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Holocene glaciation of the Green River drainage, Wind River Range, Wyoming

Nigel Davies
Western Washington University

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Holocene Glaciation of the Green River Drainage, Wind River Range, Wyoming

By

Nigel Davies

Accepted in Partial Completion

of the Requirements for Degree

Masters of Science

________________________________________________

Moheb A. Ghali, Dean of the Graduate School

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Master’s Thesis

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Holocene Glaciation of the Green River Drainage, Wind River Range, Wyoming

A Thesis
Presented to
The Faculty of
Western Washington University

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

By Nigel Davies

January, 2011
Abstract

The Wind River Range (WRR) has long been the focus of glacial investigations, yet the Holocene record remains poorly understood. Moraines in the Green River Lakes drainage, on the northwest end of the Wind River Range, preserve a remarkably complete moraine record of late-Pleistocene recession, late-glacial and late-Holocene advances. At last glacial maximum (LGM) the study area supported large valley glaciers that extended beyond the rangefront; in historic times, however, glaciers are restricted to high alpine cirques. The largest remaining active glacier (Mammoth Glacier) has retreated to 2 km² and is the primary source of meltwater and outwash to the Green River lakes. In addition to the active glaciers and pro-glacial lakes the study area contains abundant rock glaciers, glacially sculpted and polished bedrock, and post-Pleistocene glacial deposits. To assess the timing and magnitude of late-Holocene glaciation in the Green River Lakes drainage this project used a combination of geomorphic mapping, lake coring and assessment of equilibrium line altitudes.

As part of the funding through the United States Geologic Survey (USGS) EDMAP program, a product of this project was a 1:24,000 scale surficial geologic map of Green River Lakes drainage centered on the Gannett Peak Quad, and incorporating portions of Squaretop, Green River Lake and Dow Mountain Quads. The methods and units I used in this map follow the protocols of a similar mapping effort in the North Cascades National Park and other national parks in Washington (Reidel et al., 2007).

I recovered twelve sediment cores from Upper and Lower Green River Lakes; these lakes trap the majority of the rock flour produced by the Mammoth Glacier. As a result, there is a
continuous record of late Holocene sedimentation in the lakes that serves as a proxy record of glacial activity. Of the eleven cores recovered from the Upper Green River Lake, I focused analyses on the longest Livingston core (3.65m) and one shallow, high-resolution Glew core (75cm). The depth-age model for these cores are based on six AMS radiocarbon analyses (ranging from 430 – 4245 cal. yr B.P.) and $^{210}$Pb dates from the upper-most portion of the Glew core. The one high resolution Glew core (67 cm) from the Lower Green River Lake has a depth-age model that relies on two statistically indistinguishable AMS radiocarbon analyses (3060-3160 cal. yr B.P.).

Rock flour input from Mammoth Glacier is recorded in visual stratigraphy (VS), organic content (OC), magnetic susceptibility (MS), and grain size distribution (GSD) in the Upper Green River Lake cores. The rock flour flux appears to have increased significantly shortly after ~1000 yr B.P., and rapidly after ~500 yr B.P., culminating at ~280 yr B.P. The increase in rock flour after 1000 yr B.P. likely records the rapid growth of Mammoth Glacier at the onset of the Little Ice Age (LIA). A similar rock flour signal is observed via VS, OC, MS and GSD in the Glew core from Lower Green River Lake at ~300 yr B.P.; however, the signal is muted compared to that in the upper lake. The LIA in the region (Schuster, 2000) is the largest Neoglacial advance as recorded in the moraine sequence and lacustrine stratigraphy.

The cores from Upper Green River Lake record the climatic LIA maximum at 280 yr B.P. A small, secondary peak in rock flour flux is recorded by a distinct green-gray sediment dated at ~260-240 yr B.P. This sediment appears to relate to an outburst flood from Scott Lake, a tarn immediately downstream of the Mammoth Glacier. Scott Lake’s high water stand (10 m above modern shoreline - recorded by the “bathtub ring” of oxidation and the stranded delta), was
stable through most of the Holocene. The breach may have been associated with an outburst from Mammoth Glacier, possibly induced by the draining of a side-glacial lake. Multiple ponding locations have been revealed by subsequent ice retreat from LIA maximum. The additional flux from Mammoth Glacier, produced a catastrophic debris flow into Scott Lake causing water to overtop the sediment dam. The down cutting of the dam initiated the breach, flooding the valley with ~ 2,000,000,000 m$^3$ of water, suspended sediment, and debris. The coincidence of the outburst with the maximum rock flour flux suggests a causal link between the flood and the LIA maximum of Mammoth Glacier.

The pre-LIA rock flour record in the Green River lakes is less clear, but consistently high MS and low OC values in sediments deposited between 1000-4500 yr B.P. suggests that the Mammoth Glacier was active but significantly smaller than its historical extent through much of that period. The Upper Green River Lake cores do not indicate any period in the last 4500 years when the Mammoth Glacier was absent entirely.

To constrain temperature and precipitation conditions associated with the LIA and late-glacial glacier maxima, the glaciers of Green River Lakes drainage were modeled using end moraines, interpreted cirque headwalls and till coverage. These reconstructions enabled an assessment of the equilibrium line altitudes (ELA) which vary between 3050 m (late-Pleistocene -THAR) to 3610 m (modern - Mammoth Glacier - AAR). Using SNOTEL, PRISM and imported lapse rates from Colorado (Brugger and Goldstein, 1999) the best estimate of modern climatic conditions (average summer temperature and winter precipitation) at the modern Mammoth Glacier ELA are 3.78 °C and 68.48 cm SWE. Calculated conditions at the paleo-ELAs [LIA: (<1 °C and 2 cm SWE); late-glacial: (<1 °C and 4 cm SWE)] suggest only minor cooling and increased precipitation
was required to cause these advances, consistent with Leonard’s (2007) inference of the sensitivity of WRR glaciers to climate change.
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I would like to thank my advisor Doug Clark for his endless patience and never ending enthusiasm for this project. On day one as we looked at Google Earth, there was a twinkle in Doug’s eye as he finally introduced a student to this dream area. I appreciate his ability to stick with me even when I lost focus (skiing, mountain biking, and triathloning) and went on walkabout. He is not only a good friend, skiing partner, Ski to Sea team member but he has welcomed me into his family, for which I am ever grateful. I would also like to thank my other committee members Andy Bunn and Scott Linneman for their valuable input to the research and writing process.

This project benefited from a multitude of funding sources. The USGS through the successful EDMAP grant (designed to train the next generation of geologic mappers) was the major source. Additional funding came from: the Geologic Society of America, the Tobacco Roots Geologic Society, the Wyoming Geologic Association, and Western Washington University Geology Department. I was paid to teach students geology by Western Washington University, which was a wonderful experience. In the final months the Washington Department of Natural Resources, helped support my need to map rocks.

To make a project in Wyoming take flight from Washington many people on the ground are required. First, the Wyoming Geologic Survey (WGS) and acting State Geologist Ron Surdam provided me support for the EDMAP grant. In the field Seth Witkie (WGS) spent a day reviewing my progress and discussing mapping strategies and philosophies. The United States Forest Service (USFS), who manages the Bridger Teton National Forest where the field area is wholly located, were a great source of support. Ted Porwoll and Terry Svalberg from the science and air
quality monitoring division provided me unending assistance including maps, guidance and emergency PVC splitting in a home workshop. The district ranger was kind enough to provide guidance on the application and permit to retrieve cores and complete work within the Wilderness boundaries. I was also assisted by Hilda Sexauer who provided not only bathymetry maps for the lakes within the drainage but also refrigerated storage for the core samples.

To complete the field work in Wyoming I relied on a host of field assistants who volunteered their time, only to be compensated by creative backpacking meals and the prospect of beer. Sam Bruno, Matt Meyer, Joe Goshorn-Maroney, Joe Mahaffey, Deborah Molsberry, Kyle Kroger and Joe McClenahan all donated 10-28 days of their lives to the cause. These were not easy days; they required 100+ lbs backpacks and the assault of mountains described by WWU professor Robert Mitchell as “the talus field of death”. All of them gave everything they had while battling mosquitos, backcountry cooking, warm beer, and hours on coring rafts. Thankfully the scenery and incredible beauty of the Wind River Range made any inconvenience mostly tolerable.

To complete the analysis of the cores I must thank the folks at Lawrence Livermore National Lab (LLNL) Center for Accelerated Mass Spectrometer (CAMS). Both Tom Guilderson, who provided $^{14}$C dates and advice, and Paula Zermeno who arranged my visit (I was a foreign national at the time) and helped prep my samples were fantastic. Just getting to step foot onto the LLNL campus was a scientific joy. At Western Washington University I was assisted by David Schull and Maggie Esch in completing the $^{210}$Pb dating. I would especially like to thank David who treated me like a professional scientist, not watering down the math needed to properly tease
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1.0 Introduction

The Holocene is defined as the period of relatively stable climate that followed the last global glaciation (Roberts, 1998). Despite its relative stability, small but significant climatic perturbations during the Holocene have resulted in the advance and retreat of glaciers worldwide (e.g., Grove, 1988; Alley et al., 1997). The deposits associated with this glacial activity (e.g., moraines, lake sediments, outwash terraces, etc.) serve as proxy records of these glacial fluctuations (Nesje, 1994; Leonard, 1985; Clark and Gillespie, 1997). Linkages between the depositional histories in the lakes and the surficial deposits can help constrain the climatic conditions (e.g., temperature, precipitation) that ultimately caused those changes (Larsen and Mangerud, 1981; Karlen and Matthews, 1992; Leemann and Niessen, 1994; Souch, 1994).

Whereas studies of large modern glaciers can be used to derive past global climate conditions (mainly Greenland and Antarctica), mountain glaciers are well suited for studies of regional climate. High alpine glaciers respond quickly to minor fluctuations in temperature and precipitation, thus preserving high resolution local records (e.g., Leonard, 1985 and 2007). By combining multiple local study areas across a region, such as the Western United States and Canada, it is possible to develop a regional climate history for the Holocene (Osborn and Karlstrom, 1988; Osborn and Luckman, 1988; Desloges and Ryder, 1990; Clark and Gillespie, 1997; Burrows, 2000; Clague et al., 2001; Laroque and Smith, 2003; Cashman, 2004; Bilderback, 2004; Menounos et al., 2004; Bowerman, 2005; Mijal, 2008).

Although the timing and magnitude of climate change during the last glacial maximum in the inter mountain region of western North America have garnered substantial attention in the past century, fluctuations in Holocene climate and glacier change have only recently achieved wider
interest. For alpine glacier systems, most research has focused on relative and semi quantitative
dating techniques of moraines (e.g., Hall, 1984; Shaw, 1988; Mahaney, 1993; Dahms, 2002).
However, these methods have poor resolution that has led to substantial errors in age
assessments (cf., Osborn and Bevis, 2001). The method of quantitative surface exposure dating,
particularly with in-situ cosmogenic radionuclides (CRN), holds promise for more rigorous dating
of these deposits (Gosse et al., 1995a, 1995b); however, large uncertainties associated with
dating such young deposits and the substantial costs has limited their use for Holocene
sequences. Another limit of moraine sequences is that subsequent larger advances typically
destroy moraines of any prior advances. In contrast, sediment in proglacial lakes provides an
attractive complementary proxy record to the moraines for past glacial fluctuations in that it
they capture a continuous, datable record of glacial sedimentation instead of discrete
“snapshots” of maximum glacier extents provided by moraine records (Leonard, 1985; Nesje,
1994; Matthews et al., 2000).

I investigated late-Holocene glaciation in the Green River Lakes drainage in the Wind River
Range of Wyoming (Figure 1 and Figure 2). I chose this specific drainage because it contains five
important characteristics: 1) active glaciers; 2) pro-glacial lakes to trap the entire Holocene
sedimentary record; 3) simple drainage system such that all rock flour is sourced from
Mammoth Glacier (Figure 3); 4) well preserved moraine evidence up valley; and 5) crystalline
gneiss and granite bedrock. In the high alpine and actively glaciated regions, I mapped
geomorphic features and landforms (ice limits, moraines, rock glaciers, debris fans, till, etc.).
The focus of mapping in the lower valley was on alluvial fans, colluvial slopes, floodplains, and
outwash terraces. In the context of this mapping, I collected multiple lake sediment cores from
the two prominent lakes that directly capture alpine glacial runoff, Upper and Lower Green River lakes. By documenting variations in character and fluxes of sedimentation in the cores, I was able to construct a timeline of late-Holocene glaciations in the headwaters that feed the lakes. By linking the sediment pulses in the lakes to complementary moraines below the glaciers these data can be tied to paleo-temperature and precipitation conditions using estimates of ELA. Additionally, I became interested in the history leading to the formation and subsequent lake level lowering of Scott Lake, a high alpine tarn in the Wells Creek drainage. Glacial outburst flood are documented in the Wind River Range (Oswald and Wohl, 2008), as a result of a climate change, revealing clues for the mechanism of landscape evolution. Furthermore, ice core studies in the range provide a valuable temperature and mass balance proxy record that allows a calibrated and correlated sedimentary record. The results from my study provide a valuable complement to the growing body of proxy constraints on Holocene climate change in the montane west. As such, this study tests the driving climatologic factors (e.g., temperature, precipitation) that have and continue to affect this portion of west-central Wyoming and the Western Cordilleran.

2.0 Background

2.1 Location and Physiography
The Wind River Range, west-central Wyoming (Figure 1), runs ~350 km along a northwest – southeast trend. The range supports 64 named glaciers, many high alpine lakes, rugged peaks > 4000 m high (including Gannett Peak 4052 m – highest point in Wyoming), U-shaped valleys, hanging tributaries and alpine meadows and marshes. Major valleys on both flanks of the Wind
River Range supported large Pleistocene valley glaciers, some extending > 50 km in length with a minimum thickness of 600 m (Blackwelder, 1915; Chadwick et al., 1997; Locke, 2002). The plains surrounding the range are high desert prairie with a general elevation of 2,000 m, creating a sharp elevation gradient along the flanks of the range. On the southwest side of the Wind River Range Pleistocene glaciers created a broad moraine belt characterized by numerous large southwest-northeast trending lakes (e.g., Fremont Lake, New Fork Lake) (Figure 1). The combination of significant active glaciers and abundant lakes make the range an excellent target for a geomorphic mapping and lake-coring study (Figure 1 and Figure 2).

2.2 Bedrock Geology

The Wind River Range was initially uplifted during the Laramide Orogeny (Late Cretaceous to mid-Eocene), exposing a lower Precambrian plutonic and high-grade metamorphic core (Figure 4); Cerveny and Steidtmann, 1993). The lower valleys expose upper Paleozoic and Mesozoic sedimentary rocks that mostly dip to the southwest, perpendicular to the crest of the divide (Figure 4). In the Green River Lakes area, these formations include the Bighorn Dolomite, Madison Limestone and TenSleep Sandstone (Green and Drouillard, 1994 and Figure 4). The Eocene Green River Formation records sediment shedding off the range during the Laramide uplift and denudation. During the Quaternary, the range experienced repeated extensive alpine glaciation; Bull Lake noted by Blackwelder (1915) occurred between 130-100 ka during marine isotope stage 6 and 5d and Pinedale reached its local maximum between 23-16 ka during the later portion of marine isotope stage 2 (Chadwick, 1997). Since the Pleistocene, recessional pauses and high alpine advances have left indications of minor glacial episodes (Mahaney 1993). Products of latest Pleistocene and Holocene glaciations are ubiquitous in the high country,
including abundant moraine and outwash deposits, highly sculpted bedrock, horns, tarns, and other ice flow features.

### 2.3 Climate

The Wind River Range is affected by both Pacific-dominated Westerlies in the winter and summer monsoonal moisture originating in the Gulf of Mexico (McCabe et al., 2004; Gray et al., 2004). Although the instrumental climate record is relatively short and spatially restricted, there are a few reliable sites in the range. In particular, the Natural Resources Conservations Services (NRCS) maintains weather station and SNOTEL sites across the region including the closest at Gunsight Pass, ~15 km west of the Lower Green River Lake (Figure 5). Over the past 30 years the Wind River Range has averaged 20 – 150 cm/yr of precipitation, depending on aspect and altitude (NRCS – SNOTEL) (Figure 5). Average annual precipitation at Gunsight Pass (2993 m elevation) on the western flank of the Wind River Range is approximately 23 cm (NRCS-SNOTEL-Gunsight Pass) (Figure 5). In an 800 m PRISM (Parameter-elevation Regressions on Independent Slopes Model) grid (lat. -109.66, long. 43.16) centered on Mammoth Glacier the average winter snow fall (Oct.-April) over the past 115 yr is 63.14 cm SWE (PRISM, 2011). The large elevation gradient, contrasting the high rugged mountain topography with nearly 100 km of low, gentle topography to the southwest, creates substantial orographic and convective flow driven by prevailing winds from the west-southwest. Precipitation across the range falls primarily as snow during the winter months; although rain dominates during summer months, snow can occur at much of the higher elevations (Naftz, 1993a). As noted by Leonard (1989) summer snow accumulation does not play a major factor in the health of most temperate glaciers.
As in many continental alpine settings, daily temperature fluctuations can be substantial. During the summer months daily temperatures can vary from 0-30 °C, whereas during winter month’s temperature can vary from -40 to 5 °C (NRCS – SNOTEL). Average annual temperature at Gunsight Pass (2993 m elevation) is approximately 1.5 °C (NRCS-SNOTEL-Gunsight Pass). In the same 800 m PRISM grid (lat. -109.66, long. 43.16) centered on Mammoth Glacier the average summer temperature over the past 115 yr (June-July-August) is 6.27 °C (PRISM, 2011). The combination of relatively cool summer temperatures and abundant winter precipitation in the Wind River Range supports the relatively large glaciers.

2.4 Previous Work

Because of its extensive evidence of glaciation and well-preserved moraine sequences, the Wind River Range has been the focus of glacial research for more than a century (e.g., Hayden, 1878). This history provides one of the foundations for the regional glacial stratigraphy in wide use throughout western North America.

