Regional Correlations of Late Pleistocene Climatic Changes Based on Cosmogenic Nuclide Exposure Dating of Moraines in Idaho

Cody Sherard

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REGIONAL CORRELATIONS OF LATE PLEISTOCENE CLIMATIC CHANGES BASED ON COSMOGENIC NUCLIDE EXPOSURE DATING OF MORAINES IN IDAHO

By Cody Sherard
Accepted in Partial Completion of the Requirements for the Degree of Master of Science
Western Washington University 2006

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Cody Nelson
March 24, 2018
REGIONAL CORRELATIONS OF LATE PLEISTOCENE CLIMATIC
CHANGES BASED ON COSMOGENIC NUCLIDE EXPOSURE DATING OF
MORAINEs IN IDAHO

A Thesis
Presented to
The Faculty of
Western Washington University

In Partial Fulfillment
Of the Requirements for the Degree
Master of Science

By
Cody Sherard
August 2006
Abstract

Multiple post-last glacial maximum (LGM) moraines occur in the Sawtooth Range of the Rocky Mountains in central Idaho. Although relative ages of the moraines have been studied, few numerical ages exist for these moraine sequences. In this study, the ages of LGM and late glacial (Younger Dryas) moraines in the Redfish Lake drainage of Idaho are constrained with 23 new, cosmogenic, $^{10}$Be dates. Results of this study include developing a new chronology for the Redfish Lake late Pleistocene moraines that are consistent with regional morainal ages.

New cosmogenic $^{10}$Be data from boulders on Bull Lake moraines at Redfish Lake are consistent with other dates obtained from Bull Lake moraines elsewhere in the Rocky Mountains. Ages of 107 to 206 ka and 238 to 323 ka suggest synchronicity with other Bull Lake glacial advances in the Wind River Range, Tetons, and Yellowstone.

New cosmogenic $^{10}$Be data from boulders on terminal moraines in Redfish Lake valley, Idaho, show a LGM at 18.5 ± 0.9 ka to 15.6 ± 0.8 ka, recessional (or minor readvance) moraines deposited between 15.4 ± 0.7 ka and 14.2 ± 0.7 ka; and a Younger Dryas (YD) phase at 11.4-11.7 ± 0.5 ka. Boulders on an LGM moraine at Alturas Lake, Idaho were $^{10}$Be dated in order to supplement and test the existing concepts of glaciation there. Ages of 17.2 ± 1.0 ka and 16.8 ± 0.8 ka indicate that this moraine is late Pinedale, not mid-Pinedale as previously suggested by Thackray et al. (2004).

The late Pleistocene ice chronology at Redfish Lake is consistent with the glacial chronology in the Wind River Range, WY (LGM 24 ka to 18 ka; recessional deposits 18 ka to 16 ka and YD 11.5 ka); Yellowstone, WY (LGM advances: 29.5-22.5 ka, and 19.5-15.5 ka); Wallowa Mountains, OR (two $^{10}$Be dated LGM advances: 21.1 ± 0.4 ka and
11,330 ± 220-10,100 ± 70 ¹⁴C yrs B.P.); and many other places throughout western North America. This research provides chronological constraints on Pinedale LGM (~18.5 ka) moraines, Pinedale recessional (between 18.5 and 14.2 ka) moraines, and Younger Dryas (~11.4 ka) moraines in the Sawtooth Range of the Rocky Mountains, and assists in determining the areal extent of the late Pleistocene cold period in the Western United States.
Acknowledgements

I would like to thank the following people who have helped immensely in the completion of this project. My thesis committee; Dr. Don Easterbrook, Dr. Scott Babcock, Dr. John Gosse, and Dr. Scott Linneman; those who helped in the field, lab and office, Dr. John Gosse, Lisa Stockli, Dr. Doug Clark, Quentin Zumhoffe, Jeffery Weiss, and Shannon Petrisor, thank you for your continuing support throughout this project. A special thanks to Don Easterbrook; without his substantial guidance and assistance, this project would not have been possible.

I would also like to thank the Geological Society of America and the WWU Geology Department for their assistance in funding this project.

This study is part of a larger study by Dr. Don Easterbrook (Western Washington University) and by an international group of geologists to determine if glacial/climatic changes during the late Pleistocene occurred synchronously in western North America, Europe, and the southern hemisphere.
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1. Introduction

In North America, studies have suggested two phases of the last glacial maximum (LGM), one occurring around 21 ± 2 ka and the other occurring around 17 ± 2 ka. Deposits of both ages are recorded in some areas, whereas deposits of only one age are recorded in others. These ages are based on years of accumulation of data through extensive studies across North America (the literature is too extensive to list in this paper, see references for specific areas of study related to this paper).

The Cordilleran Ice Sheet (CIS) and alpine glaciers throughout western North America responded to this fluctuating climate, depositing moraines until about 17 ka (e.g., Armstrong et al., 1980; Easterbrook, 1992, 1994; Gosse et al., 1995b; Licciardi et al., 2004). Several thousand years after the onset of the recession of the CIS and alpine glaciers, the abrupt (~20-60 yr onset) Younger Dryas (YD) cooling period took place and lasted about 1000 years (documented extensively by moraines throughout the world and in the Greenland ice cores: 11,500 to 10,200 $^{14}$C yrs. B.P.; 12.7 to 11.5 cal ka; Alley et al., 1993; Stuiver et al., 1995; Alley, 2000). The YD was recorded by alpine glaciers and ice sheets in many regions, but not in others. This cooling event is the second of two of the largest climate changes since the LGM, the other being the Intra-Allerød Cold Period (IACP) shortly before about 11,500 $^{14}$C yrs B.P. (13.4 cal ka; Johnsen et al., 1992; Dansgaard et al., 1994; Stuiver et al., 1995).

Multiple post-LGM moraines occur in the Sawtooth Range in central Idaho (Figures 1, 2), but until recently, were not well dated. Relative dating techniques have been applied to these moraines (Williams, 1961; Breckenridge et al., 1988), but a dependable age correlation cannot be made with other moraines in the region without numerical dates. The Redfish Lake drainage in Idaho was chosen for study because of its spectacular, well-preserved
moraines laden with granitic boulders suitable for $^{10}$Be cosmogenic nuclide exposure dating. Dating of the Redfish Lake morainal sequence allows comparison with moraines in the Stanley Basin at nearby Alturas, Pettit, and Yellowbelly Lakes and in Hell Roaring Valley (Borgert, 1999; Lundeen, 2001; Thackray et al., 2004) (Figure 1), and provides additional insight into late Pleistocene glacial advances in this region.

A numerical glacial chronology is needed to compare events from region to region. Relative dating techniques are, at best, useful for relative chronology in adjacent valleys. Even then, differences in authors’ opinions and observations can create discrepancies in interpretation of glacial sequences. The results presented herein are from cosmogenic $^{10}$Be dates, and all ages are reported with 2$\sigma$ precision. Age comparisons are made in this study between dates obtained through different methods (i.e., $^{14}$C, $^{36}$Cl, and $^{10}$Be) without normalizing production rates and re-scaling comparison data. This comparison of data without normalizing production rates and re-scaling can result in age differences as high as 15%. Correlation between the new ages reported here and other ages at a high level of precision is complicated because dating techniques have improved over the last 10 years and these improvements have caused changes in production rates and scaling calculations that result in changes in ages that are on the order of thousands of years in the late Pleistocene. However, establishing a numeric chronology of alpine glacier fluctuations in the Sawtooth Mountains can be used to evaluate the synchronicity of LGM and late Pleistocene (specifically YD) alpine glacier
Figure 1. DEM overview of the Stanley Basin, Idaho, bordered by the Sawtooth Mountains on the west and the White Mountains on the east. The massive, well defined moraines of Redfish Lake are visible at the north end of the valley. To the south are the smaller, albeit also impressive moraines of Hell Roaring Fork, Yellowbelly Lake, Pettit Lake and Alturas Lake. Figure courtesy of Louden Stanford, Idaho Geological Survey.
fluctuations in North America with better precision than through relative dating techniques.

II. Previous Work

The general Pleistocene glacial sequence in the Rocky Mountains recognized by Blackwelder (1915), based on field mapping of moraines and terraces on the northeast flank of the Wind River Mountains of Wyoming, is as (follows from oldest to youngest):

- **Buffalo Glaciation**, composed of three stades termed Cedar Ridge Glaciation, Sacagawea Ridge Glaciation and Dinwoody Lake Glaciation
- **Bull Lake Glaciation**, composed of an early and late stade
- **Pinedale Glaciation**, composed of three stades (early, middle, and late)

Since 1915, these glaciations have been considered the typical sequence for correlation in the Rocky Mountains. In 1962, the terms were accepted by the American Commission on Stratigraphic Nomenclature.

*Sawtooth Mountains, Idaho*

**Bull Lake Glaciation**

Williams (1961) mapped Pleistocene moraines in the Sawtooth Mountains and developed a relative chronology based on topographic and weathering criteria. The Bull Lake moraines are distinguished by a greater degree of erosion than Pinedale moraines and they have more subdued crests and less steep sides. Boulders are less abundant, more rounded, and have more spalls.
Figure 2. Redfish Lake and associated moraines, Sawtooth Mountains, Idaho. DEM courtesy of L. R. Stanford, Idaho Geological Survey.
Williams (1961) recognized two groups of Bull Lake moraines, Bull Lake I (older) and II (younger) distinguished by their relative topographic positions. Bull Lake I moraines (herein called Elk Meadow moraines) are oriented to the northeast and are limited to lateral moraines and a very dissected group of terminal moraines (Figure 2). These are cross-cut by the Bull Lake II moraines (herein called Salmon River moraines) that are oriented south, parallel to the range front. The Salmon River moraines include two sets of bounding lateral moraines and several recessional moraines.

**Pinedale Glaciation**

Williams (1961) recognized late Pleistocene Pinedale moraines in all of the large drainages in the western Stanley Basin. They are steep-sided, sharp-crested, and have some depressions. Large, slightly weathered, granitic boulders are abundant on these moraines. Outwash extent is limited on the west side of the basin (as compared to the east in the White Cloud Mountains) so disruption of moraines by fluvial erosion is minimal.

At Redfish Lake (elevation ~1770 m, 5800 ft) Williams (1961) mapped extensive Pinedale moraines that are up to 300 m high and extend nearly 10 km into the basin, forcing the Salmon River to the eastern side of the basin. More than a dozen nested moraines lie upvalley from the Pinedale maximum at Redfish Lake. They are less massive and more discontinuous than the outermost moraines, ranging from 5-20 m high (Figure 3).

Pinedale recessional moraines are present in several adjacent valleys as well. These include the “Fourth of July Creek” deposits and the “Milky Creek” deposits mapped by Breckenridge et al. (1988).
Figure 3. Redfish Lake morainal group, Redfish Lake, Idaho. Moraine crests indicated in blue. Boulders sampled and dated for this study are indicated in red, and boulders visited but not suitable for sampling are indicated with a blue X. ID-RL-04-012 and ID-RL-04-013 are the most terminally located boulders dated, and are located along the pinedale terminal moraine. They are $^{10}$Be dated at $15.7 \pm 0.8$ ka and $18.5 \pm 0.9$ ka respectively. Dates from various nested Pinedale moraines range from $14.2 \pm 0.7$ ka to $15.5 \pm 0.7$ ka (see Table 1 for sample numbers).
Figure 4. Bench Lakes morainal group located northwest of Redfish Lake, Idaho. Moraine crests are indicated in blue. Boulders sampled and $^9$Be dated for this study are indicated with a red dot, boulders sampled but not yet dated are indicated with a green dot, and boulders visited but not suitable for sampling are indicated with a blue X.
Latest Pinedale moraines

Williams (1961) mapped late Pinedale moraines around the outlet end of four of the five Bench Lakes, located northwest of the Redfish Lake left lateral moraine (Figure 4). Post-Pinedale glaciation is also indicated by some other small north and northeast facing cirque moraines above 9,000 ft (Williams, 1961).

Recent studies in adjacent valleys

Thackray et al. (2004) obtained radiocarbon dates from lake and marsh cores between moraines in four of the southernmost valleys in the western Stanley Basin. Seven to nine moraines were mapped and named in valleys of Hell Roaring Creek, Yellow Belly Lake, Pettit Lake, and Alturas Lake (Figure 5). Minimum limiting dates of moraines from the valleys of Pettit Lake and Yellow Belly Lake are 13.94 ± 0.15 cal ka and 13.96 ± 0.14 cal ka (\(^{14}\)C dated). In the valley of Alturas Lake, the minimum limiting date for the Perkins Lake (Pinedale) moraine group is 16.86 ± 0.41 cal ka, and Thackray et al. (2004) suggest it represents the maximum Pinedale advance. This date also represents the youngest age for their Busterback Ranch moraine group farther downvalley. A core from Moose Kettle in the valley of Alturas Lake contains a minimum limiting date of 15.59 ± 0.17 cal ka (\(^{14}\)C dated) from basal peat on a Busterback III moraine (Figure 6) (Borgert, 1999). Borgert (1999) suggests that this material beneath the peat is possibly of non-glacial origin (post-glacial slumping, sliding, or moraine denudation processes).

