/g)	Age (10^3 years)
		Terminal moraines		
QPt1-SL ≤18.0 to ≥21.7 (2.0)				
91-003	2.276	14	5.18	18.1 (0.5)
91-003	2.276	14	5.13	18.0 (0.5)
91-004	2.279	5	6.71	21.7 (0.7)
QPt1-FL =15.8 to ≥21.7 (0.9)				
91-035	2.262	2	6.25	19.9 (0.6)
91-035	2.262	2	6.23	19.9 (0.6)
92-108-1	2.274	2	6.76	21.4 (0.6)
91-032	2.311	5	6.85	21.7 (0.7)
92-105-2	2.290	3	5.86	18.5 (0.6)
92-110-1	2.287	4	5.89	18.8 (0.6)
92-107-1	2.271	2	5.00	15.8 (0.5)
92-106-2	2.271	2	6.02	19.1 (0.6)
91-029	2.299	5	5.48	17.5 (0.5)
91-009	2.310	6	5.50	17.6 (0.5)
91-041	2.360	5	5.40	16.6 (0.5)
91-013	2.302	6	5.94	19.1 (0.6)
91-031	2.311	5	6.15	19.5 (0.6)
91-033	2.302	5	5.30	16.9 (0.5)
91-010	2.302	4	5.07	16.0 (0.5)
91-030	2.305	5	4.54	14.4 (0.4)
QPt1-HL ≤19.2 to ≥19.5 (0.8)				
92-117	2.357	3	6.45	19.5 (0.6)
92-119	2.319	5	6.09	19.2 (0.6)
		Recessional moraines		
QPt2-FL 26.0 (1.4)				
93-308	2.264	5	9.21	30.2 (0.9)
93-306	2.268	3	6.81	21.8 (0.7)
QPt2-HL 16.0 (0.6)				
92-123	2.369	3	5.34	16.0 (0.4)
92-155	2.375	2.5	5.35	15.9 (0.5)
QPt3-HL 16.3 (0.6)				
92-129	2.390	5	5.61	16.9 (0.5)
92-124	2.337	5	5.03	15.7 (0.5)
QPt4-FL 15.5 (0.6)				
91-020	2.352	5	5.03	15.5 (0.5)
QPt4-HL 15.6 (0.6)				
92-127	2.335	4	5.04	15.6 (0.5)
QPt5-HL 16.0 (0.7)				
92-130	2.341	5	5.41	16.8 (0.5)
91-024	2.323	10	5.02	16.5 (0.5)
92-156	2.341	5	5.28	16.4 (0.5)
91-026	2.342	4	4.68	14.4 (0.4)
		Other retreat positions		
>HL5 13.8 (0.6)				
91-039-b	2.352	2	5.09	15.3 (0.5)
91-028-b	2.378	5	4.08	12.4 (0.4)
Seneca Lake 15.9 (0.6)				
93-346-b	3.146	5	8.42	15.9 (0.5)
>QTLt 12.1 (0.6)				
93-345-b	3.247	4	7.28	12.8 (0.4)
93-344	3.247	2.5	6.52	11.3 (0.3)
>QTLt 13.3 (0.6)				
93-343-b	3.226	4	7.45	13.3 (0.4)
		Pre-Pinedale moraine		
QBLt5-FL 137.8 (6)				
91-034	2.299	6	41.6	138 (4)

relation between boulder age and position on the moraine (see Fig. 2, inset) imply that exposure of the boulders before deposition onto the moraines was not significant. In contrast with (*21*), evidence at Fremont Lake (*22*) suggests that broad crested moraines such as the Pinedale terminal have eroded by less than 1 m in the past 20,000 years, which would not be enough to affect the age of any of the sampled boulders (all standing more than 1.5 m above the surfaces of the moraines).

The QPt1-FL moraine is voluminous, and its construction presumably would have required considerable time (*23*). The Fremont Lake lobe evidently advanced to the LPGM position by 21,700 ± 700 ^{10}Be years (WY-92-108 and WY-91-032) and remained there until 15,800 ± 500 ^{10}Be years (WY-92-107 and WY-91-010), but not much later (Table 1). It cannot be determined if the Fremont Lake lobe actually remained at its terminal position for the entire 5900 years; it may have first occupied the LPGM position 21,700 ± 700 ^{10}Be years ago and oscillated near the LPGM position before retreating permanently after 15,800 ± 500 ^{10}Be years (*7*, *10*).