2.4.1 Western Cordilleran Glacial Chronology

Recent studies suggest similar yet distinct patterns of Holocene glacial fluctuations in the western maritime and Rocky Mountain ranges. Radiometrically controlled proxy records suggest glacial advances in maritime mountains are centered at 8 ka, 5 ka, 3 ka, culminating in the Little Ice Age (LIA) maximum ca. A.D. 1850 (Clark and Gillespie, 1997; Leonard, 1997; Menounos et al., 2004; Bowerman, 2005). The LIA is a well-documented historical period of northern hemisphere cooling affecting the Rocky Mountains (Carrara and McGimsey 1981; Grove, 1988).
As documented by Leonard (2007), glaciers in the Wind River Range are particularly sensitive to small temperature perturbations (Figure 6). Leonard modeled the effects of temperature depression with constant precipitation to document the rate of regrowth toward last glacial maximum (LGM) ice extents across multiple Rocky Mountain ranges. With only 2°C of modeled cooling in the Wind River Range between 5-10% of the LGM ice extent had returned, whereas other ranges had yet to respond (Figure 6). This climatic sensitivity exercise emphasizes the value of investigating the Holocene glacial record in the Wind River Range.

2.4.2 Wyoming Glacial Research

Most studies of glaciation in the Wind River Range have focused on the Pleistocene record. Blackwelder (1915) designated the deposits at Bull Lake and Fremont Lake (Pinedale) as the type localities for Quaternary glaciation in the Rocky Mountains. Blackwelder’s initial chronology was refined by recent investigations of moraine and outwash stratigraphy around the range (e.g., Holmes and Moss, 1955; Hall and Shroba, 1995; Fall et al., 1995). Cosmogenic radionuclides exposure dating (³⁶Cl and ⁱ⁰Be) of moraines at Bull Lake and Pinedale further refined the timing of these events to include 15 and 3 advances, respectively (Gosse et al., 1995a, 1995b). The majority of glacial investigations in the Wind River Range have focused on the northeast side of the range (Blackwelder 1915; Chadwick 1997). Locke (2002) documented the extent of the Green River Lake glaciers as part of the entire LGM Wind River Range extent. These studies have helped to constrain the timing and magnitude of advances during LGM in the Wind River Range; however, the post-Pinedale and Neoglacial chronology has received far less attention.

Small moraines in Temple and Titcomb basins (Figure 1) were initially interpreted as early Neoglacial [Indian Basin Advance] (Miller and Birkeland, 1974; Mahaney, 1984). This age
assignment was based on relative moraine weathering and morphostratigraphic relationships. Subsequent lake coring, radiocarbon dating, and palynology work by Davis and Osborn (1987), Zielinski and Davis (1987), and Fall et al. (1995) conclusively established the Indian Basin moraines to be 11,000-12,000 cal. yr B.P. Further evidence for the reassignment of Indian Basin moraines to late-glacial is provided by Gosse et al. (1995a and 1995b) who dated the Inner-Titcomb Lakes moraines as having formed 11 – 13 ka using $^{10}$Be CRN exposure dating. This dating placed the Inner-Titcomb Lakes moraine as Younger Dryas in age, although Gosse et al. leave the question of ice re-advance or recessional pause from Pinedale maximum unanswered.

Mahaney (1993) developed a Holocene glacial chronology in the headwaters of the Green River Lakes drainage, comprising three Neoglacial advances [Indian Basin (5-5 ka), Audubon (2-1 ka), and Gannett Peak (LIA)], based on relative weathering, lichenometry, and soil development (Figure 7). He notes a lack of suitable organic material for radiocarbon dating from moraine material especially at his site GR3 (Mahaney, 1993 - figure 1) at the base of Tourist Creek and a lack of tephra deposits from Cascade or Yellowstone volcanoes. Mahaney’s mapping indicates that a Neoglacial advance (Indian Basin) deposited a till dam at the western margin of Scott Lake. Mahaney mapped a relict shoreline of Scott Lake, with lacustrine deposits between relict and modern shorelines (Indian Basin deposits lining the exposed shore with Audubon deposits draped over in places). However, Mahaney provides no explanation for the exact timing or mechanism allowing the lake level to lower and establish the current shoreline. Furthermore, Mahaney suggests that rock glaciers in the Wells Creek basin only record two phases of advance in the Neoglacial period, a more extensive Indian Basin advance and a less extensive Audubon advance (Mahaney, 1993). From Mahaney’s mapping the Gannett Peak event appears not to
have any rock glacier activity. Mahaney’s reliance on low-resolution dating techniques such as relative weathering and lichenometry, which use complicated and multivariable erosion and growth curves respectively, calls into question many of his conclusions, particularly in light of the other recent studies in the range that do not document pre-Gannett Peak (LIA) Holocene advances (Figure 7). Mahaney’s work, however, remains the only published research of Holocene-age glacial deposits in the basin and helps provide a foundation for my study.

2.4.3 Fremont Ice Core

In the last 20 years, the United States Geologic Survey (USGS) has collected and analyzed two ice cores (Naftz, 1993a; Naftz, 1993b; Naftz et al., 1997; Schuster et al., 2000) from the Upper Fremont Glacier located immediately east of the continental divide (Figure 8) and less than 1 km from Mammoth Glacier. In 1991, Naftz et al. (1993a and 1993b) drilled a continuous 160 m ice core to bedrock. A fortuitous grasshopper leg from 152 m depth in the ice core allowed for a radiocarbon date which indicates the record spans 270 years of accumulation (Schuster et al., 2000). In addition to the age constraints, the ice core was analyzed for δ18O, direct-current electro-conductivity (ECM) and dust composition. The most notable shift from mean isotopic δ18O Standard Mean Ocean Water (SMOW) values occurs below 101 m. This perturbation from the modal δ18O SMOW is thought to represent the warming trend which terminated Little Ice Age (LIA) and initiated glacial retreat at 1840 A.D. (160 cal. yr B.P.) in the Wind River Range (Naftz, 1996) (Figure 8). This shift corresponds with cores from other mid-latitude glaciers (e.g., Quelccaya Ice Cap (Peru)) (Naftz, 1996). Beyond the LIA climate shift the isotopic record and annual accumulation in the ice core may provide future studies a potential high-resolution proxy
record that may help to link the shorter, problematic instrumental climate record to longer period proxy records like lake sediment cores.

2.4 Research Site Selection

The headwater of the Green River has five important characteristics that make the study area an attractive location in the Rocky Mountain region for a detailed glacial-lacustrine study: 1. It has a single dominant active valley glacier (the Mammoth Glacier) with a documented set of apparent Holocene-aged moraine deposits; 2. The modern outwash is rich in suspended rock flour from the Mammoth Glacier, which flows into a set of pro-glacial lakes (e.g., Upper Green River Lake, Scott Lake) that are downstream of the Holocene terminal moraines; the lakes should thus record the entire Holocene outwash record for the glacier; 3. Although there are a few other active glaciers upstream of Upper Green River Lake, their contribution of rock flour to the Upper Green River Lake is insignificant compared to that originating at the Mammoth Glacier because of their small size. Also, each of these small glaciers has several paternoster tarns that capture nearly the entire suspended load before reaching the main Green River; 4. A distinct moraine sequence upstream from these pro-glacial lakes ensures that the lake coring does capture a variation in sedimentation rather than a sterile, non-glacial environment; 5. The bedrock in the study area is primarily crystalline igneous and metamorphic units, reducing the potential for contamination or reservoir effects in organic samples for radiocarbon dating. Together these factors make the Green River Lakes drainage an excellent location to constrain the Rocky Mountain Holocene glacial chronology.
3.0 Methods

The approach combined field and laboratory techniques to develop a radiocarbon-controlled multi-proxy record of Holocene climate change in the Green River drainage. This approach incorporated surficial mapping, lake sediment coring, radiocarbon dating, and equilibrium line altitudes (ELAs) to reconstruct the glacial history. These reconstructions, in turn, provide a means to constrain the climatic conditions that drove the glacial fluctuations. In accordance with a grant from the USGS as part of the EDMAP program, I produced a GIS map at 1:24,000 scale of the surficial geologic map of the Green River Lakes drainage.

3.1 Mapping

The objective of my field mapping is to characterize the Holocene and late Pleistocene geomorphic features and deposits in the Green River drainage, particularly in reference to their relationship to sediments in the Green River Lakes. Previous mapping by the Wyoming Geologic Survey has produced regional surficial maps at 1:500,000 scale (Love and Christiansen, 1985). The coarse resolution of this mapping is not sufficient to characterize the glaciers and geomorphic setting for my study. I therefore created a geomorphic map at a 1:24,000 scale using: USGS 7.5’ topographic quadrangles, 10 m/pixel Digital Elevation Model (DEM) data (Wyoming Geographic Information Science Center [WYGISC]), 2002 Digital Ortho Quarter Quadrangles (DOQQ) at 1 m/pixel (Wyoming GIS Coordination Structure [WYGIAC]), and USFS 1996 and 2007 aerial photographs, combined with traditional field mapping of local sites in the area. I compiled the mapping in a GIS in ArcMap 9.3. The geomorphic map units are based on detailed unit descriptions from work published in the North Cascades National Park and the USGS surficial mapping guidelines (Riedel et al., 2007). The mapping units located in the field of
study include: Bedrock, Glaciers, Rock Glaciers, Talus, Debris Cones, Debris Cones Apron, Debris Avalanches, Outwash, Colluvium, Meadows, Valley Wall, Valley Floor, Rivers and Lakes. Other line and polygon features that I mapped include: Rivers, Moraines, Paleoglaciar flow indicators, and equilibrium line altitudes. Moraines were distinguished based on stratigraphic position and relative weathering parameters including soil development, lichen cover, and boulder weathering (e.g. Clark et al. 1994). Reconnaissance mapping using aerial photography was field checked locally to confirm the initial interpretations.

3.2 Lake Bathymetry

I determined the best lakes to core based on geographic location, relationship to outwash source and feasibility of access. I then constructed detailed bathymetry maps for each lake using GPS and a handheld sonar depth meter. These surveys were conducted by transecting the lakes repeatedly in an inflatable 2-person kayak. On each transect, I recorded a depth and position measurement every 5 meters. The depth was measured using a Fishray Manta FR-300 handheld sonar device (nominal accuracy ~0.3 m), and the XY position was recorded using a Trimble GeoExplorer 3. Each subsequent transect was initiated from the endpoint of the previous transect. The lake margins were established from 1m/pixel DOQQ in ArcGis (v9.3). All of the GPS points were downloaded to GPS Pathfinder office and differentially corrected (nominal accuracy ~3 m) using a base station (~10 km) operated by the USFS – Bridger Teton National Forest. These points were imported into Surfer 8.0 to create contoured bathymetric maps for each lake. As I was unable to process the bathymetric mapping in the field, I relied primarily on previous bathymetric maps from Wyoming Fish and Game (1974) with the addition of point data from my own transect work. After returning to WWU, I used the XYZ data from
the transects to construct the bathymetric maps for Scott Lake, Upper and Lower Green River lakes. I present some additional bathymetry data in the appendix.

### 3.3 Lake Coring

The specific coring sites in each lake are based primarily on the lake bathymetry. The best coring sites typically are located in the deepest parts of a lake, distal from any actively aggrading deltas at the inlets to the lakes, and also away from steep slopes that might release sub-lacustrine slumps or turbidity currents. These sites typically preserve the longest, most complete, and least disturbed sedimentary records. The limitations of coring techniques included: the challenge of maintaining position of Livingston coring raft; the limited amount of Livingston casing; and the unavoidable tangling of Glew core lines at water depth of > 30 m. As such, I was forced to seek cores from shallower, sloped portions of the lakes.

#### 3.3.1 Livingston Corer

The Livingston corer consists of a 1.2 m long by 5 cm wide stainless steel core barrel with a 10 cm double 0-ring piston attached via wire to the surface and capped with a steel core-barrel head (Wright, 1967). During coring the piston is held fixed by tensioning the wire. The piston acts to reduce “coning” of the sediments along the core barrel during coring, and creates suction to hold the ~1 m sediment core in place during extraction. To reach the desired coring depth additional 1.5 and 2 m Magnesium-zirconium alloy extension rods are added sequentially to the core-barrel-head. During the coring, the coring stem is lowered from the lake surface through 8-cm PVC sectional tubing that is seated on the lake bottom and extends between 0.5 – 1 m above the lake surface. The casing helps to reenter the hole during multiple pushes with the corer and also prevents external material from entering the core hole between pushes. This
also allows multiple pushes at the same coring location. After each push, the filled tube is removed and capped, and a clean core barrel is attached to the core head. The cores were extruded back on shore into cellophane-lined 2” PVC for transport to the lab. The Livingston corer was used in summer from an anchored raft.

### 3.3.2 Glew Corer

The Glew corer is a single-push system designed to recover the young, poorly consolidated sediments at the sediment-water interface as well as the top 100 cm’s of sediment. It consists of a 1.15 m long by 8 cm diameter clear plastic core barrel attached to a one-way compression valve (Glew, 1989). A free-sliding hollow percussion weight (5 kg) slides over the metal rod (50 cm) that connects the plastic, one-way valve, core barrel system to the surface lash rope. Both the percussion weight and coring devices are connected to the surface by separate ropes (a main line attached to the rod – which connects to the valve and core barrel, and a secondary line attached to the percussion weight). This setup allows the device to be deployed in water of depths < 50 m (although complications arise at water depths of more than 30 m due to tangling of ropes and the inability to maintain lateral position). Coring involves dropping the entire core barrel and percussion weight from 2 m above lake bottom, then further pushing the barrel into the mud with repetitive blows with the weight. Upon refusal, the weight is removed and the core is raised to the surface of the lake, the sediment is held into the core barrel by the one-way valve. Cores are capped and returned to shore to record visual stratigraphy photographically and magnetic susceptibility analysis (see core analysis for details). At each coring site 1-2 parallel cores were collected at water depths of between 2-30 m, and recovered ~35 - 85 cm of
sediment. Because of their poor consolidation and transportability, sediments in Glew cores were sampled in the field at 1 cm intervals, and stored in sealed ziplock bags.

3.4 Core Analyses
The objective of coring lake sediments is to characterize the depositional environments that produced the types of sediments. Alpine lacustrine sediments typically have both organic and inorganic components. The inorganic or clastic sediments are mainly derived from local bedrock and hillslope processes in the basin, with minor wind-blown dust from outside the basin. A major additional source of clastic material in the Green River drainage are the glaciers in the highest parts of the drainage. Temperate glaciers can produce prodigious amounts of fine silt-and-clay sized particles (rock flour) due to abrasion at their base as they slide over bedrock. The production of rock flour scales with rate of sliding, bedrock lithology, and surface area of the glacier (e.g. Leonard, 1985; Leemann and Niessen, 1994; Zillen et al., 2003); the rate of deposition of rock flour in downstream lakes therefore tends to vary over time as glaciers advance and retreat. Stochastic events, like debris flows and outwash floods, can dramatically change the instantaneous rate of deposition of the clastic component. Thus the clastic deposition into lakes typically responds rapidly to changes in climate or geomorphic processes (Fritz, 1996).

Organic content tends to be buffered against dramatic changes and thus provides a more consistent background signal from the lake. The organic sediment is generally locally derived from decaying biotic material and in most alpine regions, maintains a relatively consistent flux to lakes during the Holocene (Filippelli et al., 2006). The rate of change of alpine biota relative to changes in hillslope processes that produce non-organic sediments (i.e., clastics) is slow.
Increases in organic sedimentation in alpine regions typically accompany warmer periods (e.g., Mackereth, 1966) whereas cooler periods typically reduce the flux of organic sediments. Because of the rate of change of alpine biota relative to changes in hillslope processes that produce non-organic sediments (i.e., clastics) is slow, sharp changes in organic content values likely represent changes in clastic flux rather than organic production.

### 3.4.1 Magnetic susceptibility

Magnetic susceptibility (MS) provides a rapid, non-destructive quantitative assessment of stratigraphy in the cores, both in the field and after transport to the laboratory. In basins dominated by crystalline bedrock (e.g., Green River drainage), most clastic sediment contains iron-bearing minerals such as biotite, hornblende, and magnetite. Climate variations dictate the weathering and erosion rates of the watershed’s bedrock, resulting in variation in magnetic susceptibility over time. By measuring the effect of these particles on an induced magnetic field, MS provides a reliable first-order assessment of the organic vs. clastic sediment ratio along a core (Thompson et al., 1975). The measurements of MS reflect bulk sediment composition allowing sediments from different core sites in a lake to be correlated (Thompson et al., 1975; Nowaczyk, 1992). However, as the catchment size increases and bedrock composition becomes more complex, such correlations can become more tenuous. Because of the relatively dramatic flux in clastic sedimentation from the Mammoth Glacier, this concern does not appear to be a substantial problem in the Upper Green River Lake.

For Glew cores from both Upper and Lower Green River lakes, I measured MS values at 1 cm intervals using a Bartington MS2C core-scanning susceptibility meter. For the Livingston cores
from Upper Green River Lake, I measured MS at a 2 cm intervals. For each core, I applied a linear drift correction using the program Multisus.

### 3.4.2 Visual Stratigraphy

The color of sediment offers a valuable qualitative index to the depositional environment and dominant sediment type. Organic-rich sediment is usually dark brown to black whereas glacially derived sediment (rock flour) is usually gray to light gray blue (Leonard, 1997). Additionally, sediment derived from hillslope processes is usually coarser or biomodal in grain size, and composed of an oxidized component (medium to light brown) as they are derived from previously weathered grains. As the Glew cores could not be transported intact to the lab, the cores were analyzed in the field for color, gross sedimentary structures, and photo documentation prior to extrusion, however smearing affected these observations.

In the lab, the Livingston cores were split lengthwise and logged for sedimentary structures, macrofossils, and grain-size estimation. Subsequently, they were photographed for archival and graphical presentation. Additionally, I prepared smear slides of specific horizons to detail microscopic grain size difference and composition. After these initial non-destructive assessments, the cores were sampled for organic content, quantitative particle-size analysis, radiocarbon and lead dating.

### 3.4.4 Loss on Ignition (LOI) – Organic Content (OC)

Loss-on-ignition (LOI) is a quantitative analysis of organic sediment content that can be used to evaluate paleo-environmental conditions in alpine settings. Studies have shown an inverse correlation between glacier extent and LOI/OC in alpine glacial settings similar to the Wind River
Range (Leemann and Niessen, 1994; Levesque et al., 1994; Meyers and Teranes, 1998; Meyers and Lallier-Verges, 1999).