Thackray et al. (2004) proposed that the older set of moraines in these valleys, the Busterback Ranch I, II, III moraines, are at least 10 ka older than the younger moraines.
Figure 5. Moraine groups in the Alturas, Pettit, Yellow Belly, and Hell Roaring valleys, mapped by Thackray et al., 2004. They conclude that the Group 1 Busterback (Bull Lake equivalent) moraines are 10,000 years older than the Group 2 Perkins Lake (Pinedale equivalent) moraines.
Figure 6. Alturas Lake Valley moraines mapped by Borgert, 1999. Moose Kettle core is located upvalley from the Busterback Ranch III moraine (MK) and gave a basal date of $15.6 \pm 0.2$ cal ka. Lost Boots Marsh core (LBM) gave a basal date of $16.9 \pm 0.4$ cal ka. Moraine Bog (MB) yielded no dates.
and more likely are at least 27 ka, based on moraine morphology and weathering characteristics of boulders. This interpretation would mean that the outmost moraines in each of these valleys are older than the late Pinedale LGM.

Thackray et al. (2004) acknowledge that mountain glacier advances are commonly attributed largely to cooling, but they hypothesize that increased precipitation was the dominant factor in the alpine glacier advances in the Sawtooth Mountains and that “the end of the last glaciation involved large-scale reorganization of oceanic, atmospheric, and cryospheric systems”. This hypothesis is based on the assumption that atmospheric circulation in the western U.S. was dominated by changes produced by the Laurentide Ice Sheet far to the east. If that assumption is correct, Younger Dryas moraines should be absent from the Sawtooth Mountains. Thackray et al. (2004) found no evidence of YD moraines in the valleys they studied and concluded that the youngest moraines predated the YD by at least 1 ka on the basis of the 13.96 ± 0.14 cal ka age and 13.94 ± 0.15 cal ka age from their McDonald and Pettit Lake cores and a 13.98 ± 0.12 cal ka date from Lost Boots Marsh core. However, they did not rule out a YD equivalent advance of cirque glaciers in the high Sawtooth Mountains upvalley.

II. Cosmogenic Nuclide Exposure Dating Methods

$^{10}$Be Dating

Bombardment of rocks by cosmic radiation produces in situ, radioactive and stable isotopes and cosmogenic nuclide dating techniques rely on the measurement of those isotopes. Six cosmogenic nuclides are widely used in the study of geology, $^{10}$Be, $^{26}$Al, $^{14}$C, $^{3}$He, $^{21}$Ne, and $^{36}$Cl. Nishiizumi et al. (1989) were the first to study in situ $^{10}$Be and $^{26}$Al.
radionuclides from eroded rock surfaces. High-energy galactic nucleons (primarily protons) with energy \( \leq 10^{10} \text{ GeV} \) produce nuclear disintegrations in the upper atmosphere (Lingenfelter and Flamm, 1964). The products of these interactions are collectively known as secondary cosmic radiation and it attenuates rapidly as it approaches the earth due to absorption by the atmosphere. Cosmogenic nuclides such as \(^{10}\text{Be}\) are produced in quartz by nuclear spallations of oxygen and silica when a neutron collides with a \(^{28}\text{Si}\) or \(^{16}\text{O}\) nucleus, and breaks the nucleus into several lighter particles (Templeton, 1953; Lal, 1991; Gosse and Phillips, 2001). For example, \(^{10}\text{Be}\) and \(^{26}\text{Al}\) are almost exclusively produced \textit{in situ} by spallation reactions. Muons interact with nuclei in oxygen and silica, but account for less than 18% of the production at sea level (Lal, 1991; Nishiizumi et al., 1994; Stone, 2000; Gosse and Phillips, 2001). Because the production rate of cosmogenic \(^{10}\text{Be}\) in quartz is reasonably well calibrated for the time scale of interest in this study (Stone, 2000; Gosse and Phillips, 2001), sampling of rocks with an abundance of quartz was a priority. The granite boulders sampled in the field areas have between 10% and 25% quartz.

Many atmospheric and physical factors affect the production rate of cosmogenic nuclides, which varies through time in accordance with the flux of cosmic radiation reaching the surface. Gosse and Phillips (2001) cite the controlling factors as: variations in primary cosmic radiation intensity, changes in the magnetic field and solar modulation of galactic cosmic rays, dipole and non-dipole changes in the geomagnetic field effects on galactic cosmic rays, variations in atmospheric shielding, elevation of samples, and long term changes of the ground surface.
The standard model used in scaling in situ cosmogenic nuclide production rates with a reasonable uncertainty of $<$10%, was developed by Lal (1991). The amount of the cosmogenic nuclide produced depends on:

(1) the decay constant of the isotope,

(2) the production rate at depth in the rock:

\[ P_x = P_0 e^{(kx)} \]  

(Gosse and Phillips, 2001)

\( P_x \) is the production rate at depth \( x \), in cm

\( P_0 \) is the production rate at the rocks surface (5.1 atom g\(^{-1}\) yr\(^{-1}\) for \(^{10}\)Be (Stone, 2000))

\( k \) is a density dependant constant (cm\(^{-1}\)) and is equal to \( \rho/\Lambda \), where \( \rho \) is the bulk density of the rock and \( \Lambda \) is the mean attenuation length for the given cosmic radiation.

(3) The amount of time the surface has been exposed to radiation,

(4) The erosion rate of the rock’s surface,

(5) Any inherited component of the isotope concentration (Gosse and Phillips, 2001),

(6) Any disturbance in exposure due to rolling, snow or ash cover, etc.

Ideally, the time since boulder deposition is therefore obtained since the length of exposure since deposition is measured. All samples dated for this paper are reported in \(^{10}\)Be ka.

The absolute production rates for \(^{10}\)Be and \(^{26}\)Al nuclides were first measured in rocks of the Sierra Nevada by Nishiizumi et al. (1989). Their production rate was based on 16 samples with an assumed age of 11,000 cal yrs B.P., which yielded a production rate in quartz of \( 6.01 \pm 0.4 \) atoms g\(^{-1}\) yr\(^{-1}\) (sea level, above 60\(^{\circ}\) lat.). Stone (1999, 2000) has
calculated an average global production rate of 5.1 ± 0.3 \(^{10}\)Be atoms·g\(^{-1}\)yr\(^{-1}\) and is the value used in this study. This is based on a 2.2% contribution of muons to the production of \(^{10}\)Be at sea level, and gives reasonable production rates for both high and low elevations. The production rate can be adjusted for any altitude because atmospheric shielding reduces the intensity of radiation and for latitude because magnetic field strength varies spatially (Lal, 1991; Gosse and Phillips, 2001). Generally, production rates are lower at lower elevations and higher at higher elevations because radiation intensities increase with altitude, and decrease at lower altitudes (e.g., Stone et al., 1998; Stone, 2000). Total uncertainty in production rates is estimated at 10-20% for \(^{10}\)Be (Gosse and Phillips, 2001).

Variations in the cosmogenic nuclide production rate in rocks of the same composition and at the same altitude and latitude are due mostly to inheritance and shielding. This includes snow, sediment cover, vegetation shielding (mostly tall, dense forests), topographic shielding (mountains), and sample inconsistencies such as edge effects. Edge effects are the dispersal of cosmic particles out of the edge or a corner of a rock before they interact with the mineral nuclei in the rock. Sampling should be done no closer than 30 cm from the edge of a rock surface to minimize dispersal through the edge of the boulder. Gosse and Phillips have developed an equation to estimate the yearly effects of shielding by snow,

\[
S_{\text{snow}} = \frac{1}{12} \sum_{i} \left( \frac{Z_{\text{snow}, i} - Z_{\text{sample}}}{\rho_{\text{snow}, i} \Lambda_{f,e}} \right) e \left( -\frac{Z_{\text{snow}, i} - Z_{\text{sample}}}{\rho_{\text{snow}, i} \Lambda_{f,e}} \right) \]

\begin{align*}
Z_{\text{snow}, i} & \text{ is the monthly avg. snow height (cm) above the sample surface} \\
Z_{\text{sample}} & \text{ is the height of the boulder (cm)} \\
\rho_{\text{snow}, i} & \text{ is the avg. monthly snow density} \\
\Lambda_{f,e} & \text{ is the effective attenuation length}
\end{align*}

(Gosse and Phillips, 2001)
The effects of shielding by snow can be seen in Figure 7 (Figure 17 of Gosse and Phillips, 2001). Snow density and rock shielding cause additional absorption of cosmic particles. A younger age is reflected in the rock when snow is deeper, denser, and coverage persists for a longer amount of time.

Nuclide production rates by spallation decrease exponentially with depth in a rock. Flat-topped rocks are ideal for sampling because the influence of foreshortening (reduction of cosmic ray flux on a dipping surface) is minimized. Mean attenuation lengths are affected because radiation enters the rock at oblique angles (Gosse and Phillips, 2001). This does not, however, affect samples obtained for $^{10}$Be age dating because they are only 2-3 cm thick. Error due to topographic shielding effects is generally low because about 80% of the effective cosmic radiation comes from angles $>45^\circ$ from horizontal, so topographic barriers have to be quite large or the sample must be very near the base of the barrier. Attenuation lengths for shielded rocks (due to mountains or trees, but not snow or sediment cover) are longer because the radiation enters the rock from near vertical angles.

Snow cover prevents some radiation from reaching rock surfaces, depending on snow depth and density (see Figure 7). For example, at Redfish Lake, if an average one meter of common density (0.2 g/cm^3) snow accumulates on a boulder surface during three months per year, an age discrepancy of approximately 3% could occur (the exposure age would be younger than it actually is). For a boulder dated at 18,000 yrs B.P., that is equal to a decrease in apparent age of about 540 years. On the other hand,
Figure 7. The effects of shielding by snow. Deeper, denser snow creates a larger difference in perceived age (younger) and true age. From Gosse and Phillips, 2001.
one meter of dense snow (0.4 g/cm^3) could cause a younger apparent age by approximately 6%; for the same boulder, that is 1080 years.

For this study, the estimated age reduction by snow was not calculated due to a record of insignificant snowfall at Banner Summit SNOTEL station. Snow depths were recorded at an average of between 60 to 70 cm for six years between 2000 and 2006. This amount of snow would not be sufficient to build up next to, and cover the surface of the boulders sampled. Because the snow accumulation is not significant for the record given, no correction was made. There is no way to know historical snow depths however, especially during Pleistocene glacial periods, when snowfall was likely very great, and snow depth may have been on the order of meters, not centimeters. However, most of the post-depositional history of Pinedale boulders is post-Pleistocene.

Erosion rates may not be insignificant, particularly for long (>50 ka) exposure ages (Gosse and Phillips, 2001). In this study, the effect of boulder erosion is examined by assuming constant erosion rates of the boulder top during the exposure time.

Differences in lithology of the Sawtooth and Idaho granitic batholiths, from which erratics originate, may contribute to different erosion rates. The Idaho batholith is a Cretaceous, grey to white, coarse-grained, equigranular biotite-granodiorite (Worl et al., 1991). The Sawtooth batholith is an Eocene, pink, coarse-grained, equigranular to porphyritic granite (Worl et al., 1991). The Eocene granite may be more resistant to erosion and boulders on moraines are 0.5 m to several meters larger than those composed of Cretaceous granodiorite. Dates from the Eocene granite may have less uncertainty due to weathering. Ages obtained are, however, a minimum due to uncertainties in erosion rates. Additionally, the difference in boulder size could be due to the original fracture density in the
source rock. The Cretaceous unit may have undergone more intensive fracturing and therefore had smaller spacing between fractures than the Eocene granite, producing smaller boulders. Conversely, the Eocene granite may be highly fractured and vulnerable to glacial plucking, whereas the Cretaceous granite could be more massive and less vulnerable to plucking.