The exposure age of a boulder (WY-91-034) situated on a moraine ridge immediately in front of the Pinedale terminal moraine (Fig. 1 and Table 1) is 137,800 ± 4100 ^{10}Be years if we assume that there has been no erosion [179,200 ± 5400 ^{10}Be years if adjusted for an estimated upper-limit erosion rate of 1.5 $\mu m\ year^{-1}$ (*12*)]. These ages are consistent with the deposition of the moraine during the Bull Lake glaciation as originally interpreted by Richmond, who mapped it as the fifth moraine of the Bull Lake glaciation at Fremont Lake (QBt5-FL) (*2*). The Bull Lake age of this boulder gives us confidence that the outermost boulders on QPt1-FL correspond to the beginning of the Fremont Lake lobe LPGM (21,700 ± 700 ^{10}Be years).

During and after the LPGM, the three lobes formed well-developed lateral moraines, recessional moraines nested within the terminal moraine, and interlobate moraines where the lobes coalesced (Fig. 1). The similar ice volumes of the Fremont Lake, Half Moon Lake, and Soda Lake lobes, the equal terminal elevations, and the shared accumulation zones imply that the three lobes behaved synchronously throughout the Pinedale glaciation. The ^{10}Be dates from two boulders (WY-92-119 and WY-92-117; mean = 19,400 ± 800 ^{10}Be years) from the QPt1-HL moraine and from a boulder (samples WY-91-003 and WY-91-004; mean = 19,900 ± 200 ^{10}Be years) on QPt1-SL lie within the range of ages of QPt1-FL (Table 1 and Fig. 2) and support the implication.

We determined exposure ages for 11 boulders on Pinedale recessional moraines of the Half Moon Lake lobe (QPt2-HL to QPt5-HL) and Fremont Lake lobe (QPt2-FL and QPt4-FL) where the Pinedale recessional moraines are best preserved (Fig. 1). The ^{10}Be exposure ages of the four Half Moon Lake recessional moraines overlap (Table 1 and Fig. 2). The ^{10}Be chronology indicates that once retreat began, all four moraines were deposited within a short time, approaching the resolution time of the ^{10}Be technique to date a single deposit. If all four moraines are assumed to be coeval, the uncertainty in their combined mean, 16,000 ± 900 ^{10}Be years, is less than 1000 years, implying an initial average retreat rate no less than 2 m year^{-1}. The coefficient of variation (including analytical and geological uncertainties) for a single sample measurement about the mean age of the nine boulders measured on the Half Moon Lake recessional moraines is 4.1%. This is the best estimate of the precision with which the age of a glacial deposit can be determined from the exposure age of a single boulder.

Considering the other 25 dates from the Fremont Lake and Half Moon Lake Pinedale moraines, the exposure age of the two boulders on QPt2-FL are too old (Table 1). Although the samples have not been remeasured yet, we suspect that these boulders may have inherited ^{10}Be from a prior exposure. The age of the QPt4-FL moraine, 15,500 ± 600 ^{10}Be years, is indistinguishable from the age of the QPt4-HL moraine and the other recessional moraines of the Half Moon Lake lobe (Table 1 and Fig. 2), again implying that the ice lobes behaved synchronously.

We conclude that the main phase of deglaciation from the Half Moon Lake glacial maximum position began by 16,000 ± 700 ^{10}Be years. The QPt1-HL terminal moraine provides additional support for a LPGM in the Fremont Lake basin lasting at least 3400 years (19,400 ± 800 to 16,000 ± 700 ^{10}Be years). If the range in ages from the QPt1-FL terminal moraine represents the entire duration of the last glacial maximum, then retreat from the QPt1-FL moraine began no earlier than 15,800 ± 500 ^{10}Be years ago.

To further date the last major deglaciation, we measured ^{10}Be in bedrock surfaces along a transect from the recessional moraines to a series of post-Pinedale moraines near the Wind River Range divide [Temple Lake glaciation equivalent (*24*)]. Samples WY-91-039 and WY-91-028 are from glacially polished bedrock surfaces situated behind the moraines on the distal side of Half Moon Lake (Fig. 1). The ^{10}Be dates for these two >HL5 samples are younger (15,200 and 12,200 ^{10}Be years) than for the recessional moraines (Fig. 2). The dates obtained for these samples record the time when till cover was removed from the surfaces and therefore provide a minimum date for ice retreat. As the three ice lobes appear to have behaved approximately synchronously, retreat from the entire Fremont Lake basin was complete by 15,200 ^{10}Be years ago.

A third striated bedrock surface located at Seneca Lake, approximately 25 km in the direction of ice retreat from the Pinedale terminal moraine, yielded an exposure age of 15,900 ± 600 ^{10}Be years. This exposure age indicates that ice on the Wind River Range highland plateau (3250 m) dissipated approximately coevally with the main phase of ice retreat from the Fremont Lake basin, and consequently the highland plateau may have been deglaciated even before ice completely left the lowland valleys (*25*). Although complete deglaciation of highland plateau surfaces occurring before outlet valleys surfaces has been documented elsewhere (*26*), we cannot rule out the possibility that the ages are biased by preexposure.