LOI analysis involves multiple steps; for my study, I sampled sediment volumes of ~1 cm\(^3\) at 1 cm (Glew corer) and 2 cm (Livingston corer) increments. Samples were weighed then heated to 100 °C for 12-24 hours until dry; they were reweighed, heated at 450 °C for 4 hours, placed in a desiccator to cool and weighed again. The initial water content of the sediment can be evaluated using this equation:

\[
H_2O\text{ content} = \frac{(W_{wet})-(W_{crucible})-(W_{dry100})-(W_{crucible})}{(W_{wet})-(W_{crucible})} \times 100 \tag{1}
\]

The modified and simplified LOI equations below are used to estimate the total organic content (Heiri et al., 2001):

\[
LOI = \frac{((W_{dry100})-(W_{crucible})-(W_{dry450})-(W_{crucible}))}{(W_{dry100})-(W_{crucible})} \times 100 \tag{2}
\]

\[
LOI = \frac{(W_{dry100})-(W_{dry450})}{(W_{dry100})-(W_{crucible})} \times 100 \tag{3}
\]

where \(W_{wet}\), \(W_{dry100}\) and \(W_{dry450}\) represent sediment mass and \(W_{crucible}\) is the crucible mass.

Typically, LOI/OC values are measured at a maximum burn of 550 °C, but Santisteban et al. (2004) notes that heating samples between 425 °C and 520 °C induces breakdown of magnesite,
siderite, rhodochrosite, and dolomite, that can cause overestimates of the total biotic organic content. To minimize this effect, I used a longer burn time at the lower temperature.

3.4.5 Grain Size Distribution
Lake-sediment grain size reflects the sources and processes of sediment transport, and typically varies from coarse sand to very fine silts and clays in alpine lakes. Coarse-grain sediments in such lacustrine settings can have several origins, such as storm sheet-wash, debris flow, avalanche, and ice rafting (Karlen and Matthews, 1992). Finer grained sediments are typically transported as suspended sediment in stream flow, or as eolian dust. In glaciated regions with crystalline bedrock, the suspended sediment flux is typically dominated by outwash from extinct glaciers. Such outwash contains abundant suspended silts and clays (rock flour) produced as the glaciers scour the underlying bedrock (e.g., Karlen and Matthews, 1992; Clark and Gillespie, 1997; Bowerman, 2005; Mijal, 2008). Pro-glacial lakes trap this sediment and develop detailed proxy records that preserve a record of rock-flour flux from the glacier(s). Many workers (e.g., Leonard, 1989; Bowerman, 2005; Mijal, 2008) have interpreted a close relationship between glacier extent and rock flour flux to pro-glacial lakes.

I sampled sediments at 1 cm (Glew corer) and 2 cm (Livingston corer) intervals through each core in preparation for grain size analysis using a Malvern Mastersizer 2000 laser particle-size analyzer with autostage (nominal particle-size range: 0.01 – 2000 μm). Prior to running each sample in the Mastersizer, all sediment samples were soaked overnight in baths of 30% hydrogen peroxide and then 1M sodium hydroxide to digest organics, diatoms and other biogenic materials. The objective of the analysis is to measure only the clastic grain size derived from geomorphic processes. However, when a sample showed a bimodal distribution with a
coarser peak in the known diatom size range this portion of the analysis was excluded. Thus, the remaining percentages were normalized to reflect the distribution of only clastic grain sizes.

### 3.5 Radiometric Dating

#### 3.5.1 Radiocarbon Dating (\(^{14}\text{C}\))

Radiocarbon dating is one of the most widely accepted, reliable, and trusted dating methods available. Accelerator Mass Spectrometry (AMS) allows extremely small organic samples (~0.1 mg) to be analyzed accurately. Lake sediment cores commonly contain small macrofossils (e.g., wood, needles, and leaves visible to the naked eye). Ideally, multiple, regularly spaced \(^{14}\text{C}\) samples are analyzed throughout a give core, from which a continuous record of sediment age (depth-age model) can be developed.

I sampled small needles, seeds, and twigs for AMS \(^{14}\text{C}\) dating. Such small macrofossils are typically incorporated into lake sediments the same year they die. Because they rapidly decompose sub-aerially, these samples should provide a close date to the age of the sediments in which they are encapsulated. Large organic samples such as branches or bark, could potentially last many years after death, and thus may induce significant error in a date. Similarly, charcoal is chemically quite stable at Earth’s surface and is less suitable for dating than small macrofossils. As such, I used only smaller macrofossils. I collected samples at or near major stratigraphic horizons (e.g., color gradient changes, sand lenses, rock flour pulses, MS transitions, etc.). Each sample was dried overnight and sealed in 20 ml scintillation vials until analysis at Lawrence Livermore National Lab (LLNL).
AMS radiocarbon sample preparation involves soaking of each sample in baths of 1 normal HCL and NaOH at 90˚ C for 30 minutes to remove any residual mobile carbon from the soil or sediment in which the material was preserved (Pilcher, 1991; Trumbore, 2000). The NaOH soak is repeated until the solution is clear and free of humic acid. I prepared the samples at Lawrence Livermore National Laboratory (LLNL). The samples were then sealed into vacuum tubes prior to the graphitization process. The samples were run at LLNL on the AMS under the guidance of Tom Guilderson and corrected by using the CALIB 6.0 software (Stuiver and Reimer, 1993).

Calibration of ¹⁴C ages often produces multiple age windows with varying probabilities (Stuiver and Reimer, 1993). To report the calibrated ages, I chose the interval (or “window”) with the highest probability and report both the upper and lower bounds in calibrated years B.P. For specific dates to input to depth-age models I used a mean of this calibrated age window.

**3.5.2 ²¹⁰Pb Analysis**

To obtain a high-resolution record from the uppermost portions of the Glew cores, I used ²¹⁰Pb dating (Eakins and Morrison, 1977; Appleby and Oldfield, 1979). The short half-life of ²¹⁰Pb (~22.3 yr) only allows dating the past 100 to 120 years. However, this is a crucial interval as it spans the historic and instrumental record at the end of the Little Ice Age in particular. It is also an interval where ¹⁴C dating is not useful because of production-rate decline in the atmosphere. The ²¹⁰Pb data provides age constraints for the uppermost portion of the sediment record, and a means to link lake sedimentation to climate changes recorded in the Fremont Glacier ice core (Schuster et al., 2000).
$^{210}\text{Pb}$ is a naturally occurring radioisotope of $^{238}\text{U}$ decay series, with its closest parent being $^{222}\text{Rn}$. The dating model assumes a constant flux of $^{210}\text{Pb}$ from the decay of $^{222}\text{Rn}$ followed by the adsorption on suspended particles which carries the excess $^{210}\text{Pb}$ into the lake. However $^{210}\text{Pb}$ is also naturally occurring within the clastic sediments in secular equilibrium with $^{226}\text{Ra}$ (measured by $^{214}\text{Pb}$).

The $^{210}\text{Pb}$ activity was measured at Western Washington University with a Canberra GL 282R Ge gamma spectrometer under the supervision of Dr. David Shull. I analyzed $^{210}\text{Pb}$ in sediments from a Glew core from the Upper Green River Lake. Samples were prepared by drying for 24 hours in a 120 C oven. The resulting sediment was powdered and accurately weighted out to 10 g (when available) and placed into 70 mm plastic containers suitable for use in the spectrometer. The samples were capped and set aside for 14 days to ensure proper accumulation of $^{214}\text{Pb}$. Samples were run at 1 cm spacing for the first 5 cm and ~ 3 cm spacing from 5 – 15 cm, the estimated point where excess $^{210}\text{Pb}$ reached zero (supported = excess). To calibrate the results, a standard was run as mixture of a preprocessed sample and pitch blend ore to establish the efficiency of the detector.

To calculate the age of a known sediment depth I used this equation for constant supply model:

$$A_z = A_0 e^{-\lambda z/s}$$  \hspace{1cm} (4)

Where $A_z$ is the excess at depth $z$, $A_0$ is the excess as the top of the core, $\lambda$ is the radioactive decay constant (yr$^{-1}$) of $^{210}\text{Pb}$ and $s$ is the sedimentation rate.

As expected, with increased depth of sample, the excess $^{210}\text{Pb}$ diminishes until $^{210}\text{Pb}$ and $^{214}\text{Pb}$ are in secular equilibrium. Secular equilibrium occurred where the excess decays per minute
approach zero and remain at the same value with continued sampling at greater depth. To establish a value for the initial amount of excess $^{210}\text{Pb} \left(A_0\right)$ I used a linear interpretation of the data above secular equilibrium (included the first point where secular equilibrium was reached).

### 3.5.3 Depth-Age Models

In the simplest situation, sediment age increases with depth as a function of sedimentation rate and compaction from increased overburden. Bioturbation and slumping may disrupt this pattern by mixing, removing, or repeating a period of deposition. Visual stratigraphy in all cores indicated these problems were not significant in the Green River Lakes cores. To construct depth-age models for each core, I input the mean age of the highest-probability calibration window at each sample depth, then assumed a linear interpolation between each sample. By interpolating between known dates the depth-age model accounts for sediment compaction. Because of variations, the rate of sedimentation between dated horizons is unknown. I used a simple linear interpolation; however, it remains possible that a power or exponential interpolation better defines the reconstruction. At the bottom of cores, I extrapolated linearly below the deepest date using the slope of the overlying interval. I assumed that the very top of the core (0 cm depth) reflects modern deposition (0 yrs B.P.) although it is possible that the uppermost sediment was recently disturbed by the coring process. The consistency of the uppermost stratigraphy implies any such disturbances were likely minor.

I processed and created depth-age models for two cores from the Upper Green River Lake, UGL-5 and UGL-LIV5. The Livingston core (UGL-LIV5) preserves a detailed record of radiocarbon dates including one just below the peak MS value. The Glew core (UGL-5) had some viable dateable material, but all returned dates younger than 200 cal. yr B.P. As such, I was forced to
look elsewhere for age control in UGL-5. To accommodate this problem, I correlated the two cores from Upper Green River Lake (UGL-LIV5 and UGL-5), based on thickness and relative position of the Little Ice Age MS peak. Using the ratio of LIA sediment thickness between cores, I transferred a date from UGL-LIV5 to UGL-5; the date is located at the base or initiation of the LIA MS peak in both occurrences.

3.6 Equilibrium-Line Altitudes

Equilibrium line altitudes (ELA) are the theoretical lines of zero-net mass balance on glaciers (e.g., Benn and Lehmkul, 2000). The primary controls on the ELA of temperate alpine glaciers are summer temperature, winter precipitation, and topographic characteristics such as hypsometry and aspect (Figure 9). The latter factor is particularly important for small cirque glaciers in which northern aspects and tall cirque headwalls both enhance wind-drifted accumulation and greatly decrease summer ablation. As glaciers extend down valley from their cirques, these topographic factors decrease (Leonard, 1989). The focus of my study, the Mammoth Glacier, is large enough to extend well past the zone of cirque shading and enhanced accumulation. Its mass balance should be therefore more closely tied to regional climate.

3.6.1 Methods

ELA for bare ice glaciers can be determined by a variety of methods, with the most commonly used methods being the Accumulation-Area Ratio (AAR), Toe-to-Headwall Altitude Ratio (THAR), and Mean Altitude (MA). The AAR method is a simple method developed by Porter (1975), in which the entire glacier area is measured and an assumed value (generally 0.60-0.65) is used to calculate the fraction of area above the ELA (0.60 for this study). The THAR ratio differs by identifying the ELA as roughly the mid-point altitude between the cirque headwall and the
terminal moraine and applying a ratio 0.6:0.4 (accumulation: ablation) (Benn and Evans, 2003).
The MA approach simplifies the glacier even further from THAR ratio by finding the mean
elevation (1:1) between cirque headwall and terminal moraine.

In my study, I used the AAR, THAR, and MA methods for the well-constrained modern glaciers
and the LIA reconstructions in Wells Creek drainage. I evaluated three extents (LIA maximum-
1850 – mapped, 1966 – topographic map based on air photos and 2002- air photos and satellite
imagery) to establish the rate of ELA rise and total glacier area decrease. To establish the area
of Little Ice Age, I used the mapped extent of LIA till from the geomorphic map I created. For
the 1963 time step I used the USGS topographic map in geotiff format and for 2002 I used the
2002 DOQQ air photo from WYGIS clearinghouse.

For advances that have poorly defined ice extents, AAR reconstructions would have been
difficult and included large uncertainties and error. For these situations, I relied on the simpler
methods of THAR and MA. While THAR and MA are not as rigorous a methodology they still
serve to establish the ELA for all glaciers in my field area.

3.6.2 Climatic Implications

In compiling data of 32 well studied glaciers around the world, Leonard (1989) defined a
graphical bound window of precipitation and temperature conditions at ELAs of modern glaciers
(Figure 9). The window is defined by plotting mean summer temperatures (measured as the
average temperature for June, July and August) against the total winter accumulation
(measured as the winter snowpack in cm of snow water equivalent [SWE]). This relationship
was mathematically defined by Kotlyakov and Krenke (1982);
\[ A = 1.33(T + 9.66)^{2.85} \] (5)

\[ A = 1.33(T + 6.66)^{2.85} \] (6)

where \( A \) is winter accumulation (cm SWE) and \( T \) is summer temperature averaged for June, July and August (°C) (Figure 9). Using this mathematical, a comparison of modern climatic conditions at paleo-equilibrium line altitudes to this window can constrain the magnitude of climate change required to “regrow” more extensive glaciers.

To estimate modern conditions at any given elevation within the Green River drainage, I used data from seven proximal SNOTEL sites. While it would be ideal to have data from a 30 year running record, the sensors have, in places, only recently been added. As such, the record at some sites date to the past 25 years, but others have a truncated record based on the availability of data for both temperature and precipitation (Figure 5). The total winter precipitation was measured as the maximum reading of Snow Water Equivalent (SWE cm H₂O) at the end of the winter. This measurement mostly occurred in either late March or early April depending on the year and late season accumulation. The summer temperature was derived from the mean monthly temperatures (June-July-August). The USFS and USGS are not actively monitoring any glaciers in the Wind River Range therefore there is not a SNOTEL site at or even within many kilometers of the glaciers. Using multiple SNOTEL sites from a variety of elevations and distances from Mammoth Glacier (Figure 5), I calculated both temperature and precipitation lapse. These curves enable a projection of the instrumented data to any given elevation, in this study to the ELA or paleo-ELAs of a given glacier. I did not weight the value of specific sites based on the distance from the glacier. Thus each data set is equally weighted in
creation of the lapse rates for both temperature and precipitation. However, I excluded the precipitation data from the Cold Springs SNOTEL (west side of Wind River Range) from the interpolation because orographic effects indicate that snowfall totals underestimate total snowfall and are 50-75% of sites from the eastern side of the range.

To investigate the impact of climate data from the Gunsight Pass site, which only has a 10 yr record, I processed the data twice. Initially, I used only sites with 25 yr records (thus excluding Gunsight Pass) and then I reprocessed the data using truncated records to match the available data series from Gunsight Pass (10 yr) (Figure 5). This sensitivity analysis dictated which record is more valuable for climate modeling.

Based on the climate parameters from the SNOTEL data above, I plotted the extrapolated conditions at the modern Mammoth Glacier ELA (3610 m – AAR) onto Leonard’s (1989) ELA climate envelope. This enabled me to estimate the amount of temperature depression and precipitation increase to return the site to glacial balance (ablation = accumulation).
4.0 Results

4.1 Moraines

Large bedrock-cored moraines dam both the Upper and Lower Green River Lakes (Figure 10). These end moraines are heavily forested with well-developed surface soils. The end moraines that impound the lakes loop into poorly exposed, weathered lateral moraines which steadily rise up valley before disappearing on the steep valley walls. The Green River lakes moraines are 15 km and 20 km below the modern terminus of the Mammoth Glacier respectively. The Green River Lakes moraines are the highest which span the entire Green River Valley, this observation is complimented by lateral moraine deposits on both sides of the valley. Debris aprons, cones and valley wall units have likely obscured other lateral moraines, and outwash sediment and meadows may have in-filled other full valley end moraines (Figure 10 and Appendix B).

Two major tributaries to the Green River (Tourist and Wells creeks) have end moraine sequences proximal to their confluences with the Green River Valley. The location and elevation varies: the Tourist Creek moraine is located on northeastern side of Green River Valley (2530 m); and the Wells Creek (Scott Lake and Lost Pipe Lake moraines) are located above the hanging valleys (3205 m and 3295 m respectively) impounding alpine tarns (Figure 10). The Tourist Creek moraine is 8 m high and loops around the modern flow of Tourist Creek. It consists largely of large lichen-covered boulders and has minimal grasses and shrub vegetation and little surface soil. The Scott Lake and Lost Pipe Wells moraines consist of till with no vegetation and little surface soil. The subsurface till of both these (Scott Lake and Lost Pipe Wells) moraines apparently contains enough fines to enable them to act as dams on their respective lakes (Figure 10). Despite the 700 m difference in altitude between the Tourist and Wells Creek moraines,
similarities in weathering and relative distance from cirque headwalls suggests they relate to the same event. The discordant elevations appear to reflect the sharp topographic gradient near where these glaciers terminated (Figure 10 and Appendix C).

The highest set of moraines is proximal to the modern termini of the active glaciers. The light–gray, coarse bouldery deposits are unweathered (standing out well against the light orange weathered bedrock), angular, and poorly consolidated (Figure 10). Where these deposits are thinly draped over sloped bedrock (2-10 m thick) they commonly lack fines, consisting instead of boulders – pebble gravel (Appendix B – LIA-till). The end moraines are light to dark gray in color, between 10-30 m high, lacking any vegetation and commonly dam small glacially sourced lakes containing abundant suspended rock flour. All of the moraines are over steepened and minimally weathered. At the Mammoth Glacier, 0.5 km of retreat has created a braided outwash pattern leading to a shallow lake (the braided outwash continues down valley beyond the moraine). In contrast Sourdough Glacier is directly calving into a 0.5 km² till-dammed lake, and has apparently experienced less total ice retreat (Figure 10 and Appendix C). Based on their freshness and proximity, these moraines represent the most recent advance of Mammoth Glacier and other cirque glaciers in the map area (Figure 10).