Field methods/Sample collection

A cosmogenic nuclide date is only as good as the sample from which it was taken and the method by which it was obtained. In this study, every effort was made to maintain stringent sampling procedures for each boulder. Sampling boulders for cosmogenic nuclide dating was undertaken during the summer of 2004. Attempts were made to collect samples from a minimum of three boulders from each moraine, but this was dependent on the occurrence of boulders suitable for dating. Only boulders with surfaces >1 m (for this study an average of 1.7 m, n=23) above ground and with their base buried were sampled to minimize the effects of possible boulder movement and maximize the probability that the boulder has been continually exposed in situ since deposition.

Where possible, the same lithology was sampled to avoid small variations in isotope concentrations due to different weathering rates. This was generally not a problem, as most of the boulders in the Redfish Lake area are hard Eocene granite, with only a few less resistant Cretaceous granodiorites. The thickness of each sample collected was ≤2 cm in order to obtain a maximum concentration of $^{10}$Be near the surface. Samples were collected with a portable cutoff saw, chisel, and hammer. The cutoff saw insured a consistent sample depth in the center of the top of the boulders away from the edges, and allowed sampling of
much harder lithologies than possible with a chisel and hammer. Approximately 1000-2000 grams of sample were collected from each boulder.

Sampling from a flat surface is useful in order to simplify the calculations for cosmic ray flux into the rock. Samples in this study were taken from surfaces sloping <3° (Table 1), so geometry corrections were unnecessary (much less than a 1% effect). Additionally, samples were collected at least 30 cm from any edge in order to reduce edge effects. For samples that were topographically shielded, the angle to the obstruction above horizontal was recorded approximately every 10° in a 360° radius around the sample. These measurements were used to calculate the decrease in cosmic ray flux to the shielded samples (Gosse and Phillips, 2001).

Uncertainty in calibrating production rates due to differences in altitude and latitude can make comparisons between cosmogenic nuclide exposure ages and ages obtained through other methods difficult. The accuracy of the 10Be dates can be checked with 26Al from the same rock. However, due to lab and time constraints, aluminum was not measured in the rocks for this study. Human error is possible during sampling and lab processing, but every effort was made to minimize contamination and maintain reproducibility.

**Laboratory methods**

Sample preparation for extraction of 10Be was completed at the Cosmogenic Nuclide Extraction Laboratory (CNEL) at the University of Kansas in Lawrence, Kansas. All methods were according to the Dal-CNEF Lab Specific Standard Operating Procedure by John Gosse (updated 2004) and from Gosse and Phillips, 2001. The objective of the
Table 1. Boulder characteristics from moraines at Redfish Lake, and Alturas Lake, Idaho. Units are specified.
Age names are based on previous relative determination (Williams, 1961).

<table>
<thead>
<tr>
<th>Boulder ID</th>
<th>CNEF ID</th>
<th>Age (ka)</th>
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<th>2α</th>
<th>Description/Locality</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Elevation (ft)</th>
<th>Height (m)</th>
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### Late Glacial

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</table>

**Notes**

1. Chemistry was completed at University of Kansas, AMS at Lawrence Livermore National Laboratory, with 1σ precisions of 3% or better, using Kuni Nishiizumi standard KNSTD549 and normal LLNL standards, and assuming a $^{10}\text{Be}$ half life of 1.5 x 10⁶ years.

2. Concentrations were corrected for geochemical and AMS background of $^{10}\text{Be}$ which was 5% or less of the measured atoms for each sample. Precision is the 1σ AMS uncertainty. For the purposes of comparing TCN exposure ages to other chronologies, accuracy at 1σ would be dominated by a ca. 20% estimate for the total random and systematic errors, but does not include the contribution to uncertainty in age from error in inheritance or the effect of erosion.
laboratory preparation was to concentrate quartz from the granites and extract \( {^{10}}\text{Be} \) for AMS dating.

Samples were broken and crushed using BICO-Braun Chipmunk and disc pulverizer machines. Each sample was separated into four size fractions (>500\( \mu \)m, 355-500\( \mu \)m, 250-355\( \mu \)m, <250\( \mu \)m) using # 35, 45, 60 Fisher Scientific sieves. The targeted bulk quantity, from which quartz was separated, was 400 grams of size fraction 355-500\( \mu \)m. When 400 grams of that size was not available, the difference was made up from the 250-355\( \mu \)m size fraction.

**Pretreatment**

Pre-treatment of the samples included dissolving of mafic minerals and non-quartz phases using Aqua Regia, a mixture of 400 ml of 3HCl:1HNO\(_3\) per 400 g of sample. The acids were mixed with the sample in a Teflon beaker and heated on a hotplate at 200\( ^\circ \)C for a minimum of two hours. Samples were then leached in water to remove the dissolved phases in preparation for the next process of hydrofluoric (HF) etching. Several dissolutions in HF acid eliminate the meteoric \( {^{10}}\text{Be} \), which adsorbs to the surface of mineral grains. HF acid ‘etches’ off the outer rind of each quartz grain and also helps to soften and dissolve remaining non-quartz minerals. A mixture of 2HF:1H\(_2\)O was then added to the sample and heated for 20 minutes on a hotplate. The samples were rinsed in de-ionized water again and dried in preparation for ultrasonic quartz extraction.

Ultrasonic quartz extraction uses heated water (near 100\( ^\circ \)C) and ultrasonic vibrations to abrade and partially dissolve the remaining non-quartz phases and to remove any contaminating nuclides. A mixture of 59% HF, 41% HNO\(_3\) and water was added to 60 g of sample in a one-liter bottle. The sample bottles were put in the ultrasonic machine for a
minimum of three periods of 8-10 hours. This process was repeated until all of the non-
quartz phases were dissolved, or until there was <125 ppm aluminum [Al] remaining in the
solid sample. Franz magnetic separation was used when there was a small amount of
magnetic minerals in the sample, and when continuing ultrasonic quartz separation
compromises the sample quantity.

Each sample was tested for aluminum using the Quant-EM Aluminum test kit. To do
this, approximately 1.0000 g of the sample was dissolved in 10 ml of 0.5N HCl acid on a
50°C hotplate. The remaining particle material was dissolved in 5 ml of 0.5N HCl acid and
mixed with the test kit reagent to bring the pH up to 13. A simple colored strip was then
used to determine the concentration of aluminum in the sample. The $^{10}$Be chemistry could be
compromised when more than 125 ppm aluminum remains in the sample. In this case the
sample would have been returned to the ultrasonic quartz extraction phase to help dissolve
the remaining non-quartz, aluminum-bearing minerals.

**Isotope Extraction**

Once quartz concentrates were determined to be acceptably pure, the extraction of
$^{10}$Be from the quartz was completed. This part of the process separates $^{10}$Be from the quartz
and any remaining meteoric isotopes. A portion (30 ± 0.0001 g) of each sample was weighed
and approximately 0.3 ± 0.0001 mL of a Be Carrier was added to each sample (the carrier is
mostly $^9$Be with a known amount of $^{10}$Be, which increases the AMS target mass and prevents
carrying all of the $^{10}$Be through chemistry). A chemical blank is also processed with the
samples. Several acids were added to each sample to dissolve the rock and prevent the
formation of CaF$_2$, ensuring an easier removal of the silica in SiF$_4$ and H$_2$SiO$_4$; these include
120 mL HF, 12 mL HClO$_4$ (perchloric acid), 30mL Aqua Regia per 30 g of sample. The
samples were heated until dissolved on a 125°C hotplate and then evaporated at 200°C. Next, 10 mL of HClO₄ was added to the sample with de-ionized water to wash down the vessel and evaporated again. At this stage, the beryllium is in the form of BeClO₄. Samples were cooled and 15-20 mL HNO₃ was added and again evaporated, this time at 100°C. Finally, the sample was dissolved in 2% HCl acid and centrifuged to remove any insolubles.

Ion chromatography chemistry first removes unwanted elements in ion form from the sample by using an exchange resin that holds certain anions or cations according to pH. This prevents the dilution of Be by other elements so that the AMS target is pure Be oxide. After the sample was run through an anion column, it was run through a cation column to further purify the sample and remove as much Ti as possible. The cation and anion column resins were made of styrene beads with a high ion exchange capacity.

The final step was preparation of the Beryllium sample into about 1 mg of whitish-grey BeO powder oxide for AMS analysis. Quartz vials were boiled in a mixture of de-ionized water, HF and HNO₃ to remove any boron that may be adsorbed to the surface of the vial. Five mL HClO₄ was added to the dried beryllium sample, evaporated and repeated. Next, the sample was dissolved in 8 mL 0.5N HCl and centrifuged. Any insoluble residue is possibly TiO₂. The solution was decanted into a clean test tube and heated in a water bath (50-60°C) until thermal equilibrium was reached. Ammonia gas was then used to precipitate the Be (OH₂) by bringing the solution to a pH of about 9.2. No concentrated acid was added from this point on. The sample was centrifuged, decanted, vortexed with additional de-ionized water and centrifuged again. The precipitate was then put into the quartz vials, into a furnace, and heated at 120°C for a minimum of two hours. Samples were cooled and the sample oxide was scraped from the vial walls. The samples were then put in the furnace
again for a minimum of one hour to convert to BeO at 850°C. Finally, niobium powder was added in a 1:1 niobium to oxide ratio. The samples were analyzed at Lawrence Livermore National Laboratory (LLNL) by Bob Finkel. Total 1σ precisions averaged 3% or better, using Kuni Nishiizumi standard KNSTD549 and normal LLNL standards, assuming \(^{10}\text{Be}\) half life of \(1.5 \times 10^6\) years. Concentrations were corrected for geochemical +AMS background of \(^{10}\text{Be}\) which was 5% or less of the measured atoms for each sample. Precision is the 1σ AMS uncertainty. For the purposes of comparing TCN exposure ages to other chronologies, accuracy at 1σ would be dominated by a ca. 20% estimate for the total random and systematic errors, but does not include the contribution to uncertainty in the age from error in inheritance or the effect of erosion.

IV. Morainal Chronology

**Elk Meadow morainal group**

Description

The Elk Meadow morainal group has been considered Bull Lake since Williams mapped the area in 1961. The morainal group is located on the southeastern side of the Pinedale lateral moraines that now contain Redfish Lake and extends east into the Stanley Basin (Figure 8). Elk Meadow moraines are preserved because subsequent glacial advances and outwash did not override them. Several nested lateral moraines within the group are truncated by the Salmon River moraine group on the proximal end, and the Salmon River on the distal side (Figure 8, also Figure 2). Moraine crests are subdued and 20-25 m wide. Boulders are sparse and show signs of weathering such as rills, gnmmas, spalls and grussy
Figure 8. Elk Meadow morainal group east of Redfish Lake, Idaho. Moraine crests shown in blue. Boulders sampled and $^{10}\text{Be}$ dated for this study are indicated with a red dot, boulders sampled but not yet dated are indicated with a green dot.
texture at their bases (Figure 9). Four boulders have been dated from this moraine complex as a part of this study.

**Chronology**

A sample from the terminal end of a lateral moraine within the Elk Meadow morainal group (ID-RL-04-029) was $^{10}$Be dated at 205.7 ± 4.4 ka (Figure 8). This sample was situated just off the crest of a narrow (20 m wide) moraine. ID-RL-04-031 is on a terminal moraine, approximately 1 mile south of sample ID-RL-04-029, and has been dated at 164 ± 7.3 ka. Two samples from the northernmost terminal portion of Bull Lake moraine date at 108.7 ± 6.4 ka and 107.6 ± 5.8 ka (ID-RL-04-027 and ID-RL-04-028).

Boulders dated were of good quality with minimal gruss or other weathering features, and can be seen in Figure 9. The range of ages for Bull Lake I moraines at Elk Meadow is 107.6 ka to 205.7 ka and they agree well with other dated Bull Lake moraines in the Rocky Mountains (e.g. Richmond, 1986; Gosse et al., 1995b; Phillips et al., 1997; Sharp et al., 2003). Type Bull Lake moraines in the Wind River Range are approximately 150.0 ± 8.3 ka (based on $^{230}$Th/U ages from carbonate clast-rinds which formed on Pleistocene glacio-fluvial terraces; Sharp et al., 2003). Gosse et al. (1995b) obtained an age of 137.8 ± 4.1 $^{10}$Be ka on a Bull Lake moraine in the Wind River Range at Fremont Lake; its age would be much older, 179.2 ± 5.4 $^{10}$Be ka, if an estimated erosion rate of 1.5 μm a year is assumed.