Another bedrock surface (WY-93-345) and an adjacent boulder (WT-93-344) situated within 4 km of the Wind River Range divide (Fig. 2) yielded a mean age of 12,100 ± 600 ^{10}Be years. These dates indicate that the Wind River Range ice cap had completely disappeared by this time and its only remnants were expanded cirque glaciers. On the basis of these ages, the average systematic rate of ice retreat from the Fremont Lake basin was 7.4 m year^{-1}. Immediately in front of the Titcomb Lakes moraines a fifth bedrock surface (WY-93-343) yielded an exposure age of 13,300 ± 600 ^{10}Be years. This age reinforces the idea that the rate of deglaciation was rapid, although we suspect that it too may reflect the presence of a small concentration of inherited ^{10}Be.

The exposure age of the Titcomb Lakes moraines (QTLt, Fig. 1), 11,700 ± 600 ^{10}Be years based on 11 boulders (*13*), supports their correlation with type–Temple Lake moraines. If the Titcomb Lakes moraines are Pinedale recessional moraines, their age represents the time of the completion of deglaciation. However, the moraines may have been deposited after a readvance during the Younger Dryas cooling event (*13*), and, if so, deglaciation must have culminated at Titcomb Lakes earlier. All post–Pinedale glacial maximum advances of the cirque glaciers in the Titcomb lakes basin, including the large expansion at 11,700 ± 600 ^{10}Be years, were within 2 km of the cirque wall and above the Pinedale equilibrium line altitudes for the Wind River Range.

Our results demonstrate that cosmogenic nuclides can be used to date geomorphic landforms deposited within the past 20,000 years with single-sample precisions better than ~4%. Apparent differences between calibrated radiocarbon ages and ^{10}Be ages may diminish when production rates for ^{10}Be during the past 20,000 years are better known. Despite differences in source and terminus elevations, glacier volumes and geometries, and local climate, Pinedale

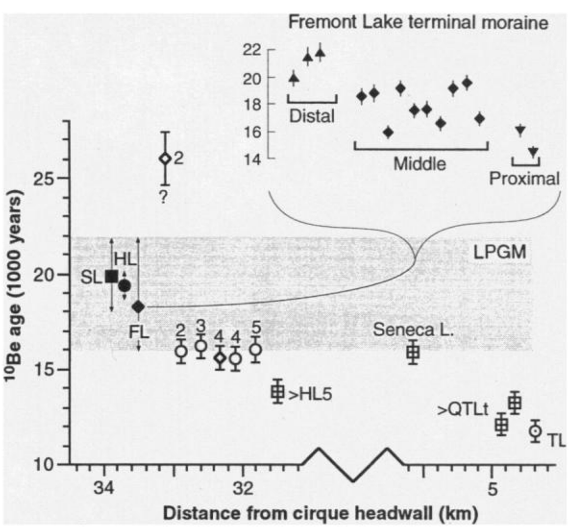

Fig. 2. The distribution of calculated ^{10}Be ages used to constrain the timing of the last glacial maximum and deglaciation. Terminal moraines are represented by solid symbols; their minimum duration is indicated by a double-headed arrow through a symbol placed at the average of the measured samples. Soda Lake (SL) lobe moraines are identified by squares, Half Moon Lake (HL) lobe moraines by circles, and the Fremont Lake (FL) lobe moraines by diamonds. Open symbols refer to samples from the recessional moraines QPt2 to QPt5. Crossed boxes are bedrock surfaces. The dotted circle is the mean of 11 boulder ages from the two Titcomb Lakes (TL) moraines (Fig. 1). The shaded area labeled LPGM marks the time when the ice margins were positioned near the last Pinedale glacial maximum position. Sample WY-91-034 could not be shown at this scale. Error bars (1σ) on individual samples represent a total analytical reproducibility of 3%, as discussed in the text; range bars on landform ages are as given in Table 1. (Inset) Samples from the FL lobe terminal moraine are arranged in order of their positions behind the leading (distal) edge of the moraine: The outer (distal), middle, and inner (proximal) portions of the broad ridge are indicated.

ice fluctuations at the type-locality suggest that deglaciation of the three Fremont Lake basin glacial lobes advanced and retreated approximately coevally with most other Rocky Mountain glaciers (*6–8, 27*). Moreover, the deglaciation history of the Fremont Lake basin is in excellent agreement with the corresponding history of portions of the Laurentide Ice Sheet for which ages are well constrained (*28*) and with many marine isotope and ice core records of the last deglaciation (*29*).