4.2 Rock Glaciers
Rock glacier deposits in the field area are steep, boulder-rich, elongate when constricted by valley walls and broader where occupying debris slopes (Figure 10 and Appendix C). They are commonly located on the steep northern side valleys of major tributaries to the Green River Valley (Figure 10). The rock glaciers can also occur on slopes close to active glaciers where slope and aspect are less influential. They occur at elevations higher that 3050 m; however, elevation
is less a factor than shading from solar radiation. The top surfaces of these deposits are often steep (25-30°) with a thin veneer of soil and windblown or weathered fines locally covering surficial boulders and cobbles. The rock-glacier termini are abrupt with a 35-39 degree slope, locally oversteepened, in which boulders are supported by a fine matrix of poorly consolidated silt and sand (Figure 10).

Commonly the rock glaciers are long and thin in map view but spread out at the termini to become bulbous. The longitudinal surface profiles are usually concave in the accumulation area and convex in the wider ablation zone. The rock glaciers in the study area lack the deflated hummocky and concave areas behind the toe. None of the rock glaciers are observed to have surface runoff and therefore do not carry any suspended sediment (rock flour).

**4.3 Core Analyses**

In the summer of 2008 I recovered twelve sediment cores from both the Upper and Lower Green River lakes (Figure 11, Figure 12, Figure 13, Figure 14, Table 1, and Appendix E). Because of budget and time constraints, I focused my analyses on three of the best preserved and most detailed cores.

**4.3.1 Lower Green River Lake LGL-1**

**4.3.1.1 Site Description**

I successfully retrieved one Glew core from the Lower Green River Lake (LGL-1). The core was situated along the north side of the lake, 500 m from the pour point where the Green River spills around the end moraine and continues down valley (Figure 12 and Figure 13). Based on the bathymetry, the bottom surface of the lake is sloped ~ 5-10 southward, toward the center of
the basin. The water depth was 25 m, close to the upper limit of the Glew core’s ability to successfully retrieve a core from a floating platform (Figure 12, Figure 13, and Appendix E). The true bottom or deepest portion of the primary basin of the Lower Green River Lake is located 1.5 km to the south from where LGL-1 was retrieved. The overall depth of the lake (80 m) dictated that coring the deepest portion of the deep basin was unfeasible due to the limitations of the Glew core. For LGL-1, the corer penetrated 80 cm into the lake-bottom sediments, but only 67 cm of core was recovered.

4.3.1.2 Magnetic Susceptibility (MS)
I measured magnetic susceptibility (MS) in the field after returning Glew core LGL-1 to dry ground. MS values in LGL-1 vary between 0 and 3 \* 10^{-6} SI units (linear drift applied); except for three significant peaks centered at 3 cm (12 \* 10^{-6} SI), 24 cm (143 \* 10^{-6} SI) and 62 cm (22 \* 10^{-6} SI) (Figure 15). The lowest peak is truncated by the MS measurements at core bottom and thus the peak is likely not actually centered at 62cm, rather it would continue with depth if a deeper and more complete core record had been retrieved. The MS peak at 24 cm corresponds to a depositional transition which is reflected in all of the other proxy records (Figure 15 and Figure 16).

4.3.1.3 Visual Stratigraphy (VS)
The visual stratigraphy (VS) was poorly expressed in the core barrel (Figure 15), and therefore was characterized during extraction. Rust brown sediments at the top of the core gradually transition, distinguished clearly at 17 cm as a dark brown color. A sharp contact occurs at 23 cm where sediment becomes light gray green. Sediments between 22-24 cm contained scattered larger grains (0.5 mm-2 mm) supported by a finer grained matrix of clay and silt size sediments.
The color contact at 23 cm is also reflected in the MS values. At 40 cm the sediments gradually grade into a light, yellow-gray becoming more prominently yellow at the base of the core (Figure 15). The yellowing toward the core bottom appears to be associated with the clastic sediment pulse that is truncated by the MS meter.

The core lacked any discernable tephras or lamination; however, smearing along interior portion of the clear plastic core barrel possibly obscured larger scale features (Figure 15). Macrofossils (e.g., pine needles, wood fragments and root fragments) were present throughout the core and retrieved for $^{14}$C dating (Table 2).

4.3.1.4 Organic Content (OC)

There are two main portions to the organic content in LGL-1. Above the horizon at 22-24 cm OC values vary between 8 and 12 % with a minor peak centered at 10 cm (12%) (Figure 15). No troughs were present in the OC readings. Below the horizon at 22- 24 cm, OC records decrease to 2 % with little variation from 24 cm down to the bottom of the core (67 cm) (Figure 15). The transition between the two portions of the core is sharp. The separate portions of the OC data, with a transition at 22-24 cm are consistent with the variations in VS and MS data (Figure 15 and Figure 16).

4.3.1.5 Grain Size Distribution

There are three distinct portions to the grain size distribution (GSD) in LGL-1. The top section of the core from 0-22 cm has approximately 20% clay, 30-35% fine silt, 35-40% medium silt and 10-15% fine sand (Figure 16). Sediments at 22 and 23 cm mark a transition where the total clay, fine and medium silt sum to 35%, whereas medium to coarse sand fraction increase from 0% (above) to 40%. In addition, prior to running this sample on the Malvern, I removed multiple
coarser sand – 2 mm pebble inclusions – because the pump on the machine cannot process grains >1 mm. The sand pebble mixture was composed of sub-rounded to sub angular fragments of limestone and minor fraction of granitic clasts with feldspar grains. As such the distribution in the core does not completely reflect the actual coarseness of the sediment at 22-23 cm (Figure 16). From 23-36 cm the clay fraction increases to 50% of the GSD. It spikes at 25 cm becoming 80% of the total GSD (Figure 16). Below 36 cm the GSD break down is such: 20-25% clay, 55-60% fine silt, 20-25% medium silt and 0-5% fine sand. Again, similar to the MS, VS and OC proxy records, the horizon at 22-24 cm is a transition between depositional environments, recorded in the GSD.

4.3.1.6 Radiocarbon Dating
I recovered six potentially viable macro fossil samples from the core (Table 2). Of these samples three were prepared for $^{14}$C dating at Lawrence Livermore National Labs and 2 of the macrofossils returned dates. These two samples are at 22 cm and 24 cm depth, indicating ages of 3000 – 3160 cal. yr B.P. and 3060-3160 cal. yr B.P. respectively (Figure 17 and Table 6).

The depth-age reconstruction for LGL-1 (Figure 17) relies on these two $^{14}$C dates. Because the two dates are analytically indistinguishable at 1-sigma and are separated by only 2-cm in the core, they are treated as a single age. Thus 22-24 cm is considered to be 3110 cal. yr B.P. I use a linear interpolation between an assumed age of 0 B.P. at the top of the core and 3110yr B.P. The age model extrapolates below 24 cm to the bottom of the core assuming the same linear gradient (Figure 17). The overlap of the dates and the scattered large sand grains between 22-24 cm appear to represent a period of rapid deposition, not seen elsewhere in LGL-1. The layer may represent singular high energy slope wash event that reworked macrofossils from near
surface sediments surrounding the Lower Green River Lake. Because the sample site is on a sloped surface and only 200 m from the northwest shore, it is possible that a laminar flow entrained macrofossils, contaminating the core.

**4.3.1.7 LGL-1 Chronology**

Using the depth-age model, I transposed the MS, OC and GSD from depth into calibrated ages (Figure 18 and Figure 19). Based on this interpretation, the core may represent 9000 cal. yr B.P. of sedimentation in Lower Green River Lake. At 3100 cal. yr B.P. there is a significant event in the Lower Green River Lake manifest in all proxy records. The peak in MS which builds slowly from 5000 cal. yr B.P., coincides with an increase in clay and fine silt (Figure 18). The peak in MS at 3100 cal. yr B.P. corresponds with a coarse clastic horizon of medium sand to small pebbles. This peak coincides with an abrupt increase in OC, which jumps from 2% to 8-12% (Figure 18). The increase in OC is accompanied by a decrease in MS to near zero.

At ~300 cal. yr B.P. there are subtle apparent changes in all proxies (MS peak, OC trough and GSD fining) (Figure 18 and Figure 19). However, this shift is only recorded by one or two measurements in each proxy, making the significance unclear. Furthermore, the lack of better age control on this core means all dates in LGL-1 are tenuous at best.

**4.3.2 Upper Green River Lake UGL-5**

**4.3.2.1 Site Description**

The coring work completed in the summer of 2008 recovered eight Glew cores from the Upper Green River Lake. Glew cores UGL-1, 2, and 3 were collected from a small sub-basin (up to 5 m deep) in the central-northern portion of the lake (Figure 11 and Figure 14). The majority of the
coring (UGL-4, 5, 7, and 8 - Table 1; Figure 14) focused on the central basin which had a maximum depth of 30 m and likely preserved the longest most detailed record of deposition. This main basin is steep sided on its western margin, likely bedrock controlled, with talus at the shoreline (and likely below the surface). The eastern margin has a gentle 5-10 slope from the shoreline to the deepest portion of the basin. One core in the northernmost portion of the Upper Green River Lake (UGL-3) was collected at 1 m in a flat bottomed very shallow portion of the lake (Figure 11 and Figure 14).

The multiple Glew cores centered on the main basin indicated no substantial variation in sediment thickness across the basin. Based on the proximity to the Green River input and the aggrading delta, I believed the southern portion of the Upper Green River Lake would likely have a thicker subsurface sedimentary record.

4.3.2.2 Magnetic Susceptibility (MS)

I completed magnetic susceptibility (MS) measurement on all eight Glew cores from Upper Green River Lake (Appendix A). The cores recovered at ~5 m in a NW sub basin of the lake (UGL-1, 2, and 3) captured a small increase in magnetic susceptibility (MS) centered at ~12 cm, this peak is consistent and correlative within this sub basin (Appendix A). The cores recovered from the deepest (17 -26 m water depth) central basin UGL-4, 5, 7, and 8 all follow the same MS trend (Appendix A). Based on this field MS analysis, I chose to further investigate UGL-5 which has the longest recovery and was best suited for complete analysis (Figure 20).

The magnetic susceptibility (MS) in UGL-5 has a prominent peak which is the major feature in the record. The peak reaches its maximum at 42 cm depth (183 * 10^6 SI units) (Figure 20). The upper limb of this peak, toward the top of the core, has gentle slope as MS values decrease
linearly to depth 0 cm ($10 \times 10^{-6}$ SI). Whereas the lower limb, toward the bottom of the core, is characterized by a sharp $100 \times 10^{-6}$ SI drop in only 8 cm of core (43 – 50 cm). On the upper limb there appears to be a secondary spike at 34 cm; while it only encompasses 2 data points it is also present in UGL-4 (Figure 20 and Appendix A). At the very top and bottom UGL-5 MS values trend to zero as a function of measurement capabilities (Figure 20).

4.3.2.3 Visual Stratigraphy (VS)

The visual stratigraphy (VS) of the Glew cores was poorly discernable while the sediment remained in the core barrel (Figure 20), so I characterized the sediment after extraction. The core is capped with a loosely consolidated light, low-density, brown to brown-gray 3 cm layer, with active bioturbation (live red worms noted). The color and degree of compaction shifted to being grayer and moderately compact by 5-10 cm, grading into a light gray greenish color by 30 to 45 cm depth. The color sharply transitioned back to a browner gray at 50 cm. This color shift at 50 cm corresponds with the magnetic susceptibility drop from 43-50 cm (Figure 20). From this point to the bottom of the core there is minimal color variation, it remains brown-gray.

The core lacked any discernable tephras, larger wood fragments or lamination; however the smearing along interior portion of the clear plastic core barrel possibly obscured such finer features. This smearing makes the color transition from gray green to brown gray appear ~10 cm below the actual transition (Figure 20). Macrofossils (e.g., pine needles, wood fragments and root fragments) were present throughout the core (Table 2).

4.3.2.4 Organic Content (OC)

There is no clear trend or partitioning in the organic content (OC) for UGL-5 (Figure 20). Two troughs occur at 56 cm and 35 cm, with one peak at 65 cm. The OC values rapidly transition
between 65 cm and 56 cm (peak - trough; 16% - 1%). However, the trough at 56 cm represents only one data point. Of interest is the drop in OC values between 52 cm and 35 cm, which corresponds with the MS peak growth from 43-50 cm. From a low of 2% at 35 cm, OC values steadily increase to 5% at the top of the core (Figure 20). Overall, at the top of the core OC values tend to be lower, on the whole, than values recorded at the bottom of the core.

4.3.2.5 Grain Size Distribution (GSD)
UGL-5 has minor variations, but no major trends or partitions in the grain size distribution (GSD). The average GSD for UGL-5 is: 10% clay, 30-35% fine silt, 40-45 % medium silt, 20% fine sand and 0-5% coarse sand (Figure 21). Unlike LGL-1, no clasts larger than coarse sand were removed prior to analysis on the Malvern. There are two major spikes in fines within UGL-5. At 68 cm the GSD contains up to 95% clay (Figure 21). Between 42 and 44 cm the clay and fine silt fraction increase to comprise 80% of the total GSD (Figure 21). The second spike in fines corresponds to the maximum MS spike centered at 42 cm. Additionally, the grain size distribution at 50 cm, the base of the MS peak, also shows a dramatic decrease in coarse faction (medium and coarse sand) and an increase in fines (up to 70%) (Figure 21). A coarse layer is noted at 15 cm depth where clay and fine silt diminishes to 10% of bulk composition and the coarse sand fraction increase to 20%. While the GSD can be affected by single point measurements, the peak in fines mimics that of peak MS values and the troughs in OC.

4.3.2.6 Radiocarbon Dating
I extracted four macrofossil samples from Glew core UGL-5 (Table 2). Of these samples three were prepared for AMS $^{14}$C dating at Lawrence Livermore National Lab and two of the macrofossils returned dates. These two dates are located at 36 cm and 66 cm and have
calibrated ages of 5-155 cal. yr B.P. and 10-150 cal. yr B.P. (Table 5). These were eliminated as resolution tends to be poor with calibrations of $^{14}$C younger than 200 cal. yr B.P.

4.3.2.7 Lead Dating

Because UGL-5 lacked suitable radiocarbon dating controls, I applied $^{210}$Pb dating to the Glew core to establish a numeric chronology. Secular equilibrium in UGL-5 was reached at 9.5 cm depth (Figure 22). The resulting linear interpretation between samples at 1.5, 2.5, 4.5, 6.5 and 9.5 cm indicate that initial excess levels at depth 0 ($A_0$) was 23.5 (dpm/g). Using equation 4, I calculated dates for each sample I processed (Figure 22). However, I chose to include only three of these dates in the depth age model. I excluded the dates which were below secular equilibrium (>9.5 cm). Of the five remaining dates I use the dates at 2.5, 6.5 and 9.5 cm. I excluded the date at 1.5 cm as it was very young (1 yr.) and so close to the active bioturbation; the date at 4.5 cm was synchronous with the date at 6.5 cm (43 and 45 yr. B.P), thus I chose to include only one of these dates (6.5 cm and 45 yr. B.P). I feel these selections serve to accurately representing the uppermost sediment.

I integrated $^{210}$Pb dates with a $^{14}$C date to construct the depth-age model for UGL-5. To incorporate age control below 9.5 cm, the deepest date obtained with the $^{210}$Pb method, I incorporated a $^{14}$C date from Livingston core UGL-LIV5 (Figure 23). To accomplish this transformation, I correlated the two cores from Upper Green River Lake (UGL-LIV5 and UGL-5), based on thickness and relative position of the Little Ice Age MS peak. The MS peak in UGL-5 was relatively broad, spanning 40 cm, whereas the MS peak in UGL-LIV5 spanned 25 cm. Because UGL-5 is collected closer to the deepest portion of the main basin in Upper Green River Lake it would be reasonable to expect it to have a higher sedimentation. Using the ratio of LIA
sediment thickness between cores, I transferred a date from UGL-LIV5 to UGL-5; the date is located at the base or initiation of the LIA MS peak. Based on this correlation, the calibrated date at 38 cm depth in UGL-LIV5 (430-490 cal. yr B.P.) is equivalent to 460 cal. yr B.P. at 67 cm depth in UGL-5 (Figure 23, Table 5, and Table 6).

**4.3.2.8 UGL-5 Chronology**

Using the depth-age model (Figure 23), I transposed the MS, OC and GSD from depth into calibrated ages (Figure 24 and Figure 25). Core UGL-5 represents ~550 yr B.P. of sedimentation in Upper Green River Lake. From 450 to 300 yr B.P. the MS values increase (40 *10^-6 SI – 185 *10^-6 SI) to a peak at 280 yr B.P (Figure 24). Also commencing at 450 yr B.P. is a rapid decrease in OC from 15% to 2% by 400 yr B.P. The peak in MS at 280 yr B.P. coincides with fines pulse in GSD. The trough in OC that occurs at ~250 yr B.P., corresponds with a minor peak in MS at the same time (Figure 24). At about 100 yr B.P. there is coarse pulse, which is only reflected in GSD (Figure 24 and Figure 25). At time zero the MS values are equivalent to the values recorded before the major peak in MS (550-450 cal. yr B.P.).

**4.3.3 Upper Green River Lake UGL-LIV5**

**4.3.3.1 Site Description**

I recovered six Livingston cores from the Upper Green River Lake. UGL-LIV 1, 4 and 6 were collected from the small sub-basin (up to 5 m deep) in the central-northern portion of the lake, proximal to Glew cores UGL-1, 2, and 3 (Figure 11 and Figure 14). UGL-LIV 2, 3 and 5 were collected from the eastern margin of the main-central basin, similar to Glew cores UGL- 4, 5 and 7 (Figure 14). This main basin is steep sided on its western margin, likely bedrock controlled with minor rock fall talus at the shore line (and below the surface). The eastern margin has a
gentle 5-10 slope consistently from the shoreline to the deepest portion of the basin.

However, the depth restriction of Livingston coring dictated the sites be away from the center of the basin towards the north and east. Cores UGL-LIV 1,2, and 3 were only 2 x 100 cm. UGL-LIV4 was extracted without the appropriate plug; as such the core was lost upon extraction. UGL-LIV6 represents the greatest depth of sediment recovered from the Upper Green River Lake (5 x 100 cm pushes with minimal overlap), but was only recovered from 5 m depth in the northern sub basin (Figure 11 and Figure 14). The core retrieved from the deepest portion of the lake was UGL-LIV5, collected from 10 m water depth and recovered 4 *100 cm of minimally overlapping core (Figure 26).