In Yellowstone National Park, Richmond (1986) obtained K-Ar ages on exposed sediments of Flat Mountain Arm at Yellowstone Lake. Two units (moraines) are divided by a pumice bed that has been dated at 225 ka, and the upper unit is overlain by a rhyolite flow that has been dated at 174 ka (Richmond, 1986). Rock Point till, also in Yellowstone contains volcanic deposits and also lies over another pumice bed K-Ar dated at 153 ka (and
Figure 9. Elk Meadow sample boulders. A: ID-RL-04-028, boulder is approximately 2 m tall. B: ID-RL-04-029, boulder is approximately 1.8 m tall. C, D: ID-RL-04-031, boulder is approximately 1.7 m tall, and thickness of sample obtained was 4.5 cm.
most likely younger than 160 ka) (Richmond, 1986). These moraines are from the Illinoian Glaciation and represent Bull Lake equivalent deposits. For this study, the age of samples ID-RL-04-028 through ID-RL-04-031 at Redfish Lake are maximum ages for the Bull Lake I moraines.

The Elk Meadow morainal group $^{10}$Be ages in summary:

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<th>$^{10}$Be age</th>
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<tr>
<td>Elk Meadow terminal</td>
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</tr>
<tr>
<td>ID-RL-04-027</td>
<td>108.7 ± 6.4 ka</td>
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<tr>
<td>ID-RL-04-028</td>
<td>107.6 ± 5.8 ka</td>
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<tr>
<td>ID-RL-04-031</td>
<td>164 ± 7.3 ka</td>
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<tr>
<td>Elk Meadow lateral</td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-029</td>
<td>205.7 ± 4.4 ka</td>
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</table>

*Salmon River morainal group*

Description

The Salmon River morainal group faces to the south parallel to the mountain front (Figure 10, also Figure 2) in contrast to the rest of the northeast-trending moraines at Redfish Lake. The Salmon River morainal group was originally mapped by Williams (1961) as a second, younger, Bull Lake advance. The two sets of bounding lateral moraines are narrow crested (10-15 m) except where they spread out near their terminal end and are better defined than the Elk Meadow moraines. The six nested moraines are also fairly narrow (10-30 m wide).

Boulders on these moraines are primarily Eocene granite and are more abundant than on the Elk Meadow moraines to the northeast. They are also much larger. Of the boulders sampled, Salmon River boulders had an average height of 2.4 m (n=6) and Elk Meadow
Figure 10. Salmon River moraine group, south of Redfish Lake, Idaho. Crests of six recessional moraines and two sets of bounding lateral moraines are shown in blue. Boulders sampled and $^{10}\text{Be}$ dated for this study are indicated with a red dot, boulders sampled but not yet dated are indicated with a green dot, and boulders visited but not suitable for sampling are indicated with a blue X.
boulders were 1.8 m (n=5). Boulders are more abundant on the right lateral group than on the left lateral group, and generally smaller (<1 m) on the end moraines than on the lateral moraines. Many have rills, gnammas, and spalls. Grussification is rare on these boulders.

Samples ID-RL-04-045 and ID-RL-04-047 from one of the nested moraines were collected from a large boulder field (Figure 11). Boulders on this moraine ranged in size from 1.5 to about 6 m$^3$. An ideal sample might be collected from the largest (~6 m$^3$) boulder, but climbing conditions were deemed too dangerous without appropriate equipment. The Salmon River morainal group is cross-cut on its proximal end by the right lateral Redfish Lake Pinedale moraine and is bounded by Decker Creek and the Salmon River at its distal end.

Chronology

A boulder from the boulder field on a recessional end moraine in the middle of the sequence (ID-RL-04-045) was dated at 238.0 ± 4.6 ka. A sample from a proximal position, high on the left lateral moraine was dated at 323.0 ± 18.5 ka (ID-RL-04-043) (Figure 10). The boulders dated were both pink Eocene granite but were significantly weathered (Figure 11). The sample obtained from the nested end moraine, should be younger than the sample from one of the two large bounding lateral moraines, and it is.

The samples obtained from the cross-cutting Salmon River moraines are older than those dated for the Elk Meadow moraines. While many explanations are possible, the most likely is that Elk Meadow moraines suffered a greater degree of weathering, therefore "erasing" some of the beryllium accumulation. In effect, the boulders continued to erode, revealing a lesser amount of $^{10}$Be with depth in the surface of the rock. The weathering
Figure 11. Photos A, B: The boulder field on one of the nested moraines in the Salmon River morainal group, Redfish Lake, Idaho. These boulders are 2 m and 1 m tall, although other boulders ranged in size from 1 m$^3$ to approximately 6 m$^3$. Photo A: ID-RL-04-047, sample currently in processing. Photo B: Some of the other boulders on the crest of this moraine. Photos C, D: ID-RL-04-043 (323 ± 18.5 ka) on the outer Bull Lake Salmon River moraine.
Figure 12. Salmon River and Elk Meadow morainal complex sample boulders. Boulders shown here are Eocene granites. A, B: ID-RL-04-045 from the Salmon River morainal group; boulder is approximately 2.6 m tall, thickness of sample obtained was 1 cm. C, D: ID-RL-04-030 from the Elk Meadow morainal group; boulder is approximately 2 m tall, sample thickness was 1.5 cm (not yet dated).
The differences between a Salmon River boulder and an Elk Meadow boulder are illustrated in Figure 12. The Salmon River morainal group $^{10}$Be ages in summary:

<table>
<thead>
<tr>
<th>Boulder</th>
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<td>Salmon River recessional</td>
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<td>$238.0 \pm 4.6$ ka</td>
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<tr>
<td>Salmon River left lateral</td>
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<td>ID-RL-04-043</td>
<td>$323.0 \pm 18.5$ ka</td>
</tr>
</tbody>
</table>

**Redfish Lake morainal group**

Description

The Redfish Lake morainal group consists of extensive nested, lateral, and end moraines around Redfish Lake (Figure 3, 13). These massive, spectacular moraines were mapped as Pinedale by Williams (1961).

The end moraines are extensive, broad-crested, and display characteristic knob and sink topography. Moraines extend completely across the Stanley Basin and have pushed the Salmon River to the eastern margin of the basin where the river cuts through the distal moraine. Right and left lateral moraines confining Redfish Lake are massive, sharp crested, and scattered with large Eocene granite boulders. These moraines are over 300 m high at the western end of the lake, and they grade to <50 m high at the eastern end near Little Redfish Lake (Figures 2, 13). The crests stay sharp around Redfish Lake while cross-cutting the Salmon River morainal group but become broader and begin to spread out near the southeastern third of the lake where they cut across the Elk Meadow morainal group. Redfish Lake is the only drainage on the eastern side of the Sawtooth Mountains where late Pleistocene ice advances did not extend fully over Bull Lake deposits.
Figure 13. Oblique aerial photo of Pleistocene last glacial maximum (LGM) and nested/recessional moraines surrounding Redfish and Little Redfish Lakes. White Mountain Range in background.
Boulders on the end moraines are abundant, but are generally less than one meter tall with few sufficiently tall to meet our sampling criteria. Boulder surface quality is fair because more Cretaceous granitic boulders occur on these moraines than on the younger moraines. Higher boulder weathering and erosion rates of these boulders may affect the exposure ages measured because less $^{10}$Be may be preserved deeper in the rock, resulting in an apparently younger age. However, several samples were collected from the group which were of ideal size (>1.5 m) and geometry to meet sampling criteria (Figure 14). They were of pink Eocene granite and displayed minor spalling and rilling, only a few showed minor grussification.

Approximately twelve prominent moraines are nested within the Redfish Lake morainal group, and several others that are non-continuous drift deposits. These recessional moraines are only preserved to the east of Little Redfish Lake (Figure 3). On the opposite side of the valley, northwest of Little Redfish Lake, a large lateral moraine about 70 m high extends from Fishhook Creek to the inlet of Little Redfish Lake; this moraine probably prevented recessional moraines from forming here during glacier retreat. Boulders on the recessional moraines are not sparse, but those suitable for sampling were not abundant. Sampled boulders consisted of the pink Eocene granite, and ranged in size from 1.5 to 3.8 m high (Figure 14). Because so many nested moraines occur behind the terminal zone, the Redfish Lake glacier was clearly sensitive to small climatic and precipitation changes.

Chronology

Thirteen dated samples obtained from the Redfish Lake morainal group confirm a clear late Pinedale age. A sample taken from the northern terminal end of one of the older outer Redfish Lake left lateral moraines was dated at 18.5 ± 0.9 ka (ID-RL-04-013) (Figure
Figure 14. Redfish Lake Pinedale LGM and recessional sample boulders. A: ID-RL-04-012, boulder is approximately 1.7 m tall. B: ID-RL-04-024, boulder is approximately 3.8 m tall. C: ID-RL-04-025, boulder is approximately 1.7 m tall. D: ID-RL-04-026, boulder is approximately 1.5 m tall.
3). A sample obtained from the southern terminal end of one of the older Redfish Lake lateral moraines was dated at 15.7 ± 0.8 ka (ID-RL-04-012). No other terminally positioned boulders were suitable for dating purposes. Samples taken from high on the Redfish right lateral moraine were dated at 19.8 ± 0.9 ka (ID-RL-04-038) and 15.9 ± 0.8 ka (ID-RL-04-039) (Figure 10). Another sample, ID-RL-04-040, taken 26 meters in elevation below sample ID-RL-04-039, but on the same moraine, was dated at 17.5 ± 0.8 ka. This boulder is in a swale on the crest of the Pinedale moraine. All three boulders are interpreted to be from the LGM advance since the ice may have been at a maximum thickness in that area for several thousand years. Samples ID-RL-04-012 (15.7 ± 0.8 ka) and ID-RL-04-013 (18.5 ± 0.9 ka) also support this interpretation.

Several suitable boulders were sampled from the multiple recessional moraines at Redfish Lake (Figure 14). Five dates were obtained from these moraines between the terminal zone and Redfish Lake. They range from 14.2 ± 0.7 ka to 15.5 ± 0.7 ka (Table 1, Figure 3). These boulder ages are in notable agreement and succession, and give the time of glacier recession during the late Pleistocene. From these new ages, the climate during the late Pleistocene was clearly fluctuating dramatically and the Redfish Lake glacier was responding to these short lived, extreme transitions between warm and cold temperatures and precipitation patterns.
The Redfish Lake morainal group in summary:

<table>
<thead>
<tr>
<th>Boulder</th>
<th>^{10}Be age</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Redfish terminal</strong></td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-013 (north)</td>
<td>18.5 ± 0.9 ka</td>
</tr>
<tr>
<td>ID-RL-04-012 (south)</td>
<td>15.7 ± 0.8 ka</td>
</tr>
<tr>
<td><strong>Redfish recessional</strong></td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-026 (distal)</td>
<td>15.4 ± 0.7 ka</td>
</tr>
<tr>
<td>ID-RL-04-025</td>
<td>15.5 ± 0.7 ka</td>
</tr>
<tr>
<td>ID-RL-04-024</td>
<td>15.4 ± 0.7 ka</td>
</tr>
<tr>
<td>ID-RL-04-023 (proximal)</td>
<td>14.2 ± 0.7 ka</td>
</tr>
<tr>
<td><strong>Redfish right lateral</strong></td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-038</td>
<td>19.8 ± 0.9 ka</td>
</tr>
<tr>
<td>ID-RL-04-039</td>
<td>15.9 ± 0.8 ka</td>
</tr>
<tr>
<td>ID-RL-04-040</td>
<td>17.5 ± 0.8 ka</td>
</tr>
</tbody>
</table>

**Alturas Lake morainal group**

Description

The Alturas Lake moraine sequence is similar to that at Redfish Lake (Figure 15). Also mapped by Williams in 1961, these moraines are not as spectacular or well preserved as those at Redfish Lake. Boulders on these moraines are sparse because they are comprised of the Cretaceous Idaho batholith granite that did not produce large boulders and they weather more readily than the Eocene granite at Redfish Lake.

Three older moraines in the lower valley are described by Borgert (1999) and Thackray et al. (2004) as broad, voluminous, and gently sloping. Borgert (1999) assigned the names Busterback I, II, and III to these moraines. Upvalley, a sequence of five steep-sided,
Figure 15. Alturas Lake morainal group. The purple overlay designates moraines mapped by Borgert (1999). Busterback II moraine, with crest indicated, is the only moraine in this valley with boulders suitable for cosmogenic dating. The dates obtained from these boulders indicate that the Busterback II moraine was deposited during the late Pleistocene rather than 10 ka earlier.
sharp-crested moraines were named Cabin Creek, Perkins Lake, Alturas Lake, Eureka Gulch, and Alpine Creek (in order from oldest to youngest; Borgert, 1999).

No boulders suitable for dating were located on the five younger sets of moraines. Only two boulders were found on the Busterback II moraine which met the sampling criteria (ID-AL-04-033, and ID-AL-04-034, Figures 15, 16); both were Cretaceous granite.