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16. Splits from samples WY-91-003 and WY-91-035 were completely reanalyzed, yielding concentrations within 0.5% of their mean (5.18×10^5 and 5.13×10^5 atoms per gram and 6.25×10^5 and 6.23×10^5 atoms per gram, respectively).
17. The calibration of (*15*) based on 16 samples, measured on surfaces in the Sierra Nevada with a calibrated radiocarbon age of 11,000 years, gives a production rate of 6.01 $\pm$ 0.4 atoms per gram per year at sea level for latitudes >60°. The altitude and latitude factors of D. Lal [*Earth Planet. Sci. Lett.* **104**, 424 (1991)] were used to adjust the production rate to the altitude and latitude of Pinedale (~35 atoms per gram per year). Sources of uncertainty in this calibration applied to Pinedale samples are (i) the assigned age of 11×10^3 years for the glacially striated surfaces [D. H. Clark and A. R. Gillespie, *Geol. Soc. Am. Abstr. Programs* **26**, A-447 (1994)]; (ii) possible variations in the strength of Earth's magnetic field {both a comparison of radiocarbon ages and U-Th ages in coral [E. Bard *et al.*, *Nucl. Instrum. Methods Phys. Res. B* **52**, 461 (1990)] and paleointensity measurements of M. W. McElhinny and W. E. Senanayake [*J. Geomagn. Geoelectr.* **34**, 39 (1982)] suggest that Earth's magnetic field was weaker 10,000 to 20,000 years ago than between 0 and 10,000 years ago, so production rates averaged over the past 20×10^3 years may have been 7% higher than those given in (*15*)}; and (iii) uncertainties in the average geomagnetic latitude of Pinedale introduce an uncertainty of about 10% in the production rate.
18. Ages are reported in units of ^{10}Be years, by analogy with the convention used for radiocarbon dates. This convention recognizes that there may be differences between the calendric time scale and the ^{10}Be time scale because of uncertainties in the production rate of ^{10}Be (*17*).
19. The uncertainty in the age range is the square root of the sum of the squares of the uncertainties of the oldest and youngest dates.
20. We calculated the maximum erosion rate, ϵ_{max} (in millimeters per year), by assuming that the concentration of a cosmogenic nuclide is determined by the rate of erosion and not exposure time; that is, the rate of removal of cosmogenic atoms by erosion is equal to the rate of their production by cosmic rays:

$$\epsilon_{max} \approx 3400/N_{10} \approx 566/T_{eff}$$

where T_{eff} is the apparent age of the sample, ρ = 2.65 g cm^{-3}, and N_{10} is the concentration of ^{10}Be (in 10^6 atoms per gram normalized to sea level, high latitude).
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Surface Displacement of the 17 May 1993 Eureka Valley, California, Earthquake Observed by SAR Interferometry

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Satellite synthetic aperture radar (SAR) interferometry shows that the magnitude 6.1 Eureka Valley earthquake of 17 May 1993 produced an elongated subsidence basin oriented north-northwest, parallel to the trend defined by the aftershock distribution, whereas the source mechanism of the earthquake implies a north-northeast–striking normal fault. The $\pm$3-millimeter accuracy of the radar-observed displacement map over short spatial scales allowed identification of the main surface rupture associated with the event. These observations suggest that the rupture began at depth and propagated diagonally upward and southward on a west-dipping, north-northeast fault plane, reactivating the largest escarpment in the Saline Range.

Measurement of the surface displacement of the Earth caused by a large earthquake is important for understanding its mechanism (*1*). Dense geodetic arrays to monitor surface deformation of the crust can feasibly be set up in only a few areas. In remote regions, SAR interferometry (*2, 3*) can provide high-resolution maps of coseismic surface displacements over broad areas, giving new insights into fault geometry.

We studied the magnitude (*M*) 6.1 Eureka Valley earthquake that occurred on 17 May 1993 on the border between California and Nevada. This earthquake occurred at a depth of 13 km along the west side of the Eureka Valley (*4*) (Fig. 1), one of the westernmost valleys in the wide zone of extension distributed across the Great Basin (*5*). The focal mechanism of the main shock indicates that the earthquake ruptured a north-northeast–striking fault, steeply dipping to the west (*4*). The aftershocks define a north-northwest trend (*6*) and include two shocks of M ~ 5 and several of M > 4. Small surface ruptures formed in the central part of the Eureka valley (*7*) (arrow A1, Fig. 1).

To map the earthquake displacement, we combined SAR images of the epicentral region acquired by the European remote sens-

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