4.3.3.2 Magnetic Susceptibility (MS)

I processed the magnetic susceptibility (MS) after returning the cores to WWU. I completed MS measurement on the remaining five Livingston cores from Upper Green River Lake (UGL-LIV4 excluded). All but UGL-LIV5 had minimal variation in MS with values ranging between 0 – 10 * 10^{-6} SI and no major peaks. As such, I focused my analytical efforts on UGL-LIV5. The proxy records from UGL-LIV5 are best analyzed as one continuous core. To enable this process, I reconstructed the four separate 100 cm pushes into one complete 364 cm core. In total 36 cm of overlap within the 400 cm of core was determined from measurements made in the field during the coring process (Figure 26). Using these field measurements, in conjunction with MS, VS and OC allowed the continuous core to be investigated.

The MS values recorded in UGL_LIV5 range between 10 and 25 *10^{-6} SI units with the exception of a large peak, which reaches its maximum MS value (166 *10^{-6} SI) at 24 cm depth (Figure 27). The lower limb (toward the bottom of the core) of the largest MS peak (centered at 24 cm) is
sharply truncated at its base where MS vales return to background values (~25 *10^{-6} SI) by 32 cm depth (Figure 27). The upper limb (toward the top of the core) linearly returns to 25 *10^{-6} SI at 0 cm depth, with one secondary peak occurring at 20 cm (Figure 27). There are two additional minor peaks one at 130 cm (23 *10^{-6} SI [within modal core range of MS values]) and one at 346 cm (34 *10^{-6} SI).

4.3.3.3 Visual Stratigraphy (VS)

The visual stratigraphy (VS) for all Livingston cores other than UGL-LIV5 was remarkably consistent. They had massive, brown-gray brown, organic rich muds, lacking any visual laminations or stratigraphy. The lack of evidence for glacially derived sediment in all but UGL-LIV 5 focused the subsequent analysis exclusively on UGL-LIV5 (Figure 26).

The VS of UGL-LIV5 is intricate in the upper 150 cm whereas the bottom 215 cm is more uniform (Figure 26). The uppermost meter of UGL-LIV5-1 (0-100 cm) there are two fine (<2 mm) gray to light-gray laminations (5 and 7 cm deep), within predominately reddish, buff-brown sediment (Figure 26). At 19 cm depth, a 2 mm thick lamination of alternating light gray and dark brown sediment caps a 5 cm thick light gray band. The top of this light gray band has a sharp transitioning back to the light brown colored sediments (Figure 26). The bottom contact of the 5 cm light gray band is more diffuse; however, the brown sediments return within 2 cm of its base. Below this band, at 30 cm depth, is a light-tan to gray, wavy lamination of ~4 mm. From 30 cm to 100 cm, the rest of the top push, the core consists of light brown sediments without any noticeable lamination or major color variations. In the second push 100 – 200 cm there was only one light gray lamination at 30 cm depth of ~ 2mm (Figure 26). From this depth down to
200 cm the sediment was uniform, brown, non-layered and lacking any rock flour laminations in the second push (UGL_LIV5-2).

The basal two pushes UGL-LIV5-3 and UGL-LIV5-4 both appear visually similar to the bottom of UGL-LIV5-2 (Figure 26). They have no laminations or beds and are brown organic homogeneous sediment, similar to the unanalyzed Livingston cores from the Upper Green River Lake (UGL-LIV-1, 2, 3, and 6). Macrofossils (e.g., pine needles, wood fragments and root fragments) were present throughout the core (Table 3 and Table 4).

4.3.3.4 Organic Content (OC)
UGL-LIV5 has little variability in organic content (OC). The OC values within the core have a baseline that ranges between 7 and 9 % (Figure 27). A trough occurs at 22 cm depth where OC drops to 2%. This trough in OC negatively correlates with the maximum MS peak located at 24 cm depth. The trough at 22 cm has a sharp transition on its lower limb (bottom of the core) and a linear return to baseline (7% OC) on the upper limb (top of core). This pattern is also similar to the MS pattern with a sharp base and a gentle top. The only other variation within the core occurs at a peak centered on 200 cm depth where values reach 25 % OC (Figure 27). This peak is 10 cm wide with a sharp tail above and below, as values quickly return to baseline. There is no variation in either MS or grain size distribution (GSD) at this depth. Upon further investigation, cross push correlation (UGL-LIV5-2 bottom and UGL-LIV5-3 top) and rerunning the samples, this peak appears to be accurately located (Figure 27).

4.3.3.5 Grain Size Distribution (GSD)
The grain size distribution (GSD) has little variation throughout, UGL-LIV5. GSD generally displays 10% clay, 30 - 40% fine silt, 40 - 50% medium silt and 10 - 20 % fine sand with no
coarser sediments (Figure 28). Unlike LGL-1, there are no clasts larger than coarse sand. The only variation occurs at 99 cm depth where the clay fraction increases to 70%. Because this finding is based on a single sample, sediment at 99 cm was analyzed petrographically on a smear slide to better characterize the tri-modal distribution indicated by particle-size analysis. There were a few diatoms still present after grain-size preprocessing, but the sample was primarily very fine, sub-angular silt with minor angular coarse sand. The spike in fine component in the standard GSD analysis thus appears to be accurate but anomalous.

4.3.3.6 Radiocarbon Dating
I successfully recovered twenty macro fossil samples from UGL-LIV5 (Figure 26, Table 3 and Table 4). Of these samples eight were prepared for $^{14}$C dating and seven of the macrofossils returned dates ranging from 460 – 4190 cal. yr B.P. (Table 5 and Table 6). I used six of these dates in the creation of the depth age model (Figure 29). I excluded sample ND_UGL-LIV5-3-14 (3450 -3590 cal. yr B.P.) (Table 6) which was sampled from a section of core interpreted to represent either overlapping and repetitive coring or hole-sluff. This interpretation seems convenient, removing a date that ruins the continuity of the depth-age model, as ND_UGL-LIV5-4-26 (2010 - 2190 cal. yr B.P.) is stratigraphically below ND_UGL-LIV5-3-14 (3450 -3590 cal. yr B.P.). To justify this omission, I made multiple depth-age plots choosing to omit various points and settled on this configuration which increased the rate of deposition during the period of glacial activity and incorporated most of the radiometric dates (Figure 29, Table 5, and Table 6). Another explanation could be that the sample, ND_UGL-LIV5-3-14, represents an older detrital sample that was reworked, resulting in the contamination. As this problem affects all samples, in all cores, it is less favorable an explanation.
4.3.3.7 UGL-LIV5 Chronology

Using the depth-age model (Figure 29) I transposed the MS, OC and GSD from depth into calibrated ages (Figure 30 and Figure 31). The MS values vary between 10 -25 from 4500 yr B.P. till present. While the MS peaks at 3900 and 1250 yrs B.P. are minimal, they indicate a moderate rate of clastic sedimentation throughout the core (Figure 30). The major peak in MS occurs at 280 yrs B.P., with a secondary peak occurring at 240 yr B.P. The sedimentary history for Upper Green River Lake is consistent, both UGL-5 and ULG-LIV5 have major MS peak and an OC trough at 280 yr B.P. and secondary MS peaks between 260-240 yr B.P. (Figure 24 and Figure 30). In UGL-5, which was retrieved from a deeper basin, the MS peak is broader and overlain by more subsequent deposition than in UGL-LIV5 (Figure 30).

The large peak in OC that occurs in UGL-LIV5 at 1600 yr B.P. is associated with the core boundary interface (Figure 30 and Figure 31). However, since the sloughing top sedimentary material is very low in organic content, the top 500 yr B.P. averages ~ 5% OC, this peak might not be an artifact. While measurement error is always a possibility, another explanation would be a local fire event; this would increase the flux of organic material to account for the jump to 25 % OC. The visual stratigraphy does not show any evidence of darker organic rich charcoal deposits, which I investigated carefully for possible material for radiocarbon dating. Additionally, there is minimal variation in MS values at this date; however, OC and MS values are not necessarily correlative, especially in non-glacial deposits (Figure 30 and Figure 31).

4.4 Scott Lake

The nature and characteristics of Scott Lake and associated deposits are crucial to reconstructing the Holocene climate history of the basin. The lake has numerous features that
are characteristic of a recently drained lake. One obvious feature is a distinct “bathtub ring” of oxidized outcrops that coincide with a distinct change in lichen development. The uniform heavy iron staining is exposed on bedrock 10 m above present lake level, in places bedrock is covered by alluvial fan deposit (Figure 33). These alluvial deposits have an inflection in slope 10 m above the modern shore (Figure 32 and Figure 33). Coinciding with the level of this change is a stranded Gilbert delta at the main inlet to the lake. The sub-rounded boulder to cobble gravel in the stranded delta is much coarser than the bed load capacity of the active modern stream feeding the lake (Figure 32, Figure 33 and Figure 34). The modern creek entrains a coarse sand to silt fraction in addition to suspended rock flour, as it grades into a modern Gilbert delta in Scott Lakes (Appendix E).

The bathymetry of Scott Lake reaches depths of 25 m, with one major basin centrally located (Appendix E). The modern lake is dominantly a bedrock tarn, occupying an over-deepened, Pleistocene, glacially carved depression. At the outlet from the modern lake there is a 1-2 m cascading waterfall (Figure 34), which dissecta a moraine and drops to a small narrow, bedrock-sided lake. This narrow lake is dammed by alluvium and rock fall (Figure 32). The high water mark of the main lake is continuous to the outlet of this smaller lake, indicating that the incised sediment at the outlet, possibly reworked moraine material, served as a dam for a single, deeper lake. The moraine dam (Figure 10 –Scott Lake moraine) that partitions the modern lakes remains mostly intact (Figure 32).

The area of modern Scott Lake is 0.12 km² (Figure 32 and Table 7). From the geomorphic evidence above, the area of Scott Lake during the highstand represented by the paleo-shoreline
was 0.25 km$^2$ or twice the modern area. The volume of water represented by the difference is ~2,000,000 m$^3$ (Table 7).

### 4.5 Glacial Reconstructions

In order to constrain the magnitude of climate change from late-Pleistocene till present, including the smaller fluctuation within the Green River Lakes drainage, I modeled the glaciers’ past extents from the geomorphic mapping. These glacial reconstructions allow me to establish paleo and modern equilibrium line altitudes (ELAs) using several methods. I then established the best estimate for modern climate (summer temperature and winter precipitation) at Mammoth Glacier. Using the modern climatic conditions at Mammoth Glacier, I extrapolated to the paleo glacial ELA’s using the modern lapse rates for both temperature and precipitation. This extrapolation enabled an amount of temperature depression and precipitation increase to be measured.

#### 4.5.1 Valley Glaciers

Using the moraine deposits in the Green River drainage, I assessed equilibrium line altitudes (ELA) using the mean altitude (MA) and toe headwall altitude ratio (0.6) (THAR) (Figure 10 and Table 8). The ELA data groups into three unique altitudes for paleo-glaciers (Table 8). The glaciers of the low broad valley group at 3100 m (THAR). These large valley glaciers, with moraine deposits spanning the valley width, were fed from glaciers that covered the entire catchment, which would include the Tourist Creek, Pixely Creek, and Wells Creek drainages and those on the southwest side of the Green River Valley. The glacier that deposited the Tourist Creek moraine falls into this range (3100 m) based on the moraines elevation at the terminus of Tourist Creek, where it meets the Green River. The glaciers responsible for the moraines at Scott
Lake, Lost Pipe Lake and Inner-Titcomb Lake are grouped at 3525 m (THAR – ELA; Figure 10 and Table 8). All of these deposits are similar in size and the lakes they impound are similar in area. The gray rubble moraines proximal (~1 km) to the present glaciers has an ELA increase of 100 m above the three moraines coalesced at 3525 m (3625 m) (THAR – ELA; Figure 10 and Table 8). To further refine the ELAs from these highest moraines I performed a detailed ELA analysis in the Wells Creek drainage.

### 4.5.2 Cirque Glaciers

I completed a detailed analysis of the glaciers in the Wells Creek drainage. I used the more rigorous area altitude ratio (AAR) method (Figure 35, Table 7 and Appendix D). Based on AAR the ELA has risen from 3575 m to 3610 m from Little Ice Age (LIA) maximum (1850 A.D.) to present (firn line-2002) (Figure 35 and Table 9). While all three glaciers in the Wells Creek drainage have retreated since LIA maximum, Mammoth Glacier retreat rate, based on area, is twice that of either Baby or Minor Glacier (Appendix D). All of these glaciers are also experiencing thinning and thus volume reduction.

### 4.5.3 Temperature and Precipitation

To establish the modern climatic variables at the present Mammoth Glacier equilibrium line altitude I used the regional SNOTEL data to construct temperature and precipitation lapse rates (Figure 36, Figure 37, Figure 38, Figure 39, and Table 10). As the data from Gunsight Pass has a limited period (1999-2010), I want to assess the importance of incorporating data from close to the field area (Figure 5) versus an increased duration of the SNOTEL record (Table 10). As such, two lapse rates were constructed; one with a 25 year data set (1986–2010) and a second, which included Gunsight Pass SNOTEL site, but only spans a 10 year period (1999-2010)(Table 10).
Temperature lapse rates based on the 25 yr and 10 yr data sets are -1.6 and -1.5 °C/km, respectively (Figure 36 and Figure 38). Extrapolating these to the modern ELA of the Mammoth Glacier (3610 m) indicates mean summer temperatures of 9.98 °C and 10.52 °C for the 25 yr and 10 yr extrapolations, respectively. The $R^2$ value for both regressions is only 0.11, indicating the original lapse rates are not reliable for a study of this nature (Figure 36 and Figure 38).

Precipitation (SWE) lapse rates for both the 25 yr and 10 yr data sets are 1.7 and 8.6 cm SWE/km, respectively (Figure 37 and Figure 39). I excluded the SNOTEL site at Cold Springs (used in Naftz, 1996; Schuster et. al. 2000) because it is located on the northeastern flank of the range. Due to the orographic effect, precipitation totals at Cold Springs are between 50 to 75% less than the sites from the southwestern flank of the Wind River Range, thus including this site would underestimate the actual winter snowfall at Mammoth Glacier. From the extrapolations using these lapse rates the winter precipitation at the modern ELA of the Mammoth Glacier (3610 m) would be 33.98 cm SWE and 37.98 cm SWE for the 25 yr and 10 yr extrapolations, respectively. In both cases, however, the fit of the lapse-rate regressions is poor. The $R^2$ value is only 0.02 for the 25 yr average; the $R^2$ is 0.25 for the 10 yr average which includes Gunsight Pass data (Figure 37 and Figure 39).

Using the coldest (9.98 °C - 25 yr record) and wettest (38.64 cm SWE – 10 yr record) values of these lapse rates, the ELA climate conditions plot well outside the stable climatic ELA window of modern glaciers spanning a wide range of conditions (Leonard, 1989) (Figure 40). Based on the poor $R^2$ values this finding is not unexpected.
To better constrain the modern climate at Mammoth Glacier I used modeled PRISM data, (PRISM, 2011). PRISM allows for a climate reconstruction focused on a single 800 m grid, which I centered on Mammoth Glacier (lat. -109.66; long 46.16; average elevation 3850 m). The standard 30 year normal of modern climate is 1971-2000; a data set which PRISM has as a standard default (4.8 °C and 75.74 cm SWE). To compare the PRISM data over a similar period as the SNOTEL data (1985-2010) I also found the average of both winter precipitation and summer temperature from 1981 to 2010 (7.16 °C and 63.09 cm SWE) (PRISM, 2011). These two 30 year averages still indicated, based on PRISM data, the modern glacier was out of balance (Figure 40). One final PRISM modeled data set for a 115 yr period, from 1895 to 2010 (6.27 °C and 63.14 cm SWE), again centered on the Mammoth Glacier, continued to indicate that the modern glacier was out of balance based on Leonard’s climatic window (Figure 40).

To better constrain the climatic condition at the modern Mammoth Glacier, I applied temperature and precipitation lapse rates from Brugger and Goldstein’s 1999 work in the Sawatch Range in Colorado. I pinned the starting points, for both temperature and precipitation (25 yr record), to the elevation of their specified SNOTEL site and extrapolated to the modern Mammoth Glacier ELA (Table 11 and Table 12). I then averaged the results from the SNOTEL sites to incorporate the strength of locally sourced data. Using the averaged results, the coldest (western central CO – 8.44 °C/km) and wettest environment (Sawatch Range - 37.4 cm/km) indicated the modern climatic conditions were 3.78 °C and 68.48 cm SWE (Table 11 and Table 12). Using these coldest and wettest conditions the modern glacier is closer (1.5 °C or 40 cm SWE) to being in balance (Figure 40).
5.0 Discussion

5.1 Upper Green River Drainage Geomorphic Deposits

The surficial deposits in the upper Green River drainage, above Lower Green River Lake are primarily related to the valley’s glacial history and mass wasting processes. The Pinedale and Bull Lake moraine sequences in the Green River region extend well outside the field of study more than 50 km down valley from present day Mammoth Glacier. The two Green River lakes are dammed by bedrock-cored recessional moraines from the Pinedale Glaciation. The size and continuity of moraines from this LGM stand suggest that the glaciers were at these extents for an extensive period (> 1000 yr). Based on the Pinedale maximum chronology established by Hall and Shroba (1995) and Phillips et al. (1997) these moraines at the Upper and Lower Green River lakes should date to younger than 16 Ka. Since deglaciation, the valley flanks above Upper Green River Lake have collected debris via mass wasting including debris avalanches, fans and aprons. Colluvium deposits throughout the valley are controlled by the high relief valley walls which grade into the gently meandering Green River.

The moraine at the terminus of Tourist Creek is likely a late-glacial (Younger-Dryas?) advance or still stand. Although the elevation of the Tourist Creek terminal moraine is only 50 m higher the Upper and Lower Green River lakes moraines, the size of the glacier would be an order of magnitude smaller in area and volume (Figure 10 and Table 8). Thus it is likely that the Tourist Creek moraine is the same advance that deposited the till dams at Scott Lake, Inner-Titcomb Lake and Lost Pipe Lake. These advances or still stands (Scott Lake, Inner-Titcomb Lake and Lost Pipe Lake) are correlated based on the distance from the headwall, moraine characteristic and ELA’s to be late-Glacial (Younger Dryas?, 11.2 ka)(Gosse, 1995a and 1995b). The fact that these
moraines are between 650 and 700 m higher than the Tourist Creek moraine, would indicate that the late-Glacial (YD?) advance or still stand was larger in Tourist Creek than other valleys (Wells, Elbow and Pixely). However, the volume of ice below the crux of hanging valley would be minimal compared with the total glacier area.