Chronology

In order to determine if the outermost moraines at Alturas Lake were the same age as the outermost moraines at Redfish Lake (i.e., Pinedale), two samples from the terminal moraine at Alturas Lake were obtained for comparison with dates at Redfish Lake. Ages of 17.7 ± 1.0 ka and 16.8 ± 0.8 ka (ID-AL-04-033 and ID-AL-04-034 respectively) from the Busterback Ranch II indicate that they are indeed Pinedale and correlative with the outermost moraines at Redfish Lake, not mid-Wisconsin (30-60 ka) as implied by Thackray et al. (2004)

The Alturas Lake morainal group $^{10}$Be ages in summary:

<table>
<thead>
<tr>
<th>Boulder</th>
<th>$^{10}$Be age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Alturas Lake terminal</td>
<td></td>
</tr>
<tr>
<td>ID-AL-04-033</td>
<td>17.7 ± 1.0 ka</td>
</tr>
<tr>
<td>ID-AL-04-034</td>
<td>16.8 ± 0.8 ka</td>
</tr>
</tbody>
</table>
Figure 16. Alturas Lake Pinedale sample boulders. A: ID-AL-04-034. B: ID-AL-04-033. Each boulder is approximately 1 m tall.
Figure 17. Oblique aerial photo of Bench Lakes, just north of Redfish Lake, Idaho. Sawtooth Mountain Range in background.
**Bench Lakes morainal group**

Description

Moraines are present around the lower four of the five lakes comprising Bench Lakes north of Redfish Lake (Figure 2). The lower two Bench Lakes moraines and surrounding area were mapped as Pinedale by Williams (1961), but he did not include moraines around lakes Three, Four and Five (Figure 17).

A broad, flat-topped moraine, which is generally free of large (>1m) boulders surrounds Bench Lake One, the lowest lake. This end moraine is about 85 m higher than the Redfish Lake Pinedale left-lateral moraine to the south and is cut at its distal end by the outlet stream for the five lakes (Figure 4). Sample ID-RL-04-015 was obtained from high on the right lateral moraine of this lake.

Bench Lake Two does not have an obvious terminal moraine crest, but two broad lateral moraines are preserved on the south side of the lake. Sample ID-RL-04-014 was taken from among several other large angular boulders on the outermost lateral at Bench Lake Two.

Bench Lake Three is much smaller than the other lakes. The terminal moraine is small (10 m wide) and flat-topped, while the lateral moraines are voluminous and broad-crested. Boulders are abundant on the left lateral moraine around Bench Lake Three, but most were not on the crest and could have moved since deposition. The dated sample (ID-RL-04-020) is from the crest of the 20 m–wide, right lateral moraine. Boulders were abundant here as well, but only a few were >1.5 m.

Bench Lake Four has a rocky terminal moraine approximately 15 m wide (Figures 4, 18). No lateral moraines are present because most of the lake is surrounded by steep
Figure 18. A: Moraine enclosing the Fourth Bench Lake, view east; Redfish Lake is beyond the end moraine. B: ID-RL-04-017 on the Fourth Bench Lake moraine, $^{10}\text{Be}$ dated at 11.4 ± 0.5 ka. C: Boulders not suitable for dating on the Bench Lake Four moraine.
bedrock. Ages from samples ID-RL-04-016, ID-RL-04-017 and ID-RL-04-018 from this moraine date the youngest advance in the Bench Lakes area.

Bench Lake Five is held in by a rock ridge—no moraine was visible during an air reconnaissance.

Chronology

A late Pinedale lateral moraine just outside of the Bench Lakes moraine sequence was dated at 16.0 ± 0.8 ka (ID-RL-04-014). Likewise, ID-RL-04-015 is from a right lateral moraine extending around Bench Lake One, and is dated at 14.7 ± 0.5 ka. The 16 ka date is from a moraine that is nested inside the moraine dated at 14.7 ka. The two dates overlap within 1σ.

The end moraine around Bench Lake Three upvalley was dated at 11.7 ± 0.6 ka (ID-RL-04-020). Dates of 11.7 ± 0.5 ka (ID-RL-04-016), 11.5 ± 0.5 ka (ID-RL-04-018), and 11.4 ± 0.5 ka (ID-RL-04-017) were obtained from the moraine around Bench Lake Four. These ages indicate that the moraines were deposited during the Younger Dryas chronozone (about 11,000 to 10,000 ¹⁴C yrs. B.P.; 12.6 to 11.4 ka; Figure 19). The following sequence of glacial events is suggested:

1) Deposition of a high Pinedale lateral moraine and end moraine around Bench Lake One about 14.7 ± 0.5 ka

2) Retreat of the glacier to Bench Lake Two, stabilization of the terminus to deposit a lateral moraine and end moraine around the lake about 16.0 ± 0.8 ka (this age overlaps that at Bench Lake One).

3) Retreat of the glacier to Bench Lake Three, and stabilization of the terminus to deposit a lateral and end moraine around the lake about 11.7 ± 0.6 ka.
4) Retreat of the glacier to Bench Lake Four, stabilization of the terminus to deposit a lateral and end moraine around the lake about 11.4–11.7 ka.

The Bench Lakes morainal group in summary:

<table>
<thead>
<tr>
<th>Boulder</th>
<th>( ^{10}\text{Be age} )</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pinedale lateral</strong></td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-014</td>
<td>16.0 ± 0.8 ka</td>
</tr>
<tr>
<td><strong>Bench Lake One</strong></td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-015</td>
<td>14.7 ± 0.5 ka</td>
</tr>
<tr>
<td><strong>Bench Lake Three</strong></td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-020</td>
<td>11.7 ± 0.6 ka</td>
</tr>
<tr>
<td><strong>Bench Lake Four</strong></td>
<td></td>
</tr>
<tr>
<td>ID-RL-04-16</td>
<td>11.7 ± 0.5 ka</td>
</tr>
<tr>
<td>ID-RL-04-17</td>
<td>11.4 ± 0.5 ka</td>
</tr>
<tr>
<td>ID-RL-04-18</td>
<td>11.5 ± 0.5 ka</td>
</tr>
</tbody>
</table>
Figure 19. Oxygen-isotope ($\Delta^{18}O$) values from the GISP2 ice core in Greenland on the left. These are compared to cores from Crawford Lake in Ontario Canada on the right. This graph shows the onset of the Younger Dryas at approximately 11,000 $^{14}$C yrs B.P., and its decline just under 10,000 $^{14}$C yrs B.P. From Yu and Wright, 2001.
VI. Correlation

**Bull Lake**

New sample ages from the terminal end of a lateral moraine within the Elk Meadow morainal group range from 107.6 ± 5.8 ka to 205.7 ± 4.4 ka (this study). This moraine represents the earliest advance still intact at Redfish Lake.

Combined cosmogenic $^{36}$Cl and $^{10}$Be ages from boulders on moraines at Bull Lake, Wyoming give a minimum limiting age of 130.0 ka for the Bull Lake moraines (Phillips et al., 1997). Two older $^{10}$Be dates (238 and 323 ka) from the Salmon River morainal group are considerably older than ages from the Bull Lake I Elk Meadow moraines (107 to 205 ka). Because the ages of the Elk Meadow moraines closely match Bull Lake moraines dated elsewhere in the region, inheritance in the Salmon River boulders is most likely.

**Last Glacial Maximum**

Deposits of the LGM in the Western United States are widespread and generally well preserved. In some areas the LGM is recorded at 21 ka and in others the LGM occurred at 17 ka; yet in several places, both phases of glacial advance were preserved. In several areas of Wyoming two separate phases of the LGM have been dated at approximately 24 ka and about 18 ka (Gosse et al., 1995b; Chadwick et al., 1997). In Yellowstone, evidence of both phases of the LGM occurs, one between 29.9 and 22.5 cal ka, and the other between 19.5 and 15.5 cal ka (Sturchio et al., 1994). In the Wallowa Mountains of Oregon, Licciardi et al. (2004) have $^{10}$Be dated moraines at about 21.1 and about 17.0 ka. In the Puget Lowland of Washington and Canada, the Cordilleran Ice Sheet produced the Coquitlam Drift between $25.7 \pm 0.2$ and $22.3 \pm 0.2$ cal ka$^1$ ($^{14}$C dated; Hicock and Armstrong, 1981), and the Vashon
Drift after 22.2 ± 151 ka (recalculated from 18,700 ± 170 14C yrs B.P. \(^1\)) near the Canadian border, and after 18.2 ± 0.53 ka (both of the Fraser Glaciation; recalculated from 15,000 ± 400 14C yrs B.P. \(^1\); Mullineaux et al., 1965, Yount et al., 1980; Easterbrook, 1992). At Redfish Lake, Idaho, the maximum age for the LGM is approximately 18.5 ka and ice from a younger moraine building phase appears to have come close to this maximum position at 15.7 ka (this study).

While several areas have two LGM deposits, others do not. This discrepancy may be attributed to different localized events (i.e., location relative to water bodies, or orographic effects) influencing each area. For example, cosmogenic \(^{36}\)Cl and \(^{10}\)Be ages are 23.0-15.8 ka for boulders on the three Pinedale terminal moraines at Bull Lake, Wyoming (Phillips et. al, 1997). The cosmogenic nuclide age data presented in this study shows clearly the advance and retreat of the Redfish Lake late Pleistocene and YD glaciers, and fills a significant gap in the number of dated LGM deposits in North America.

**Cascade Mountains**

Dated LGM deposits occurring in Western Washington include those on the Olympic Peninsula, in the North Cascades, and in the Puget Lowland. On the Olympic Peninsula, Thackray (2001) obtained a minimum radiocarbon age of 22.9 ± 0.3 ka (recalculated from 19,300 14C yrs B.P. \(^1\)) for the Hoh Oxbow 3 advance, and 22.7 ± 0.3 to 21.8 ± 0.3 ka (recalculated from 19,100 to 18,300 14C yrs B.P. \(^1\)) for the Twin Creeks 1 advance. He also noted the presence of an undated Twin Creeks 2 advance slightly upvalley from the Twin Creeks 1 deposit, which was also likely deposited during the end of, or shortly after the LGM.
North Cascades LGM deposits have been distinguished primarily by moraine morphology and relative weathering characteristics (Porter 1976, 1978; Porter et al., 1983; Waitt et al., 1982) with limited $^{14}$C dates. At Snoqualmie Pass, Porter previously correlated the Domerie member of the Lakedale Drift with the Vashon Drift of the Puget Lobe deposited between 18.2 ka and 16.1 ka (recalculated from 15,000 and 13,500 $^{14}$C yrs B.P. $^1$). Two pre-Domerie Lakedale advances are thought to have occurred during the late Pleistocene as well (Ronald and Bullfrog members) as suggested by reconstructed glacier profiles, although there are no numerical dates from this area (Porter, 1976). North of Snoqualmie Pass, near Glacier Peak, various LGM Cascade glaciers terminated as far as 50 km east of the divide. These also include the more extensive Leavenworth Drift (in the Wenatchee River valley) and the Cordilleran Ice Sheet's (CIS) Okanogan Lobe on the Waterville Plateau that is thought to have reached its maximum position shortly before 15 ka (Porter, 1978; Easterbrook, 1979). Limiting tephra deposits helped to determine relative ages. The maximum LGM advance at Mount Rainier is called Evans creek, but no numeric ages are available (Crandell and Miller, 1974).

Northeastern Oregon Cascade LGM deposits are the Suttle Lake Member of the Cabot Creek Formation and the Waban drift, which have similar maximal positions (Scott, 1977; Carver 1973). In the Wallowa Mountains of Oregon, Licciardi et al. (2004) have $^{10}$Be dated two sets of LGM moraines around Wallowa Lake that suggest at least two successive Pinedale advances in the area. Nine samples with a mean age of $21.1 \pm 0.4$ $^{10}$Be ka date the older advance; and nine samples with a mean age of $17.0 \pm 0.3$ $^{10}$Be ka date the younger advance. Licciardi et al., (2004) show that the older moraine was reoccupied during the younger advance and boulders of both ages are deposited on the right lateral moraine.

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Cordilleran Ice Sheet

During the Fraser Glaciation, the Cordilleran Ice Sheet (CIS) Coquitlam advance in British Columbia produced glacial drift with peat and wood that has been dated at 25.7 ± 0.2 ka to 22.3 ± 0.2 ka (recalculated from 21,700 ± 130 to 18,700 ± 170 $^{14}$C yrs B.P.; Hicock and Armstrong, 1981).