Because there is little to no post-Pleistocene debris, except in the main valley of the Green River, the field area has an abundance of bedrock promontories. As such, it is fair to assume that the glacial ice was clean and bedrock bounded while occupying the high valleys. Neoglacial readvances of Mammoth Glacier have produced considerable debris and till (which is easily distinguishable from older Pleistocene debris based on weathering characteristics), the youngest of which was the largest Neoglacial advance (Little Ice Age). While it is possible and likely that active glaciation was occurring in a limited scope within the Green River Lakes drainage during the early Neoglacial, subsequent advances during the Little Ice Age have overprinted and eroded any geomorphic evidence.

The rock glacier deposits may locally record glacial activity predating the Little Ice Age, following Giardino et al. (1984) and Konrad and Clark (1998). Down valley from the LIA till of Minor Glacier is an active rock glacier, which was interpreted by Mahaney (1993) to have an Indian Basin (~5 Ka) advance outside the Audubon (2-1 ka) advance (Figure 7). The lack of deflation, convex profile and steep front are good indicators for modern activity; however, it is not actively producing rock flour (Figure 10 and Appendix B). As this rock glacier in contiguous, lacking deflated margins, I interpreted the Little Ice Age advance to be the largest with in the Neoglacial. It would be difficult to accumulate that amount of rubble and talus during only the LIA, as such I believe the rock glacier proximal to Minor Glacier existed in some form during the
early Neoglacial. Based on solar shielding, high elevation (3500 m), increase precipitation from wind-blown accumulation and relative position (close to Minor Glacier), I believe it grew from a smaller earlier Neoglacial configuration to its LIA maximum (Figure 10, Figure 32 and Appendix B).

Mahaney (1993) indicates three separate glacial advances occurred within the Neoglacial in the Wells Creek drainages, with Indian Basin (5 ka) being the largest. In his assessment, Scott Lake is partially dammed (correlative to my Scott Lake moraine – Figure 10) with till from the Indian Basin advances, whereas I would reassess this moraine into the late-glacial and likely the Younger Dryas advance. The other locations of Indian Basin till are questionable as well. In the valley between Scott Lake and Mammoth Glacier the orientation of Indian Basin till parallels the valley which to me indicates it is more likely a sub-glacial deposit, below a bedrock promontory, accreted during the late-glacial. Mahaney also indicates the Audubon (2-1 ka) advance is larger than the Gannett Peak (LIA) based on the moraine at Minor Glacier. I interpreted the entire LIA moraine at Minor Glacier to be uniform and synchronous. Thus, I see no evidence in the geomorphic formations and depositions for any advance associated with either the Indian Basin or Audubon advances. Mahaney (1993) properly distinguishes the Gannett Peak extent, which I interpreted to be the largest Neoglacial advance, overprinting and destroying any older Neoglacial activity within the drainage.

5.2 Green River Lakes Sediments

The Upper Green River Lake’s distinctive pea green color reflects the abundant suspended sediment, mostly clay and silt, transported down from the modern Mammoth Glacier. In contrast the dark blue color of the Lower Green River Lake signifies limited input from
suspended glacial sediments. Clear Creek, which contributes to Lower Green River Lake, contributes a minor non-glacial suspended sediment load (Figure 3). Thus, I can be confident that any glacial sedimentation entering both Upper and Lower Green River lakes is directly attributed to Mammoth Glacier.

The Little Ice Age advance established an up valley glacier which produced the peak rock flour input into the Upper Green River Lake, over the documented record. All proxy records reflect this: maximum values of magnetic susceptibility (MS); a light, gray-green sediment color in visual stratigraphy (VS); a trough in organic content (OC); and the fining of sediment size to predominately clay and fine silt (GSD) are recorded in both cores from the Upper Green River Lake. In both UGL-5 and ULG-LIV5 the LIA maximum builds from building from 450 yr B.P. and reaches its maximum at ~280 yr B.P. (Figure 24 and Figure 30), which corresponds to other regional records (Leonard, 1997; Mijal, 2008). In UGL-5, which was retrieved from a deeper water depth in the central basin, the pulse of LIA is thicker and overlain by more subsequent deposition than in UGL-LIV5. Lower Green River Lake (LGL-1) records a small increase in MS centered on ~300 yr B.P. This increase is also interpreted to be the LIA advance; however, the signal strength in the proxy records is weaker as the suspended sediment load entering the Lower Green River Lake is diminished. Based on both geomorphic evidence and radiometrically controlled core records the LIA is the Neoglacial maximum in the Wind River Range, which corresponds to other studies across the Cordillera (Leonard, 1997; Armour et al. 2002; Bowerman, 2005; Mijal, 2008). As such, it is likely that similar climatic forcings affected the entire Western Cordillera during the Little Ice Age.
Based on the proxy record in UGL-LIV5 (high MS values, relatively minor variations in GSD, and low OC value) a steady rock flour flux has entered the Upper Green River Lake during the entire core chronology. This suggests that a glacier of some magnitude has existed in the drainage from at least ~4500 yr B.P. to present (Figure 30). The glacier, of a magnitude smaller than the Little Ice Age maximum, would most likely have occupied the cirque of Mammoth Glacier. However, this interpretation is questionable because the Green River Valley, above the upper lake, has ample amounts of glacial sediment deposited after the Pinedale recession, which could be reworked to produce the rock flour signal. The absolute magnitude of the LIA advance noted in the proxy record indicates that present glacial configuration does not create a spike in clastic glacial sedimentation (rock flour). This is interesting as the Upper Green River Lake, and to a less degree the Lower Green River Lake, is visually observed in summer months to have a high flux of rock flour. This observation lends weight to the claim that glaciers occupying the Well Creek drainages for the entire Neoglacial.

The core record from the Lower Green River Lakes indicates a fundamentally different sedimentary history, one that is non-glacial. The maximum MS values in Lower Green River Lake occurs ~3000 yr B.P., likely representing slope wash, characterized by a coarse sandy interval with small cobbles and pebbles (Figure 18 and Figure 19). The coarseness of the sediments and the absence of any similar event in Upper Green River Lake indicate that this sediment is non-glacial, and instead likely relates to a local mass wasting event. At the oldest dated portion of LGL-1, which dates to ~9,000 yr B.P., there is an increase in rock flour, which might be a late-glacial rock flour pulse or a reworked tephra (the yellow color in VS could supports this second
possibility) (Menounos and Reasoner, 1997). However, the core lacks suitable distribution of radiocarbon dates to confidently associate this MS peak with the late-glacial or an older tephra.

The Wind River Range is possibly too distal to the Pacific Maritime climate to record the glacial advances at 5000 B.P (Cashman, 2005; Marcott, 2005) and 3200 B.P (Bilderback, 2004; Koch et al., 2004; Bowerman, 2005). Previous work in the Northern Rocky Mountains (Osburn and Karlstrom, 1988; Smith et al. 1995; Leonard, 1997; Wood and Smith 2004) recorded advances at 3200 B.P. While I do not have evidence for this advance, the Wind River Range physiography, especially in the Upper Green River drainage, in conjunction with magnetic susceptibility proxy records for clastic sedimentation indicate that a small cirque glacier did exist throughout the late-Holocene.

5.3 Scott Lake Evolution

The evolution of Scott Lake is unusual for an alpine lake. Based on the similarity of the geomorphic deposits to Inner-Titcomb Lake (Gosse et al., 1995a), superposition and lack of any larger Neoglacial rock flour pulses in Upper Green River Lake, I strongly believe the alluvial dam that impounded Scott Lake to its high water stand was late-glacial (YD?) and not Indian Basin (5 ka) as previous interpreted by Mahaney (1993). These factors and age assessment would indicate the high water stand of Scott Lake spanned the Holocene.

At the inlet to Scott Lake much of the deposit of coarse boulder material in the stranded deltaic system may relate to outwash from early Neoglacial advances of Mammoth Glacier. Melt water from the modern glacier does not appear competent enough to create the coarse sediment to build the delta in its present form. Modern coarse material is largely contained behind the Little
Ice Age terminal moraine, preventing it from reaching the delta. However, an outburst flood at or just after LIA maximum could create the same style deposit. The southeast portion of the stranded delta is partially buried by a debris cone at the inlet to the high water stand of Scott Lake. This debris cone is not graded to the higher lake level, thus it must coincide with or post-date the lake level lowering. The apparent lack of competence of the modern stream suggests this debris was deposited during the same event that caused the breach at the western margin of Scott Lake (Figure 32).

The breaching of the dam at Scott Lake, which appears to have been stable for >10 millennia, suggest an unusual catastrophic cause, such as an outburst flood from ice-marginal lake that flooded Scott Lake overtopping the outlet alluvial dam. An event of this magnitude would likely occur when Mammoth Glacier was at maximum LIA extent; climatic conditions would likely be associated with increased precipitation and thus melt water, ponding summer lakes behind topographic barriers emplaced by the ice margins (Figure 32). While increased summer precipitation, possibly in a single event (thunderstorm), might account for the outburst flood from Scott Lake it is likely that increased discharge from Mammoth Glacier played a role.

Without direct evidence of what caused Scott Lake to catastrophically fail, I looked to other outburst flooding research in the Wind River Range. Oswald and Wohl (2008) documented a 2003 Jökulhlaup from Grasshopper Glacier (3.2 million m³), ~7 km from Mammoth Glacier. The recognition of similar processes affecting the Wind River Range as glaciers retreat from LIA extents, serves to strengthen the glacially derived outburst story. Regardless of the mechanism, the breach of the dam rapidly lowered Scott Lake to its modern stand. The iron staining along
the walls is uniformly thick with no indication of intermediate lake level stands between maximum and present.

After breaching the Scott Lake dam 2.2 million m$^2$ of water flooded the main valley of the Green River (Figure 32 and Table 9). The Oswald and Wohl paper details zones of aggradation and accumulations in response to the flood, thus it is fair to speculate this large volume of water mobilized valley sediments depositing them into Upper Green River Lake. The flood signal is documented in the cores of the Upper Green River Lake (ULG-5 and UGL-LIV5) at ~240 yr B.P. (Figure 24 and Figure 30), but not in the record of the Lower Green River Lake. The flood is documented in UGL-LIV5 as a pulse of sedimentation defined by a 4 cm thick layer of light, green-gray rock flour noted in the visual stratigraphy (VS) and magnetic susceptibility (MS) (Figure 26 and Figure 30). Glew core, UGL-5, depicts a similar MS signal, one that is younger (~240 yr B.P.) than LIA maximum and that corresponds to fining GSD and a trough in OC accompanied by light-gray sediments (VS) (Figure 24 and Figure 25). UGL-5 records up to 8 cm of deposition from this single event; it is congruent that 2x more sediment would be deposited based on relative position within the lake. This fine grained clastic sediment likely comprises suspended rock flour from Scott Lake and reworked clastic sediment from the upper Green River Valley. It is unlikely that the flood mobilized a significant amount of bottom sediment from Scott Lake. In addition to the geomorphic evidence for a recent lowering of Scott Lake to modern lake level the core record also supports the outburst occurring just after the LIA maximum at 240 yr B.P.
5.4 Equilibrium Line Altitudes (ELA), regional and local climate implications

ELAs for the Pinedale recessional, late-glacial (Younger Dryas?), Little Ice Age and modern are in agreement throughout the Green River Lakes drainage. The 3100 m ELA for Pleistocene recessional is in agreement with Munroe et al. (2006) work from the Uinta Mountains LGM ELAs (2950 m) and Thackray et al. (2004) from the Sawtooth Range. The highest ELA, for the LIA extent of the Mammoth Glacier (3610 m AAR), is the highest altitude considered in my project for climatic conditions.

The minimal variation between climatic conditions (summer temperature and winter precipitation) at elevations between 3525 and 3610 m (late-glacial to modern) extrapolated from the lapse rates indicates that there must be variations not accounted for by the simple linear regressions of SNOTEL data (Figure 36, Figure 37, Figure 38 and Figure 39). Such regressions of modern climatic conditions at the Mammoth Glacier’s ELA (3610 m) appear to substantially underestimate total winter accumulation and overestimate summer temperatures compared to those conditions on other glaciers around the world (e.g., Leonard, 1989). One explanation for these anomalies is that the SNOTEL sites do not accurately portray conditions near the crest of the range. The SNOTEL sites in the foothills will not account for avalanching and wind-blown accumulation proximal to the cirque head wall, or decreased rates of ablation due to solar shielding. A more accurate assessment of precipitation might include a threshold process, which would treat elevations below 3400 m similarly but increase precipitation dramatically above 3400 m. This possibility would account for the non-linear local orographic and drifting effects near the crest of the range. Additionally by measuring only the maximum
SNE for the snow year, melting and re-accumulation was not included in the SNOTEL total, adding another potential source of precipitation error.

There is no clear explanation for the temperature data. Using only the SNOTEL data (Figure 5) the regression of the temperature data indicates a lapse rate of ~1.5 °C/km (Figure 36 and Figure 38). This value is far lower than the standard saturated adiabatic lapse rate of 5.8 °C/km (Danielson et al., 2003) or the dry adiabatic lapse rate of 9.8 °C/km (Danielson et al., 2003). Over time, the most likely lapse rate would fall between these two values. However, even the PRISM data at the modern ELA indicate the conditions at the Mammoth Glacier ELA are outside the Leonard climatic window (Figure 40).

To address this apparent problem, I applied steepest two Colorado lapse rates (8.4 °C/km for temperature, and 37.9 cm SWE/km for precipitation) constructed by Brugger and Goldstein (1999) (Table 11 and 12). Projections based on these lapse rates from the Wind River Range SNOTEL sites (Figure 5) suggest that conditions at the modern Mammoth Glacier ELA (1.5 °C and 40 cm SWE) are close to those indicated by Leonard’s climatic window (Figure 40).

The work compiled by Leonard (1989) has stood as the benchmark for studies of this nature for the past 25 years. However, Leonard’s interpretation does not include any glaciers from the American Rocky Mountain. Leonard notes that Tsentralny Tuyuksu in Soviet Central Asia and the Engabreen in Norway, fall significantly (ca. 1.5 °C) below the window. At Mammoth Glacier, the amount of temperature and precipitation departure from Leonard’s window is similar to those of the Tsentralny Tuyuksu Glacier. This similarity suggests that the projections based on the lapse rates of Brugger and Goldstein may be reasonable for the Mammoth Glacier.
Naftz et al. (1996) indicates a cooler period within the ice cores from Fremont Glacier from the bottom of the core (1725 A.D.) until 1810 A.D. coincident with a rapid 2-3 year warming shift at LIA termination. However, the shift in δ¹⁸O provides no associated change in temperature and precipitation values. Projecting the climate data to LIA and late-glacial ELAs, using the Colorado lapse rates, indicates that cooling of < 1 °C and precipitation increase of ~ 4 cm SWE might be enough to re-grow glaciers to LIA extents and beyond. There is minimal variation between summer temperature and winter precipitation based on Brugger and Goldstein’s lapse rates from modern to late-glacial as there is only 85 m of ELA depression (0.75 °C and 2 cm SWE).

The minor temperature and precipitation variation between modern, LIA and late-glacial indicate the Mammoth Glacier’s, and likely the Wind River Range’s, sensitivity. These analyses strengthen Leonard’s 2007 analysis that required only 2 °C temperature depression to translate into 5-10% re-growth to LGM glacial configuration. As the Mammoth Glacier is actively retreating it is likely the actual ELA is above that mapped altitude based on the 2002 Air photos and is therefore colder and wetter placing it closer to Leonard’s climatic boundaries. However, the sensitive nature of the Green River Lakes drainage might contradict the assertion that a glacier has existed from 4500 yr B.P till present; rather the LIA advance is a rapid re-growth from non-glacial conditions.

**5.5 Summary**

The Wind River Range, with a wealth of glacial deposits, was identified early as a crucial link in the Rocky Mountain glacial history; research dates back to Blackwelder (1915). Refinements to the glacial chronology have established the Pleistocene advances whereas my work serves to document the late-Holocene fluctuations.
Three major glacial events are recorded in the geomorphic deposits in the Green River lakes study area: late-Pleistocene recessional, late-glacial (Younger Dryas?) and Little Ice Age. This work reinterprets the three-fold Holocene advance sequence for the drainage proposed by (Mahaney, 1993). Deposits he related to Indian Basin and Audubon advances (both early Neoglacial) instead appear to be related to either late-glacial or LIA events. The glaciers reached Neoglacial maximum during the late LIA, as recorded in both the fresh character of the LIA terminal moraine and in the substantial peaks in MS and clastic sedimentation to UGL. My work identifies only the single Holocene moraine correlative to the LIA (Gannett Peak) advance. This absence of pre-LIA Holocene advances matches results from other alpine regions around the west (e.g., Mijal, 2008). An alpine cirque glacier below the Mammoth Glacier headwall has existed from at least 4500 yr B.P. until present based on the presence of rock flour sedimentation into the Upper Green River Lake.

Scott Lake was dammed by a late-glacial (Younger Dryas?) moraine and a high alluvial dam may have been subsequently breached by an outburst flood from Mammoth Glacier shortly after the Little Ice Age maximum. The outburst flood from Scott Lake occurred at ~240 yr B.P. as recorded by a sharp pulse of fine clastic rock flour in the sediments of Upper Green River Lake. The lake lowering is recorded by a stranded delta in Scott Lake, by coarse cobble-boulder fan deposits covering the upper delta, and by oxidation staining around the lake indicating relatively recent exposure of this relict shoreline.

The summer temperature and winter precipitation, based on SNOTEL data, indicate only small perturbations (\(\sim 1^\circ C\) and 4 cm SWE\(<10\%\) increase)) would enable LIA conditions to return. The glacial system in the Wind River Range appears to be extremely sensitive to climate variables. As
such future climate change, especially warming, will likely result in human-timescale changes to the remaining glacier. This will further stress the Wyoming water supply removing one of the buffers to the hydrologic system.

5.6 Future Work

The Glew core record in the Lower Green River Lake would be complimented with other Glew, Livingston and Reasoner cores from other parts of the lake. This might assess the deeper portion of the sedimentary record and potentially resolve the rise in MS at the base of LGL-1. To complement the Glew coring in the Upper Green River Lake, Reasoner coring during the winter to recover a longer, continuous record from the main basin would be ideal. As Scott Lake is thought to have existed throughout the Holocene a Glew core from the lake would serve to assess this assertion. Additionally it might capture a varved record, which at its youngest portion might correlate to annual snow data from the Upper Fremont Glacier.

To better understand the effects of hypsometry that the AAR method negates, the application of the balance area ratio (BAR) (Benn and Gemmel, 1997) would serve to better constrain ELAs in the study area. In turn this would better constrain the amount of climate change required to regrow glaciers to LIA and late-Glacial extents.

The moraine at the mouth of Tourist Creek deserves future study as it appears to be an outlier and possibly records a separate glacial advance in the late-Pleistocene, not noted elsewhere in the Wind River Range. It might be a good candidate for CRN study. Finally having an accurate gauge of temperature and precipitation at the modern Mammoth Glacier would be vital in assessing the sensitivity of the glacier. Thus far interpolated SNOTEL, PRISM models and
imported lapse rates have poorly constrained the actual temperature and precipitation. This would be best accomplished by installing a weather station at Mammoth Glacier, preferably close to the cirque headwall.
6.0 References


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PRISM Climate Group, Oregon State University, http://prism.oregonstate.edu, assessed Feb 2011


Wright, H.E., Jr., 1967, A square-rod piston sampler for lake sediments. *Journal of Sedimentary Petrology* 37: 975-976


7.0 Figures

Figure 1. Digital hillshade map of the Wind River Range in west central Wyoming. The Green River Lakes study area is outlined in the northwest portion of the range. [Data from WYGISC]
Figure 2. A hillshade DEM of the major valleys, glaciers and lakes in the Green River Lakes drainage. Lakes are shown in blue, glaciers in pink. [Data from WYGISC]
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Figure 6. Modeled extents of glaciation in individual ranges in the southern and central Rocky Mountains that would develop with progressive summer temperature depression in absence of precipitation change. Extent of glaciation is expressed as a proportion of mapped LGM glacier extent in each range. Note greater sensitivity of Wind River Range to small temperature changes compared to other mountains in the region. Modified from Leonard, 2007 (figure 6).
Figure 7. The chronology for the late-Pleistocene through Holocene deposits and climatic conditions near Mammoth Glacier as proposed by Mahaney, 1993.
Figure 8. Site of the Upper Fremont Glacier ice cores from 1991 (Naftz, 1993b) and 1998 (Schuster et al., 2000). From Schuster et al., 2000.
Figure 9. Winter precipitation and mean summer (June-August) temperature at Equilibrium Line Altitudes (ELA) for 32 world-wide glaciers (modified from Leonard 1989 (figure 1)).
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Figure 11. Upper Green River Lake bathymetry (5 ft contours) with the location of the bathymetry sounding points (black x’s) from summer 2008. (Axes are in UTM coordinates (m) UTM zone 12N).
Figure 12. Lower Green River Lake bathymetry (10 ft contours) with the location of the bathymetry sounding points (black dots) from summer 2008. (Axes are in UTM coordinates (m) UTM zone 12N)
Figure 13. Lake coring site in the Lower Green River Lake (LGL) during the 2008 summer field season. The core site is plotted on a false color digital orthophoto quarter quad (DOQQ) from the WYGIS. See Table 1 for complete details of water depths and type of core device used.
Figure 14. Lake coring sites in the Upper Green River Lakes (UGL) from this study. The core sites are overlaid on false color digital orthophoto quarter quad (DOQQ) from the WYGIS. See Table 1 for complete details of water depths and type of core device used.
Figure 15. Glew core LGL-1 showing visual stratigraphy (left), and magnetic susceptibility (MS) and organic content (OC) with sediment depth.
Figure 16. Glew core LGL-1 showing visual stratigraphy (left), and magnetic susceptibility (MS) and grain size distribution (GSD) with sediment depth.
Figure 17. The Glew core image at left details the visual stratigraphy (VS) of LGL-1 with sediment depth (cm). Two reported radiocarbon analyses are noted in core image; central intercepts of the 2-sigma calibrated age ranges of the analyses shown in the age model. The depth-age model at right uses a linear extrapolation from the top of the core to the samples, then a similar extrapolation to the core bottom below the radiocarbon control.
Figure 18. The plot of magnetic susceptibility (MS) and organic content (OC) for LGL-1 with calibrated age using the transformation detailed in the depth-age model.
Figure 19. The plot details magnetic susceptibility (MS) and grain size distribution (GSD) LGL-1 with calibrated age using the transformation detailed in the depth-age model.
Figure 20. Glew core UGL-5 showing visual stratigraphy (left), and magnetic susceptibility (MS) and organic content (OC) with sediment depth.
Figure 21. Glew core UGL-5 showing visual stratigraphy (left), and magnetic susceptibility (MS) and grain size distribution (GSD) with sediment depth.
Figure 22. Visual stratigraphy, $^{210}\text{Pb}$ analyses, and age-depth model of core UGL-5 that was processed for $^{210}\text{Pb}$ dating, respectively.
Figure 23. Depth-age model for UGL-5 using three selected $^{210}$Pb points and calibrated radiocarbon date correlated from UGL-LIV5 based on MS stratigraphy. The model assumes linear interpolation between control points.
Figure 24. The plot of magnetic susceptibility (MS) and organic content (OC) for UGL-5 with calibrated age using the transformation detailed in the depth-age model.
Figure 25. The plot details magnetic susceptibility (MS) and grain size distribution (GSD) for UGL-5 with calibrated age using the transformation detailed in the depth-age model.
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Figure 27. Livingston core UGL-LIV5 showing visual stratigraphy (left), and magnetic susceptibility (MS) and organic content (OC) with sediment depth.
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Figure 32. Map of the Wells Creek Drainage showing the extent of both Scott Lake paleo-shoreline and the Little Ice Age glacier extents. Labels indicate locations of small ice-marginal lakes formed by Mammoth Glacier along its edges during LIA maximum. Sudden release of these ponds may have initiated outburst flood(s) that resulted in breaching of the sediment dam at the outlet of Scott Lake.
Figure 33. Photo looking west (top) along the northern margin of Scott lake. Paleo-shorelines are in the stained oxidized bedrock oxidation staining. The enlarged image (bottom) details the staining on the bedrock.
Figure 34. View east toward Scott Lake and Gannett Peak with the outline of the paleo shoreline (yellow) and the remains of the Scott Lake moraine/till body (blue). Scott Lake is ~625 m long (E-W) and ~ 200 m wide (N-S).
Figure 35. The Wells Creek drainage with base layer DOQQ’s showing the three major glaciers (Mammoth, Minor and Baby) mapped at modern and LIA extents. ELA shown were derived using the AAR method (See Appendix D for details about ELA).
Figure 36. Temperature data averaged for June, July and August from 1985-2010 with linear interpreted lapse rate for temperature. Data compiled from SNOTEL sites (USDA –NRCS) from western central Wyoming (see Figure 5 and Table 10 for details).
Figure 37. Snow accumulation data at point of maxim yearly accumulation (March) from 1985-2010 with linear interpreted lapse rate for precipitation. Data compiled from SNOTEL sites (USDA –NRCS) from western central Wyoming (see Figure 5 and Table 10 for details).
Figure 38. Temperature data averaged for June, July and August from 1999-2010 with linear interpreted lapse rate for temperature. Data compiled from SNOTEL sites (USDA –NRCS) from western central Wyoming (see Figure 5 and Table 10 for details).
Figure 39. Snow accumulation data at point of maximum yearly accumulation (March) from 1999-2010 with linear interpreted lapse rate for precipitation. Data compiled from SNOTEL sites (USDA – NRCS) from western central Wyoming (see Figure 5 and Table 10 for details).
Figure 40. Winter precipitation and mean summer (June-July - August) temperature at Equilibrium Line Altitudes (ELA) (Equations 5 + 6 and Leonard, 1989). The black markers represent the modern climate data at Mammoth Glacier ELA (3610 m AAR) using extrapolated SNOTEL data (squares) (9.98 °C and 33.98 cm SWE), circles PRISM data (1980-2010 - 7.16 °C and 63.09 cm SWE, 1971-2000 – 4.8 °C and 75.74 cm SWE, 1895-2010 - 6.27 °C and 63.14 cm SWE) and Colorado lapse rates applied to SNOTEL data (3.78 °C and 68.48 cm SWE) from Brugger and Goldstein (1999) (Table 11 and Table 12). The red triangle is LIA ELA – 3575 m (AAR) (4.04 °C and 63.64 cm SWE) and the blue diamond is Late-Glacial (YD?) ELA – 3525 m (MA) (4.5 °C and 61.74 cm SWE) using the Brugger and Goldstein (1999) Colorado lapse rates.
### Table 1

The results and timing of the summer 2008 coring of the Upper and Lower Green River Lake. This table links with Figure 11, Figure 12, Figure 13, and Figure 14, showing the exact locations of each of these cores sites.

<table>
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<tr>
<th>Sediment Core Site</th>
<th>Type of Corer</th>
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<th>Date</th>
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<td>7/11/2008</td>
</tr>
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<td>UGL-2</td>
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<td>7/11/2008</td>
</tr>
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<tr>
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<td>7/17/2008</td>
</tr>
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<td>UGL-6</td>
<td>Glew</td>
<td>1.2 m</td>
<td>7/18/2008</td>
</tr>
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<td>UGL-7</td>
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<td>27.1 m</td>
<td>9/7/2008</td>
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<td>UGL-8</td>
<td>Glew</td>
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</tr>
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Table 2. Macrofossil collections LGL-1 and UGL-5 recovered from the Green River Lakes in Summer 2008.

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<th>Depth</th>
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<th>Suitability</th>
<th>Official Name</th>
<th>Submission Name</th>
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<td>9 cm</td>
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<td>13 cm</td>
<td>pine needle</td>
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<td>no</td>
<td>ND_LGL1-13</td>
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<tr>
<td>22 cm</td>
<td>Twig</td>
<td>1.5 cm</td>
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<td>ND_GRL_13</td>
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<td>ND_GRL_3</td>
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Table 3. Macrofossil collections from UGL-LIV5 recovered from the Green River Lakes in Summer 2008. UGL-LIV5 is broken into the 4*100 cm pushes where UGL-LIV5-1 is at the top and UGL-LIV5-4 is at the bottom of the compiled core. (UGL-LIV5-1 + UGL-LIV5-2)

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<td>49.5 cm</td>
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<td>90 cm</td>
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Table 4. Macrofossil collections from UGL-LIV5 recovered from the Green River lakes in Summer 2008. UGL-LIV5 is broken into the 4*100 cm pushes where UGL-LIV5-1 is at the top and UGL-LIV5-4 is at the bottom of the compiled core. (UGL-LIV5-3 + UGL-LIV5-4)

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</tr>
<tr>
<td>ND_UGL_LIV5-3-65</td>
<td></td>
</tr>
<tr>
<td>ND_GRL_10</td>
<td></td>
</tr>
<tr>
<td>98 cm</td>
<td></td>
</tr>
<tr>
<td>wood fragment</td>
<td></td>
</tr>
<tr>
<td>0.5 cm</td>
<td></td>
</tr>
<tr>
<td>no</td>
<td></td>
</tr>
<tr>
<td>ND_UGL_LIV5-3-98</td>
<td></td>
</tr>
<tr>
<td>ND_GRL_11</td>
<td></td>
</tr>
</tbody>
</table>
Table 5. Full report of $^{14}$C dating results from 3 cores (UGL-5, UGL-LIV5 and LGL-1) from the Upper and Lower Green River Lakes run at Center for Accelerated Mass Spectroscopy (CAMS) at Lawrence Livermore National Laboratory under the guidance of Tom Guilderson. The calibrations were run using Calib 6.0. Highlighted calibration ranges represent the highest relative probabilities.

<table>
<thead>
<tr>
<th>Sample Depth</th>
<th>ND_UGL5-36</th>
<th>ND_UGL5-66</th>
<th>ND_UGL-LIV5-1-38</th>
<th>ND_UGL-LIV5-1-52.5</th>
<th>ND_UGL-LIV5-1-87</th>
</tr>
</thead>
<tbody>
<tr>
<td>Corrected Core Depth</td>
<td>36 cm</td>
<td>66 cm</td>
<td>38 cm</td>
<td>52.5 cm</td>
<td>87 cm</td>
</tr>
<tr>
<td>14C age</td>
<td>70</td>
<td>110</td>
<td>365</td>
<td>385</td>
<td>965</td>
</tr>
<tr>
<td>cor. +/-</td>
<td>200</td>
<td>100</td>
<td>35</td>
<td>35</td>
<td>35</td>
</tr>
<tr>
<td>cal_BP(lower)</td>
<td>-5</td>
<td>-3</td>
<td>328</td>
<td>333</td>
<td>799</td>
</tr>
<tr>
<td>(upper)</td>
<td>0</td>
<td>-1</td>
<td>359</td>
<td>350</td>
<td>813</td>
</tr>
<tr>
<td>rel._prob. (lower)</td>
<td>0.018</td>
<td>0.013</td>
<td>0.307</td>
<td>0.201</td>
<td>0.16</td>
</tr>
<tr>
<td>(upper)</td>
<td>153</td>
<td>148</td>
<td>374</td>
<td>502</td>
<td>866</td>
</tr>
<tr>
<td>rel._prob. (lower)</td>
<td>0.569</td>
<td>0.672</td>
<td>0.054</td>
<td>0.799</td>
<td>0.478</td>
</tr>
<tr>
<td>(upper)</td>
<td>169</td>
<td>188</td>
<td>429</td>
<td>901</td>
<td>901</td>
</tr>
<tr>
<td>med._prob.</td>
<td>195</td>
<td>147</td>
<td>423</td>
<td>448</td>
<td>859</td>
</tr>
<tr>
<td>mean_prob.</td>
<td>80</td>
<td>80</td>
<td>460</td>
<td>470</td>
<td>850</td>
</tr>
</tbody>
</table>
Table 6. Full report of $^{14}$C dating results from 3 cores (UGL-5, UGL-LIV5 and LGL-1) from the Upper and Lower Green River Lakes run at Center for Accelerated Mass Spectroscopy (CAMS) at Lawrence Livermore National Laboratory under the guidance of Tom Guilderson. The calibrations were run using Calib 6.0. Highlighted calibration ranges represent the highest relative probabilities.

<table>
<thead>
<tr>
<th>Sample</th>
<th>ND_UGL-LIV5-2-40</th>
<th>ND_UGL-LIV5-3-14</th>
<th>ND_UGL-LIV5-4-26</th>
<th>ND_UGL-LIV5-4-92</th>
<th>ND_LGL1-22</th>
<th>ND_LGL1-24</th>
</tr>
</thead>
<tbody>
<tr>
<td>Corrected Core Depth</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14C age</td>
<td>1330</td>
<td>3290</td>
<td>2110</td>
<td>3800</td>
<td>2940</td>
<td>2935</td>
</tr>
<tr>
<td>cor. +/-</td>
<td>35</td>
<td>70</td>
<td>70</td>
<td>50</td>
<td>35</td>
<td>45</td>
</tr>
<tr>
<td>cal_BP(lower)</td>
<td>1187</td>
<td>3445</td>
<td>1991</td>
<td>4090</td>
<td>3008</td>
<td>3004</td>
</tr>
<tr>
<td>(upper)</td>
<td>1201</td>
<td>3590</td>
<td>2155</td>
<td>4132</td>
<td>3012</td>
<td>3161</td>
</tr>
<tr>
<td>rel._prob.</td>
<td>0.183</td>
<td>0.942</td>
<td>0.901</td>
<td>0.225</td>
<td>0.021</td>
<td>0.984</td>
</tr>
<tr>
<td>(lower)</td>
<td>1258</td>
<td>3600</td>
<td>2269</td>
<td>4137</td>
<td>3036</td>
<td>3194</td>
</tr>
<tr>
<td>(upper)</td>
<td>1297</td>
<td>3612</td>
<td>2295</td>
<td>4249</td>
<td>3049</td>
<td>3197</td>
</tr>
<tr>
<td>rel._prob.</td>
<td>0.817</td>
<td>0.058</td>
<td>0.099</td>
<td>0.74</td>
<td>0.067</td>
<td>0.016</td>
</tr>
<tr>
<td>(lower)</td>
<td>4274</td>
<td>3061</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(upper)</td>
<td>4282</td>
<td>3163</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>rel._prob.</td>
<td>0.035</td>
<td>0.81</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(lower)</td>
<td>3187</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>(upper)</td>
<td>3204</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>rel._prob.</td>
<td>0.102</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>med._prob.</td>
<td>1268</td>
<td>3523</td>
<td>2093</td>
<td>4192</td>
<td>3106</td>
<td>3098</td>
</tr>
<tr>
<td>mean_prob.</td>
<td>1277.5</td>
<td>3517.5</td>
<td>2073</td>
<td>4193</td>
<td>3112</td>
<td>3082.5</td>
</tr>
</tbody>
</table>
Table 7. Area of Scott Lake and Paleo-Scott Lake and simple measurement of the volume of water released during the catastrophic lake lowering event.

<table>
<thead>
<tr>
<th></th>
<th>Area km</th>
<th>Volume (km$^3$)</th>
<th>Volume (m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Paleo Scott Lake</td>
<td>0.2455</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Scott Lake</td>
<td>0.1200</td>
<td>0.0012</td>
<td>2204368</td>
</tr>
<tr>
<td>Difference in Area</td>
<td>0.1255</td>
<td>0.0010</td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td></td>
<td>0.0022</td>
<td></td>
</tr>
</tbody>
</table>
Table 8. Age and ELA for the valley glaciers located within the mapping area. See Figure 5 for a detailed location of each moraine.