In the Puget Lowland, organic matter dated at 18.2 ± 0.5 ka and 19.3 ± 0.5 ka (recalculated from 15,000 ± 400 and 16,070 ± 600 $^{14}$C yrs B.P.) beneath Vashon Drift in Seattle indicates that occupation by ice did not happen until after approximately 18 ka (Mullineaux et al., 1965). Wood from thin, in situ, peat layers interbedded with outwash beneath Vashon till in the central Puget Lowland has been radiocarbon dated at 18,265, 18,000, 17,350, 17,250, 17,000, and 16,510 $^{14}$C yrs B.P. (Easterbrook, 1979, 1986, 1992, 2003; Deeter, 1979). The $^{14}$C dates from coarse-grained, Canadian-provenance outwash indicate that Vashon ice had reached 115 km south of the Canadian border by the time of deposition of the peat. Vashon ice reached the Seattle area about 15,000 $^{14}$C yrs B.P. (17.6 cal ka), and attained its maximum position by about 14,500 $^{14}$C yrs B.P. (16.9 cal ka) (Porter and Swanson, 1998). The ice sheet stayed at this location for about 100 years before it retreated northward past Seattle about 14,500 $^{14}$C yrs B.P. (16.5 cal ka) (Mullineaux et al., 1965; Porter and Swanson, 1998). The Puget Lobe of the CIS retreated northward to the southern end of Whidbey Island, then floated and disintegrated all the way to the Canadian border at about 13,000 $^{14}$C yrs B.P. (15.3 cal ka) (Easterbrook, 1979, 1992, 2003a,b).

Rocky Mountains

Ages obtained from the Redfish Lake Pinedale moraines show a consistent pattern with Pinedale moraines in other areas. LGM deposits have been numerically dated in the
Wind River Range, Yellowstone National Park, the Colorado Front Range, and the Salmon River Mountain Range near McCall, Idaho (i.e.; Phillips et al., 1997; Gosse et al., 1995b; Chadwick et al., 1997; Licciardi et al., 2001; Benson et al., 2004).

Licciardi et al. (2001) sampled $^{10}$Be in boulders from moraines of the Yellowstone ice cap, which reached its terminal position near Chico Hot Springs between about 18.9 and 14.6 ka. This is a single advance and much younger than suggested by Sturchio et al. (1994). Younger still are eleven measurements from Deckard Flats that reveal the ice cap terminus depositing a moraine between 16.0 and 12.4 ka (Licciardi et al., 2001). Sturchio et al. (1994) calculated U-series data on travertines (hot-spring carbonate deposits) near an outlet glacier that controlled the drainage of the Yellowstone ice cap. Their ages show an LGM advance from about 29.9-22.5 ka, followed by a significant retreat from 22.5-19.5 ka, and a smaller readvance after 19.5 ka that ended by 15.5 ka (the Deckard Flats advance).

Several $^{36}$Cl and $^{10}$Be ages were obtained from Pinedale boulders in the Front Range and Park Ranges of Colorado by Benson et al. (2004). Their work in north-central Colorado supports a slightly older LGM advance than at Redfish Lake. Error in their ages is based on $^{36}$Cl leakage, sediment shielding, and to a lesser extent, snow shielding. Corrections were made on the $^{36}$Cl dated samples, resulting in an age increase of approximately 2,200 years. Uncorrected $^{36}$Cl ages range from 20.9 to 16.5 ka, and $^{10}$Be ages from some of the same boulders range from 23.2 to 18.4 ka. Finally, Madole (1986) suggests that glaciers in the Colorado Front Range near the location of Glacial Lake Devlin advanced between 23.5-21 cal ka, based on radiocarbon dating.

The type locality for Pinedale moraines is at Fremont Lake, Wyoming (Blackwelder, 1915), where Gosse et al., (1995b) have found a mean age for the Pinedale maximum of 18.2
± 0.9 $^{10}$Be ka. Well-developed lateral and recessional moraines nestled behind the terminal moraine at Fremont Lake are similar to those at Redfish Lake, Idaho. The terminal moraine is broad crested, and boulder ages span 6000 years (with older boulders located on the distal side of the moraine crest, and younger boulders on the proximal side). Gosse et al. (1995b) infer that boulders were dropped from the ice while it was building the moraine inward. The deposition of all Pinedale moraines at Fremont Lake occurred between 22.0-16.0 ka based on cosmogenic dates from boulders on moraines (Figure 20; Gosse et al., 1995b). Again, these boulder ages are very similar to those at Redfish Lake, and may suggest simultaneous alpine glacier advances.

A complex of five groups of terminal, recessional, and lateral moraines at Sinks Canyon in the Wind River Range are shown to be Pinedale by relative age comparisons (Dahms, 2004). The youngest boulders on the Pinedale recessional moraines at Sinks Canyon have also been $^{10}$Be dated at 17.5 ± 1.0 ka and 15.4 ± 1.0 ka (Fabel et al., 2004). At Bull Lake, Wyoming, Phillips et al. $^{36}$Cl/$^{10}$Be dated boulders on Pinedale moraines at 23.0-16.0 ka (1997). Chadwick et al. found no (Wisconsin age) moraines older than about 23.0 ka in the Dinwoody Lakes area nearby (1997).

Colman and Pierce (1981, 1986) used weathering rinds to estimate ages of moraines at McCall, Idaho (Salmon River Mountain Range) and found advances at about 20.0 ka and about 14.0 ka. The glacial sequence consists of deposits of at least four glaciations; rind measurements on volcanic clasts and soil data allowed the authors to assign estimated numerical ages to the Pinedale and Bull Lake sequences (Colman and Pierce, 1981). They have subdivided the Pinedale till into two units: younger Pilgrim Cove, and older McCall till. Colman and Pierce suggest that the rind thickness for these boulders correspond to ages
Figure 20. Moraines in the Fremont Lake basin, Wyoming. Moraines 1 (oldest) through 7 (youngest) are Pinedale, with a mean $^{10}$Be age of $18.2 \pm 0.9$ ka. Moraines i (oldest) through v (youngest) are Bull Lake, $179.2 \pm 5.4$ $^{10}$Be ka (assuming minimal erosion). From Gosse et al., 1995b.
between 10-25 ka, with Pilgrim Cove being about 14 ka, and the McCall deposits about 20 ka (Colman and Pierce, 1986).

In all of the areas discussed above (Idaho, Wyoming, Colorado, Montana, Puget Lowland, Washington, and Oregon), numerically dated deposits suggest glacial advances that span the time period between 15 ka and 20 ka. LGM and late-Glacial advances at Redfish Lake also coincide with the dated glacial deposits in these areas, suggesting that a majority of large alpine glaciers were advancing and retreating in response to climate changes that were occurring on a regional scale.

**Younger Dryas**

Based on ice core dating in Greenland (GISP2), the Younger Dryas (YD) lasted for approximately 1200 years, from 12.7 to 11.5 ± 0.1 cal ka (11,500 to 10,200 ^14^C yrs. B.P.; Alley et. al, 1993; Stuiver et al., 1995; Meese et al., 1997; Alley, 2000). The onset and decline of the YD cold phase occurred abruptly, changing about 8° C at the 20-40 yr onset and about 12° C in 20 yrs at the closing of the YD.

Ice accumulation is lower during cold periods and higher during warmer periods on the Greenland ice sheet (Alley et al., 1993). In Greenland, ice accumulation rates doubled within a few years at the end of the YD, and eventually came to their modern values of about 0.24 m ice/yr (Alley, 2000). Colder, drier times generally produce dustier ice; Mayewski et al. (1993) noted that increased fluxes of crustal materials imply either more intense circulation over continental regions, increased aridity, or changes in source area. Crustal material fluxes in the GISP2 ice core in Greenland rose sharply at the onset of the YD and declined towards the end of the YD (Mayewski et al., 1993).
Regionally, evidence for the YD ice advance has been noted in western North America in the Cascade Mountains of Washington, Oregon, and southwestern British Columbia, the Fraser lowland, the Rocky Mountains of Colorado, Wyoming, and Idaho, California, and New Mexico, among other places.

**Canadian advances**

Radiocarbon dates of $13.1 \pm 0.2$ ka to $11.8 \pm 0.7$ ka (recalculated from 11,330 to 10,100 $^{14}$C yrs B.P. $^1$) from the Canadian Rocky Mountains (Reasoner et al., 1994) are synchronous with the European YD advance. Herb-dominated pollen abundance in sediment cores is used as a cool climate indicator along coastal British Columbia, at Cape Ball on the Queen Charlotte Islands. Pollen-climate correlations from a peat bed indicate a drop in paleotemperature between $13.0 \pm 0.2$ ka and $11.8 \pm 0.2$ ka (recalculated from 11,100 and 10,200 $^{14}$C yrs B.P. $^1$), which may be correlative with the YD chronozone (Figure 21; Mathewes et al., 1993; Mathewes, 1993). At Crawford Lake, southern Canada, multiple climatic events (Bølling warming, intra-Allerød cold period, Younger Dryas, Pre-boreal oscillation and the 8.2 ka cooling) are recorded by $^{14}$C and $\delta^{18}$O in lacustrine sediments (Yu and Eicher, 2001). Crowfoot Lake sedimentary records in British Columbia indicate increased clastic deposition between $13.1$ ka and $11.8$ ka (recalculated from 11,300 and 10,100 $^{14}$C yrs B.P. $^1$; Reasoner et al., 1994).

**Washington advances**

**North Cascades**

After the CIS disappeared from the North Cascades and long alpine glaciers in the Nooksack Middle Fork retreated upvalley from their terminal positions in the lower valley, the Deming glacier stabilized in the upper valley and built a lateral moraine that incorporated...
Figure 21. This pollen diagram from Cape Ball, Queen Charlotte Islands, shows an increase in herbs and a decrease in arboreal pollen between 10,200 and 10,730 \(^{14}C\) yrs B.P. (12.7 ± 0.05 ka and 11.8 ± 0.2 ka)\(^1\). This suggests a forest perturbation, and cooler, wetter climatic conditions. From Mathewes, 1993.
Figure 22. Location of peat bog, and sample date of 11,050 +- 50 ¹⁴C yrs B.P. at Snoqualmie Pass, Washington. Tephra layers Yn (ca. 3400 ¹⁴C yrs B.P.) and Wn (ca. 450 ¹⁴C yrs B.P.) are from Mount St. Helens. From Porter, 1978.
numerous logs. Calibrated dates of 12.75 ± 0.06 and 12.52 ± 0.14 (recalculated from 10,680 ± 70 and 10,500 ± 70 \(^{14}\)C yrs B.P.\(^1\)) were obtained from the logs, indicating correlation with the YD (Kovanen and Easterbrook, 1996, 2001; Kovanen and Slaymaker, 2005).

Studies by Porter (1976, 1978) concentrated on determining the relative ages of moraines on Snoqualmie Pass, including the Hyak member of the Lakedale Drift. The Hyak member includes the youngest and highest moraines deposited during the late Wisconsin Fraser Glaciation. These moraines were formed by valley and cirque glaciers that extended downvalley at least 3.2 km from the top of the pass (975 m, 3200 ft) and Hyak (820 m, 2700 ft) (Figure 22). The Hyak drift unit is composed of semi-stratified flow till, laminated lacustrine sediments, and fluviatile sand and gravel (Porter, 1976). Because the morainal sequence at Snoqualmie Pass lies upvalley from the maximum advance moraine, Porter (1976) considered these deposits post or late-LGM. He reported a minimum age for the Hyak moraines of 12.9 ± 0.05 ka (recalculated from 11,050 ± 50 \(^{14}\)C yrs B.P.) from wood in late glacial gravel behind a type Hyak moraine (Figure 17). The Hyak moraine complex may be YD correlative, however, given the \(^{14}\)C age and the position of another moraine upvalley from the sample location.

In addition to alpine glaciers, the Cordilleran Ice Sheet responded to the YD climatic signal in Washington. Oscillations of the Cordilleran Ice Sheet (CIS) in the Fraser Lowland produced YD equivalent moraines: the Sumas III stade from 13.4 ± 0.1 to 12.9 ± 0.1 ka (recalculated from 10,980 to 10,250 \(^{14}\)C yrs B.P.\(^1\)), and the Sumas IV stade from >12.9 ± 0.2 to 11.1 ± 0.1 ka (recalculated from >10,250 to ca. 10,000 \(^{14}\)C yrs B.P.\(^1\); Figure 23; Easterbrook and Kovanen, 1998; Kovanen and Easterbrook, 2002; Kovanen, 2002). The
Figure 23. Oscillations of the Cordilleran Ice Sheet in the Fraser Lowland produced moraines correlative to the YD. Sumas III: 13.49 to 12.98 cal ka (10,980 to 10,250 $^{14}$C yr B.P.). Sumas IV: >11.95 to 11.18 cal ka (10,250 to ca. 10,000 $^{14}$C yr B.P.). Lines indicate approximate ice margin. From Kovanen and Easterbrook, 2002.
Sumas readvances correlate with late Pleistocene sea surfaces in the Pacific Ocean (Kienast and McKay, 2001) and the Greenland GISP2 ice core.

**Colorado advances**

Cirque glaciation up valley from Sky Pond in the Colorado Front Range deposited glacio-lacustrine sediments, which were used for $^{14}$C dating by Menounos and Reasoner (1997). A moraine in the lake (altitude 3360 m) is constrained by a minimum limiting core date of 13.8 cal ka (recalculated from 12,040 ± 60 $^{14}$C yrs B.P. $^1$), and shows that overlying clayey-silts indicative of glacial activity, dated at 12.9 ka (base of layer; recalculated from 11,070 ± 50 $^{14}$C yrs B.P. $^1$) and 11.2 ka (top of layer; recalculated from 9970 ± 80 $^{14}$C yrs B.P. $^1$), were deposited during the YD chron (Menounos and Reasoner, 1997).

**Wyoming advances**

In the Wind River Range of Wyoming, Gosse et al. (1995a; 1999) used cosmogenic nuclides ($^{10}$Be, $^{26}$Al) to date Younger Dryas moraines (Titcomb Basin). Eleven boulders from the Titcomb Lakes moraines near Fremont Lake have a mean age of 11.7 ± 0.6 ka. Near Temple Lake, Wyoming, an age of 13.8 ± 0.8 ka (recalculated from 11,770 ± 710 $^{14}$C yrs B.P. $^1$) for a YD correlative advance at Rapid Lake outside the type Temple Lake moraine may be too old based on “old carbon” in the bulk lake sediments (Davis et al., 1998; Zielinski and Davis, 1987). An equilibrium line altitude (ELA) reconstruction for the Titcomb Basin and Temple Lake moraines are similar, and given their proximity to each other, this may justify the fact that the Temple Lake 13.8 ± 0.8 ka age may be too old.

**Oregon, Montana, California, and New Mexico advances**

In the Wallowa Mountains, Oregon, Licciardi et al. (2004) have also $^{10}$Be-dated four granodiorite boulders on a moraine at Glacier Lake (mean = 10.2 ± 0.6 ka) and cirque
bedrock (11.0 ± 0.5 ka) as possibly late YD in age (based on a limited number of samples). This location is approximately 25 km upvalley from the Wallowa Lake Pinedale age moraines.

Several probable Younger Dryas correlative moraines occur in Montana in the Beartooth Mountains (Graf, 1971) and the Tobacco Root Range (Hall and Heiny, 1983) based on position and relative weathering, but they have not been numerically dated. Younger Dryas moraines have been $^{10}$Be dated in the San Bernardino Mountains of southern California by Owen et al. (2003) at between 10.3 ± 0.3 to 13.9 ± 0.8 ka. Carbon-14 dating of shells from Pluvial Lake Estancia in central New Mexico reveals a highstand, which may coincide with the YD cold period (Anderson et al., 2002).

VII. Discussion and Conclusions

Twenty-three new cosmogenic $^{10}$Be ages have been obtained from granitic boulders at Redfish Lake, Idaho. These new ages constrain the glacial sequence at Redfish Lake, and help in identifying a regional pattern of glaciation during the late Pleistocene. The following correlations help clarify how the dates obtained at Redfish Lake fit into the existing concepts of glaciation in the Rocky Mountains and throughout western North America.

**Bull Lake**

Hall and Shroba (1993) suggest that the depth of erosion of Bull Lake moraines in Wyoming could be up to two meters and that any boulder smaller than that could have been covered by soil when initially deposited and later exhumed, making the apparent age younger. However, no evidence of this magnitude of erosion was found at Redfish Lake, where some newly obtained $^{10}$Be ages of Bull Lake moraines appear to be too old, rather than
too young. The heights of sampled Bull Lake boulders are plotted against their apparent ages in Table 2. No clear correlation between boulder height and age is apparent; reinforcing that boulder exhumation is not an issue at Redfish Lake.

<table>
<thead>
<tr>
<th>Height (m)</th>
<th>Age $^{10}$Be ka</th>
<th>Locality</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>107.6</td>
<td>Elk Meadow (EM)</td>
</tr>
<tr>
<td>1.3</td>
<td>108.7</td>
<td>Elk Meadow</td>
</tr>
<tr>
<td>1.7</td>
<td>164</td>
<td>Elk Meadow</td>
</tr>
<tr>
<td>1.8</td>
<td>206</td>
<td>Elk Meadow</td>
</tr>
<tr>
<td>2.6</td>
<td>238</td>
<td>Salmon River (SR)</td>
</tr>
<tr>
<td>2</td>
<td>323</td>
<td>Salmon River</td>
</tr>
</tbody>
</table>

The right lateral Redfish Lake Pinedale moraine cuts across the Salmon River morainal group, which cuts across the Elk Meadow moraine group. The Salmon River moraines are clearly older than the Pinedale moraines and younger than the Elk Meadow moraines. However, two ages from a nested moraine within the Salmon River group are considerably older (238 and 323 ka) than ages from the older Elk Meadow moraines (107 to 205 ka). The reason for this is not readily apparent but could be due to inherited $^{10}$Be in the Salmon River boulders. Boulder erosion to make the Elk Meadow dates too young is also possible, but less likely in light of ages of Bull Lake moraines elsewhere.

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Bull Lake moraines at Redfish Lake are significantly older than those dated through other methods in different localities. Dates of 206 ka, 298 ka and 323 ka are older than the 100–30 ka boulders in Wyoming (based on $^{36}$Cl measurements at Bull Lake; Phillips et al., 1997). These new dates from Redfish Lake are probably still accurate since cosmogenic nuclide dating can detect $^{10}$Be up to $10^6$ years old (Noller, 1995). Cosmogenic $^{36}$Cl however, is accurate only up to about $10^5$ years (Noller, 1995).

**Pinedale**

Ice cores from Greenland and Antarctica have been used to identify past climatic changes by analyzing the isotopic concentration of $^{16}$O, $^{18}$O, $^1$H, and $^2$H (e.g., Dansgaard, 1961; Dansgaard et al., 1989; Stuiver et al., 1995; Alley, 2000). Low $\delta^{18}$O and $\delta$D (deuterium) values measured in glacial ice correlate to lower paleotemperatures (when compared to the same values measured in standard ocean water), and therefore infer a glacial period. The $\delta^{18}$O for glacial periods is usually around $-45\,^\circ$/oo compared to $-29\,^\circ$/oo for non-glacial times. According to annual ice layer counting and the $O^{18}/O^{16}$ ratio in GISP and GISP2 ice cores in Greenland, the LGM occurred sometime between 17 ka and 23 ka (Dansgaard et al., 1989, 1993; Johnsen et al., 1992). Cosmogenic nuclide exposure age data is an increasingly useful way to date deposits for comparison to the ice core standards.

Strong moraine development during two Pinedale advances is found in Wyoming, Idaho, and Oregon (Sturchio et al., 1994; Gosse et al., 1995a, b; Chadwick et al., 1997; Licciardi et al., 2004). The earlier advance gives minimum ages of approximately 18 to 24 cal ka, then recessional moraines dating 18 to 16 cal ka. The Redfish Lake moraines dated in this study seem to show a similar pattern. A boulder located on the outer Redfish LGM left
lateral moraine is $^{10}$Be dated at 18.5 ± 0.9 ka (ID-RL-04-013), and one on the right lateral moraine is 19.8 ± 0.9 ka (ID-RL-04-038), and their ages suggest an advance contemporaneous with the maximum advances of the other areas mentioned above. The age of 15.7 ± 0.8 ka (ID-RL-04-012) on the end moraine near the outlet of Little Redfish Lake suggests that the second period of moraine building was nearly of the same magnitude as the first advance. The absence of more boulders on the terminal moraines makes acquiring additional $^{10}$Be dates difficult.

Ages of recessional moraines are younger with distance upvalley from the terminal moraine of the LGM. Within 2.25 km of the terminal moraine, boulder ages of 15.4 ± 0.7 ka, 15.5 ± 0.7 ka, 15.4 ± 0.7 ka, and 14.2 ± 0.7 ka (ID-RL-04-026, ID-RL-04-025, ID-RL-04-024, ID-RL-04-023 respectively) suggests a sequence of successive brief stillstands in response to the changing climate. At least twelve definitive moraines occur between the terminal moraine and the outlet of Redfish Lake, suggesting that the climate may have been quite variable during the transformation to the Holocene.

Three ages from the right lateral Pinedale moraine at Redfish Lake date at 19.8 ± 0.9 ka, 15.9 ± 0.8 ka, and 17.5 ± 0.8 ka (ID-RL-04-038, ID-RL-04-039, and ID-RL-04-040 respectively). ID-RL-04-038 and ID-RL-04-040 were located on the moraine crest, and ID-RL-04-039 was located in a swale of that same crest. These ages appear to correspond with both the early and late maximum ice stands at Redfish Lake.

Boulders from the same moraine may not show the same ages, not only because of error, but because they were deposited at different times. This was shown by Gosse et al. (1995b) on the Pinedale moraines near Pinedale, Wyoming, where boulders on the front of the crest of the terminal moraine were deposited 6,000 years before the boulders on the inner
Lundeen (2001) suggested that the three Busterback Ranch moraines at Alturas Lake to the south are older than 25 ka based on relative weathering data, and Thackray et al. (2004) suggest an age of at least 27 ka for these three moraines. These age conclusions were based on depth to B soil horizon, moraine crest angularity, and other distinct differences in the moraine morphology between the Busterback Ranch moraines and the Perkins Lake moraines upvalley. The new $^{10}$Be dates obtained at Alturas Lake (this study) are significant because the age of the Busterback Ranch moraines (16.8 and 17.7 ka; avg.17.2 ka) demonstrates that the interpretation by Thackray et al. (2004) that these moraines are 10 ka older than the younger inner moraines is incorrect and that they are Pinedale, as originally mapped by Williams in 1961. These ages also correspond well with the LGM moraine at Redfish Lake. Unfortunately, the absence of more large boulders suitable for dating on these moraines at Alturas Lake makes further constraint of the age of other moraines with cosmogenic nuclides difficult.

The LGM, nested moraines and YD equivalent age of moraines at Redfish Lake, Idaho are similar in age to other advances in the region mentioned previously (Table 3) in the Wallowa Mtns., Oregon (Licciardi et al., 2004); for the CIS in the Puget Lowland, and Puget Lobe deposits (Porter, 1976, 1978; Waitt et al. 1982; Porter et al., 1983; Easterbrook, 1986, 1992); the Olympic Mtns., Washington (Thackray, 2001); Cape Ball, Canada (Mathewes, 1993); Crowfoot Lake, Canada (Reasoner et al., 1994); CIS Sumas III moraine in the Fraser
Lowland (Easterbrook and Kovanen, 1998, 2002; Kovanen, 2002); and at Glacier Lake in the Wallowa Mtns., of Oregon (Licciardi et al., 2004).

Table 3. Comparison of regional ages of LGM and YD moraines.

<table>
<thead>
<tr>
<th>State/Country</th>
<th>Mountain Range</th>
<th>Location</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oregon LGM</td>
<td>Wallowa Mts.</td>
<td>Wallowa Lake</td>
<td>21.0-17.0 $^{10}$Be ka</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>22.2-18.2 ka and</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>18.2-16.5 ka$^1$</td>
</tr>
<tr>
<td>Washington LGM</td>
<td>Puget Lowland</td>
<td>CIS</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Olympic Mts.</td>
<td>Hoh River</td>
<td>22.9-21.7 ka$^1$</td>
</tr>
<tr>
<td>Wyoming LGM</td>
<td>Wind River Range</td>
<td>Fremont Lake</td>
<td>21.7-14.4 $^{10}$Be ka</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Bull Lake</td>
<td>23.0-16.0 $^{10}$Be ka</td>
</tr>
<tr>
<td>Montana LGM</td>
<td>Yellowstone</td>
<td>Deckard Flats</td>
<td>29.9-22.5 ka and</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>19.5-15.5 ka (U-series)</td>
</tr>
<tr>
<td>Colorado LGM</td>
<td>Front Range</td>
<td>Roaring Fork drainage</td>
<td>23.2-18.4 $^{10}$Be ka</td>
</tr>
<tr>
<td>Idaho LGM</td>
<td>Salmon River Mts.</td>
<td>McCall</td>
<td>20.0 and 14.0 ka$^1$</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>(rim thickness)</td>
</tr>
<tr>
<td>NEW! Idaho LGM</td>
<td>Sawtooth Mts.</td>
<td>Redfish Lake</td>
<td>18.5-15.7 $^{10}$Be ka</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>and 15.5 to 14.2 $^{10}$Be ka</td>
</tr>
<tr>
<td>Wyoming YD</td>
<td>Wind River Range</td>
<td>Titcomb Lake</td>
<td>12.8-11.5 $^{10}$Be ka</td>
</tr>
<tr>
<td></td>
<td>Coast Range</td>
<td>Cape Ball</td>
<td>13.0-11.8 ka$^1$</td>
</tr>
<tr>
<td>Canada YD</td>
<td>Rocky Mountains</td>
<td>Crowfoot Lake</td>
<td>13.1-11.8 ka$^1$</td>
</tr>
<tr>
<td></td>
<td>Fraser Lowland</td>
<td>CIS</td>
<td>12.9-11.1 ka$^1$</td>
</tr>
<tr>
<td>Oregon YD</td>
<td>Wallowa Mts.</td>
<td>Glacier Lake</td>
<td>10.2-11.0 $^{10}$Be ka</td>
</tr>
<tr>
<td>Washington YD</td>
<td>North Cascades</td>
<td>Deming</td>
<td>12.5 ka$^1$</td>
</tr>
<tr>
<td>California YD</td>
<td>San Bernardino Mts.</td>
<td>San Gorgonio Mtn.</td>
<td>10.3-13.9 $^{10}$Be ka</td>
</tr>
<tr>
<td>Colorado YD</td>
<td>Front Range</td>
<td>Sky Pond</td>
<td>12.9-11.2 ka$^1$</td>
</tr>
<tr>
<td>NEW! Idaho YD</td>
<td>Sawtooth Mts.</td>
<td>Bench Lakes</td>
<td>11.4-11.7 $^{10}$Be ka</td>
</tr>
</tbody>
</table>

The Pinedale LGM at Redfish Lake is now $^{10}$Be dated at between 18.5 ± 0.9 ka and 15.7 ± 0.8 ka and recessional moraines were deposited until at least 14.2 ± 0.7 ka (this study).
Two Pinedale moraine-building episodes apparently occurred in this area, one older than 18 ka and one younger, as in the Wallowa Mountains. Moraines in the Redfish Lake drainage of the Sawtooth Mountains recorded glacier fluctuation with great sensitivity, as indicated by as many as twelve moraines behind the Pinedale maximum. Precipitation shortages during the LGM are not apparent at Redfish Lake, nor do they appear to be at Alturas Lake, since moraines do record a moraine-building episode around 16.0-18.0 ka (this study).

The ages of 16.8 and 17.7 ka from the Alturas Lake Busterback II moraine also correspond well with the Pinedale LGM moraines at Redfish Lake, suggesting the same climatic conditions for the entire Salmon River Valley. This new data may suggest climate forcing during the Pleistocene (and perhaps other glacial periods as well) which happened simultaneously over the Western United States.

A definite synchronicity can be noted between moraines in the Sawtooth Range and those in Wyoming, Montana, Washington, California, and Oregon. Perhaps the most similar Pinedale deposits to Redfish Lake are those at Fremont Lake and Bull Lake, Wyoming, where $^{10}$Be ages are nearly identical. These areas are separated by only 435 Km (270 miles), and are the two closest localities of similar, well-documented advances in North America.

**Late-GLacial**

The Bench Lakes are located above the outermost Redfish Lake LGM lateral moraines. Lake elevations are between about 2360 m (7750 ft) at Bench Lake One, and about 2495 m (8190 ft) at Bench Lake Four. Bench Lake Five elevation is about 2630 m (8620 ft), and is held in by a rock ledge. Alpine moraine elevations that have been associated in other studies with a YD advance generally occur within a few kilometers from the glacial
cirque. At Bench Lakes, the successively rising moraine and lake elevations and proximity to the cirque suggest that the equilibrium line altitudes were also rising towards the end of the Pleistocene. This was due to the relatively short accumulation periods, to which the glacier was responding. ELAs calculated for this study at Bench Lakes are summarized in Table 4, and range from 2560 m at Bench Lakes One to 2682 m at Bench Lakes Four (Figure 26a-d).

$^{10}$Be dates from the Bench Lakes demonstrate that they are YD correlative (Table 1). The ages of boulders rise successively from Bench Lake One to Bench Lake Four and are between 11.7 and 11.4 ka. Equilibrium line altitudes (ELAs) also rise successively and are between 2560 m and 2680 m for all four of the lake moraines (Table 4); this, paired with the $^{10}$Be dates, support YD equivalent advances (GISP ages: 11,500 to 10,200 $^{14}$C yrs. B.P.; 12.7 to 11.5 cal ka; Alley et al., 1993; Stuiver et al., 1995; Alley, 2000).

Ridges upvalley from Redfish Lake that were interpreted as moraines from air photos were examined from the air and on the ground. No moraines were observed; ridges were glacially scoured bedrock cirques.

Younger Dryas advances occur at Bench Lakes, but not at Redfish Lake; an aerial reconnaissance revealed no apparent moraines in cirques at the headwaters of the Redfish Lake valley. YD correlative moraines could be on the bottom of Redfish Lake; however ELA considerations make this unlikely. Alternatively, the bedrock ridges in cirques may mark the margins of relatively small cirque glaciers equivalent to YD moraines in other cirques. Other post-Pinedale glaciation in the area is also indicated by some other small north and northeast facing cirque moraines above 9,000 ft (Williams, 1961).

The Bench Lakes morainal sequence (11.7-11.4 ± 0.5 ka) at Redfish Lake suggests a YD correlative age for these moraines and correlates well with other areas in western North
America. This relationship could become clearer with additional samples from boulders on these moraines. This paper reinforces abrupt cooling on a regional scale, and supports findings from Washington, Oregon, California, and Canada (Wilson et al., 1993; Reasoner et al., 1994; Friele and Clague, 2002; Kovanen and Easterbrook, 2002; Kovanen, 2002; Licciardi et al., 2004), and the Rocky Mountains (Gosse et al, 1995a, 1995b; Menounos and Reasoner, 1997; Davis et al., 1998; Gosse et al., 1999) that indicate an abrupt cooling event in western North America during the Younger Dryas chron. Further comparison to glacial deposition in other areas of northern North America will be crucial to the timing and occurrence of both the LGM and late-glacial events regionally, and lend in determining the cause of Pleistocene cooling.

\[\text{For ease of comparison in this study, ages have been converted to cal ka (reported with } 2\sigma \text{ confidence) through the CALIB 5.0.2 age calibration program of Stuiver, Reimer and Reimer, 2005.}\]
Table 4. Lake, Moraine, and Equilibrium Line Altitude elevations for Bench Lakes, Idaho.

<table>
<thead>
<tr>
<th>Lake Name</th>
<th>Lake Elevation (ft)</th>
<th>Lake Elevation (m)</th>
<th>Moraine Elevation (ft)</th>
<th>Moraine Elevation (m)</th>
<th>Accumulation Area Ratio (AAR) ELA (ft)</th>
<th>Accumulation Area Ratio (AAR) ELA (m)</th>
<th>Toe to Headwall Altitude Ratio (THAR) ELA (ft)</th>
<th>Toe to Headwall Altitude Ratio (THAR) ELA (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Bench Lake 1</td>
<td>7760</td>
<td>2366</td>
<td>7720</td>
<td>2354</td>
<td>8400</td>
<td>2560</td>
<td>8390</td>
<td>2558</td>
</tr>
<tr>
<td>Bench Lake 2</td>
<td>7760</td>
<td>2366</td>
<td>7760</td>
<td>2366</td>
<td>8560</td>
<td>2609</td>
<td>8420</td>
<td>2567</td>
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<tr>
<td>Bench Lake 3</td>
<td>7860</td>
<td>2396</td>
<td>7860</td>
<td>2396</td>
<td>8720</td>
<td>2658</td>
<td>8510</td>
<td>2595</td>
</tr>
<tr>
<td>Bench Lake 4</td>
<td>8200</td>
<td>2500</td>
<td>8200</td>
<td>2500</td>
<td>8800</td>
<td>2682</td>
<td>8680</td>
<td>2646</td>
</tr>
<tr>
<td>Bench Lake 5</td>
<td>8600</td>
<td>2622</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
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Appendix A: *ELA estimates*

Variations in equilibrium line altitude (ELA) on a particular glacier are used as an indicator of proximal climate fluctuations. The ELA is a theoretical line that separates the zones of accumulation and ablation in a glacier. It is also a gauge of regional climate behavior because glacier mass balance fluctuations can be correlated over distances of about 500 km (Benn and Evans, 1998). In areas of lower precipitation, ELAs will be at higher elevations, and in areas of high precipitation, ELAs will be lower because temperatures needed to melt the greater amount of snow have to be higher. The elevation and aspect of the glacier surface are also important in determining an ELA. Four ways are generally employed to estimate a former glacier's ELA: (1) determined by the elevations of lateral and terminal moraines; (2) toe to headwall altitude ratio (THAR) or median altitude method; (3) cirque floor elevation (CIR); and (4) accumulation-area ratio (AAR) method. Moraine elevations and the AAR methods were used to reconstruct the ELA for the Snoqualmie Pass and the Bench Lakes areas. The lateral moraine ELA estimation is practical because lateral moraines are only deposited below the ELA, however, there should be supporting evidence such as flow direction indicators (striae, drumlins) to make this method accurate. The AAR method, which does not accommodate a wide variety of cirque floor shapes, is mostly used in uncomplicated topographical areas.

ELAs were estimated for the four Bench Lakes, Sawtooth Mountains, Idaho by using the lateral and terminal moraine elevations, THAR, and AAR methods. Glacier ablation limits were based upon the height of the lateral and terminal moraines; from this, the AAR and THAR were used to recreate the accumulation area of the glacier. The results are shown in Figure 24a-d and Table 4.
The hypsometries of the Bench Lakes accumulation and ablation areas are somewhat complicated, and the ELA reproductions vary with each method. Equilibrium line altitudes calculated by the AAR method (accumulation area is 65% of glacier area) rise successively from Bench Lake One to Four and range from 8400 ft to 8800 ft (2560 m to 2680 m). The results of the median altitude method are similar to that of the AAR method, with median altitude ELAs slightly below the AAR ELAs. To calculate the THAR, the elevation difference between the toe and headwall is multiplied by 40% then added to the toe elevation, resulting in an ablation area which is 40% of the glacier area. The THAR method generally underestimates ELAs, and it is common to use another method of ELA estimation as well (Benn and Evans, 1998). A test was done on the proposed glacial advances by calculating the basal shear stress in the ablation zones of the glaciers. In all cases, the basal shear stresses did not conform to either a temperate valley glacier with basal sliding (50-100 kPa), or cold based glaciers that have little or no basal sliding (100-150 kPa). Generally, the basal shear stresses were lower than 50 kPa, which indicates a cessation of ice flow. The discrepancy in basal shear stress calculations for the Bench Lakes study areas could be due to many factors including: (1) complex glacial hypsometries, (2) misinterpretation of glacial ice limits, or (3) misinterpretation of the thickness and surface slope of the glacier.
Figure 24a. Bench Lakes One through Four are located north of Redfish Lake in Idaho. Lakes One through Three have distinct lateral and terminal moraines. The equilibrium line altitude (ELA) was estimated for Lakes One through Three, and the results are given as Figures 14a-d. The ELA of Bench Lake One, shown here, is approximately 2560 m (8400 ft) based on the AAR method. The ELA estimated by the toe to headwall altitude ratio (THAR) is approximately 2558 m (8390 ft).
Figure 24b. The Bench Lake Two equilibrium line altitude (ELA) estimated for the Younger Dryas aged moraines is approximately 2609 m (8560 ft) based on the AAR method. The toe to headwall altitude ratio method (THAR) estimate ELA is approximately 2567 m (8420 ft).
Figure 24c. The Bench Lake Three equilibrium line altitude (ELA) estimated for the Younger Dryas aged moraines is approximately 2658 m (8720 ft) based on the AAR method. The toe to headwall altitude ratio method (THAR) estimate ELA is approximately 2595 m (8510 ft).
Figure 24d. The Bench Lake Four equilibrium line altitude (ELA) is approximately 2682 m (8800 ft) based on the AAR method. Similarly, the toe to headwall altitude ratio (THAR) ELA estimate is approximately 2646 m (8680 ft).