<table>
<thead>
<tr>
<th>Moraine</th>
<th>ELA* m</th>
<th>ELA^ m</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lower Green River Lake</td>
<td>3223</td>
<td>3066</td>
<td>Pinedale recession</td>
</tr>
<tr>
<td>Upper Green River Lake</td>
<td>3231</td>
<td>3075</td>
<td>Pinedale recession</td>
</tr>
<tr>
<td>Tourist Creek</td>
<td>3231</td>
<td>3094</td>
<td>Pinedale recession?</td>
</tr>
<tr>
<td>Scott Lake</td>
<td>3604</td>
<td>3523</td>
<td>Younger Dryas?</td>
</tr>
<tr>
<td>Lost Pipe Lake</td>
<td>3597</td>
<td>3539</td>
<td>Younger Dryas?</td>
</tr>
<tr>
<td>Inner Titcomb Lake</td>
<td>3589</td>
<td>3517</td>
<td>11.4 ± 0.5 ka (Gosse, 1995)</td>
</tr>
<tr>
<td>Mammoth LIA</td>
<td>3670</td>
<td>3611</td>
<td>Little Ice Age</td>
</tr>
</tbody>
</table>

*ELA calculated using Mean Altitude (MA)

^ELA calculated using Toe Headwall Altitude Ratio (0.4) (THAR)
Table 9. Measurements of glacial area and ELA’s for three glaciers in the Wells Creek drainage over the past 150 yr.

<table>
<thead>
<tr>
<th>Time</th>
<th>Measurement</th>
<th>Glacier</th>
<th>Mammoth</th>
<th>Minor</th>
<th>Baby</th>
<th>Average</th>
</tr>
</thead>
<tbody>
<tr>
<td>2001*</td>
<td>Area (km²)</td>
<td></td>
<td>1.945</td>
<td>0.483</td>
<td>0.217</td>
<td></td>
</tr>
<tr>
<td></td>
<td>ELA⁺ (m)</td>
<td></td>
<td>3610</td>
<td>3655</td>
<td>3565</td>
<td>3610</td>
</tr>
<tr>
<td></td>
<td>ELA++ (m)</td>
<td></td>
<td>3712</td>
<td>3696</td>
<td>3610</td>
<td>3673</td>
</tr>
<tr>
<td></td>
<td>ELA+++ (m)</td>
<td></td>
<td>3596</td>
<td>3655</td>
<td>3552</td>
<td>3601</td>
</tr>
<tr>
<td>1966^</td>
<td>Area (km²)</td>
<td></td>
<td>2.521</td>
<td>0.643</td>
<td>0.241</td>
<td></td>
</tr>
<tr>
<td></td>
<td>ELA⁺ (m)</td>
<td></td>
<td>3615</td>
<td>3635</td>
<td>3550</td>
<td>3600</td>
</tr>
<tr>
<td></td>
<td>ELA++ (m)</td>
<td></td>
<td>3679</td>
<td>3700</td>
<td>3648</td>
<td>3676</td>
</tr>
<tr>
<td></td>
<td>ELA+++ (m)</td>
<td></td>
<td>3592</td>
<td>3629</td>
<td>3636</td>
<td>3619</td>
</tr>
<tr>
<td>LIA</td>
<td>Area (km²)</td>
<td></td>
<td>2.843</td>
<td>0.865</td>
<td>0.519</td>
<td></td>
</tr>
<tr>
<td></td>
<td>ELA⁺ (m)</td>
<td></td>
<td>3580</td>
<td>3605</td>
<td>3530</td>
<td>3572</td>
</tr>
<tr>
<td></td>
<td>ELA++ (m)</td>
<td></td>
<td>3665</td>
<td>3722</td>
<td>3601</td>
<td>3663</td>
</tr>
<tr>
<td></td>
<td>ELA+++ (m)</td>
<td></td>
<td>3558</td>
<td>3592</td>
<td>3489</td>
<td>3546</td>
</tr>
</tbody>
</table>

* Using DOQQ
^ Using Topographic 1:24k
+ Accumulation Area Ratio
++ Mean Altitude Method
+++ Toe Headwall Altitude Ratio
Table 10. Temperature (June-July-August (JJA)) and Precipitation (Winter - Snow Water Equivalent (SWE)) data for from the region SNOTEL sites proximal to the Green River Lakes Drainage. The data are presented for 2 periods; a 25 year average and a 10 year average. Please refer to Figure 5 for exact locations.

<table>
<thead>
<tr>
<th>SNOTEL Site</th>
<th>Lat</th>
<th>Long</th>
<th>Elevation (m)</th>
<th>JJA Temperature (°C)</th>
<th>Winter SWE (cm)</th>
<th>JJA Temperature (°C)</th>
<th>Winter SWE (cm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gunsight Pass^</td>
<td>43.38</td>
<td>-109.88</td>
<td>2993</td>
<td>NA</td>
<td>NA</td>
<td>10.97</td>
<td>37.15</td>
</tr>
<tr>
<td>Cold Springs</td>
<td>43.28</td>
<td>-109.45</td>
<td>2935</td>
<td>10.82</td>
<td>21.79</td>
<td>11.66</td>
<td>20.36</td>
</tr>
<tr>
<td>Little Warm</td>
<td>43.50</td>
<td>-109.75</td>
<td>2856</td>
<td>10.25</td>
<td>30.58</td>
<td>11.08</td>
<td>27.13</td>
</tr>
<tr>
<td>Kendall</td>
<td>43.25</td>
<td>-110.02</td>
<td>2359</td>
<td>12.04</td>
<td>33.42</td>
<td>12.44</td>
<td>30.35</td>
</tr>
<tr>
<td>Elkhart Park</td>
<td>43.00</td>
<td>-109.77</td>
<td>2365</td>
<td>12.75</td>
<td>34.35</td>
<td>13.22</td>
<td>32.92</td>
</tr>
<tr>
<td>New Fork Lake</td>
<td>43.12</td>
<td>-109.95</td>
<td>2542</td>
<td>12.21</td>
<td>27.69</td>
<td>12.75</td>
<td>26.5</td>
</tr>
<tr>
<td>Gros Ventre Summit</td>
<td>43.38</td>
<td>-110.13</td>
<td>2667</td>
<td>10.28</td>
<td>35.95</td>
<td>10.52</td>
<td>32.17</td>
</tr>
<tr>
<td>Mammoth Glacier ELA*</td>
<td>43.17</td>
<td>-109.69</td>
<td>3610</td>
<td>9.98</td>
<td>33.98</td>
<td>10.52</td>
<td>38.64</td>
</tr>
</tbody>
</table>

*Mammoth Glacier ELA C and SWE

^Gunsight Pass is the closest SNOTEL site to the Green River Lakes drainage
Table 11. Summer temperature (June –July – August) extrapolated from 25 yr average SNOTEL data to Modern, Little Ice Age and Late-Glacial (Younger Dryas?) using temperature lapse rates from Brugger and Goldstein (1999) for two regions of Colorado.

<table>
<thead>
<tr>
<th>Elevation</th>
<th>JJA Temperature (°C)</th>
<th>EC CO * JJA T (°C)</th>
<th>WC CO * JJA T (°C)</th>
<th>MG ‐ Modern ELA @</th>
<th>MG ‐ LIA ELA #</th>
<th>MG ‐ LG(YD?) ELA $</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cold Springs</td>
<td>2935</td>
<td>10.82</td>
<td>5.96</td>
<td>5.15</td>
<td>5.42</td>
<td>5.87</td>
</tr>
<tr>
<td>Little Warm</td>
<td>2856</td>
<td>10.25</td>
<td>4.82</td>
<td>3.92</td>
<td>4.18</td>
<td>4.63</td>
</tr>
<tr>
<td>Kendall</td>
<td>2359</td>
<td>12.04</td>
<td>3.03</td>
<td>1.53</td>
<td>1.78</td>
<td>2.25</td>
</tr>
<tr>
<td>Elkhart Park</td>
<td>2865</td>
<td>12.75</td>
<td>7.39</td>
<td>6.49</td>
<td>6.76</td>
<td>7.21</td>
</tr>
<tr>
<td>New Fork Lake</td>
<td>2542</td>
<td>12.21</td>
<td>4.52</td>
<td>3.24</td>
<td>3.49</td>
<td>3.95</td>
</tr>
<tr>
<td>Gros Ventre Summit</td>
<td>2667</td>
<td>10.28</td>
<td>3.49</td>
<td>2.36</td>
<td>2.62</td>
<td>3.07</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td></td>
<td><strong>4.87</strong></td>
<td><strong>3.78</strong></td>
<td><strong>4.04</strong></td>
<td><strong>4.50</strong></td>
<td></td>
</tr>
</tbody>
</table>

Data at LIA and LG are only extrapolated using the western central Colorado lapse rate.

@ - Mammoth Glacier modern ELA = 3610 m (AAR)
# - Mammoth Glacier LIA ELA = 3575 m (AAR)
$ - Mammoth Glacier Late-Glacial (YD?) ELA = 3525 (MA)

*-Temperature Lapse rate from eastern central Colorado = 7.21 °C/km (Brugger and Goldstein, 1999)
^ - Temperature Lapse rate from western central Colorado = 8.44 °C/km (Brugger and Goldstein, 1999)

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Table 12. Precipitation (cm SWE) extrapolated from 25 yr average SNOTEL data to Modern, Little Ice Age and Late‐Glacial (Younger Dryas?) using precipitation lapse rates from Brugger and Goldstein (1999) for two mountain ranges in Colorado.

<table>
<thead>
<tr>
<th></th>
<th>Sawatch Range*</th>
<th>Elk Range^</th>
<th>Sawatch Range</th>
<th>Sawatch Range</th>
</tr>
</thead>
<tbody>
<tr>
<td>Elevation</td>
<td>Winter SWE (cm)</td>
<td>SWE (cm)</td>
<td>SWE (cm)</td>
<td>SWE (cm)</td>
</tr>
<tr>
<td>Little Warm</td>
<td>2856</td>
<td>30.58</td>
<td>59.16</td>
<td>44.08</td>
</tr>
<tr>
<td>Kendall</td>
<td>2359</td>
<td>33.42</td>
<td>80.83</td>
<td>55.81</td>
</tr>
<tr>
<td>Elkhart Park</td>
<td>2865</td>
<td>34.35</td>
<td>62.58</td>
<td>47.68</td>
</tr>
<tr>
<td>New Fork Lake</td>
<td>2542</td>
<td>27.69</td>
<td>68.17</td>
<td>46.81</td>
</tr>
<tr>
<td>Gros Ventre Summit</td>
<td>2667</td>
<td>35.95</td>
<td>71.69</td>
<td>52.83</td>
</tr>
<tr>
<td><strong>Average</strong></td>
<td></td>
<td></td>
<td><strong>68.48</strong></td>
<td><strong>49.44</strong></td>
</tr>
</tbody>
</table>

Data at LIA and LG are only extrapolated using the Sawatch Range lapse rates

@- Mammoth Glacier modern ELA = 3610 m (AAR)

#-Mammoth Glacier LIA ELA = 3575 m (AAR)

$-Mammoth Glacier Late-Glacial (YD?) ELA = 3525 (MA)

*Precipitation Lapse rate from Sawatch Range Colorado = 37.9 cm SWE(Brugger and Goldstein, 1999)

^ Precipitation Lapse rate from Elk Range Colorado = 17.9 cm SWE (Brugger and Goldstein, 1999)
Appendix

Appendix A – UGL Glew Coring

Plot of Magnetic Susceptibility ($10^{-6}$ SI) for seven Glew cores recovered from the Upper Green River Lake.
Plot of Magnetic Susceptibility ($10^{-6}$ SI) for UGL-6, a Glew core recovered from the Upper Green River Lake.
Appendix B – Surficial Geologic Units

Preliminary Surficial Geologic Map of the Green River Lakes Drainage

Landform map of selected sections of the Green River Lakes, Downs Mountain, Squaretop Mountain, and Gannett Peak Quadrangles

Polygon Features (no descriptions):

Lakes

Streams

Glaciers

Polygon Features:

Bedrock (Bb)

Description: Exposed bedrock. In the valleys and high country these exposures are typically sculpted by glacial activity (below Pleistocene glacial trimlines) or mass wasting (above trimlines), and in places form benches or plateaus.

Location: Along valley walls, valley bottoms, along ridges and occasionally in or on the lip of cirques. They are also common where two glaciated valleys meet usually on valley spurs below glacial trimlines.

Associated Landforms/ Features: Valley walls, Glaciers

Process: Glacial scouring, mechanical weathering and tectonic uplift.

Surficial Material: Bedrock.

Mapping Guidelines: Bedrock is the dominant surface unit in this heavily glaciated basin. Unit locally includes areas of thin or discontinuous covers of soil, debris, rubble till.

Rock Glacier (Rg)

Definition: A tongue-like or lobate body of angular boulders that resembles a small debris-covered glacier. Distal terminal slopes are commonly steep (≥ 32°), at or above the angle of repose. Surface generally marked by transverse ridge-and-furrow relief.
Location: Typically found at elevations greater than 9,000 ft.

Associated Landforms: Bedrock, colluvial apron, debris cone.

Process: Glacial

Age: Undifferentiated

Surficial Material: Rock debris (boulders and cobbles) and till

Mapping Guidelines: deposits typically extend down-valley from cirques and high valley walls. Distal extent constrained by prominent steep outer slopes, often greater than angle of repose. Up valley limits inferred from continuity of surface rubble and ridge-and-furrow morphology.

**Debris Avalanche (Da)**
Description: Debris chutes that funnel material directly from bedrock exposures to the valley floor typically bypassing debris cones or are superimposed upon debris cones.

Location: Typically found below bedrock and valley wall landforms.

Associated Landforms/ Features: debris cones, debris cone apron, lakes, streams, valley floor, and valley wall landforms.

Process: When large quantities of rock fall and debris activate they funnel down these avalanche chutes. The chutes typically have steep lateral walls that constrain avalanches and debris movement.

Surfical Material: Alluvium/colluvium

Mapping Guidelines: Narrow chutes that act as the active debris path on debris cones. Source and toe tend to be wider and lack well defined raised lips along flanks.

**Colluvium (Ca)**
Definition: An accumulation of rock debris and regolith derived from single or multiple rock fall or rock topple events. Typically smoother and older than debris cones or debris cone aprons.

Location: Found above valley floor units, generally adjacent to valley walls.

Associated Landforms/ Features: Debris cones, debris cone apron, lakes, meadows, valley floor, and valley wall.
Process: Colluvium is formed from detachment of rock from bedrock cliffs, valley walls, or rock towers generated by mass wasting processes. Large rock fall deposits (talus) generally accumulate over long periods of time. Surfaces can be infilled with vegetation and soil and smoothed by weathering and diffusive processes over time.

Surficial Material: Colluvium

Mapping Guidelines: Typically mapped based on vegetative coverage and in-situ relations to debris cone aprons. Upper bounds mark the transition from active processes to older well established deposits.

**Debris Cone (Dc)**
Description: Debris deposited in a conical shape with a surface slope greater than 10 degrees (perpendicular to contours), usually transported by small streams, snow avalanches, or other mass wasting events.

Location: Typically found below bedrock and valley wall landforms.

Associated Landforms/ Features: Colluvial aprons, debris cone apron, lakes, streams, valley floor, and valley wall landforms.

Process: Debris transported by small streams and snow avalanche events that originate at valley walls and are deposited at the break in valley wall slope. Debris is an accumulation of many episodes where the stream channel or avalanche path changes course back and forth creating a conical fan shape.

Surficial Material: Colluvium

Mapping Guidelines: Debris cone commonly are identified by large swaths devoid of large vegetation (trees) that exhibit a light green color when viewed in aerial photos. Large boulders are also noted to comprise debris cone surfaces in the area.

**Debris Cone Apron (Dca)**
Description: A composite of debris cones that are older than active debris cones, that are unable to be separated into individual events. Debris cone aprons generally are linear features that only occur below areas of high relief.

Location: Typically found below colluvial aprons, ridgelines, and valley wall landforms.

Associated Landforms/ Features: Colluvial aprons, debris cone apron, lakes, streams, valley floor, and valley wall landforms.
Process: Debris transported in by small, past streams and snow avalanche events that originated at valley walls and were deposited at the break in valley wall slope. Debris consists of accumulation of many episodes where the stream channel or avalanche path changes course back and forth creating a conical fan shape. After an extended period of time, debris cones coalesce into indistinguishable events.

Surficial Material: Colluvium

Mapping Guidelines: When viewed in aerial photos, debris cone aprons are identified by large, light green shrub vegetation and are devoid of large trees or expansive forests, both indicative of recent debris landsliding and avalanching.

Outwash (O)
Description: Glacially derived fluvial deposit ranging in grain size from silt and clays to boulders.

Location: Found in high valleys (>10,000 ft) where low gradients and basins trap sediment in meltwater from glaciers upstream.

Mapping Guidelines: Similar to the deposits in Valley floor but are typically restricted to higher elevations near active glaciers. Found exclusively in above the inlet to Scott Lake.

Meadow (M)
Description: Broad, low-gradient mountain wetland areas, dominated by small grasses, flowering plants, and small shrubs.

Location: Adjacent to alpine lakes, valley floors, and within cirque basins

Mapping Guidelines: Meadow areas are mapped primarily by air photo interpretation. In air photos, meadows areas are indicated by unforested grey or green areas near water sources.

Valley Wall (Vw)
Description: Steep forested slope ranging from 20° to more than 60° that is controlled by local bedrock. A thin venire of till and/or colluvium is common. Above trimlines, valley wall consists primarily of un-vegetated bedrock and colluvium.

Location: Mountain slopes below ridges and above debris cone, debris cone apron, or colluvial aprons.
Associated Landforms/Features: Ridgelines, bedrock bench, meadow,

Process: Alpine glaciation.

Surfical Material: Bedrock, till, colluvium.

Mapping Guidelines: The lower boundary of the valley wall landform is hard to distinguish between valley floor deposits, thus dashed lines are used to approximate the contact.

**Valley Floor (Vf)**
Description: Main valley unit which is controlled by bedrock and is gently sloping (less than 20°). Surface materials consist of thin venires of till, colluvium, and alluvium. At the lowest reaches of the valley floor landform, bedrock confined streams are often found. In upper cirques, valley floor landforms are also found.

Location: Valley floor materials are located in the lowest reaches of the valley and are bounded by colluvial aprons, debris cones, and debris cone aprons.

Associated Landforms: Bedrock bench, colluvial apron, debris cone, debris cone apron, lakes, streams, and valley wall.

Process: Glacial and fluvial

Surfical Material: Till, colluvium, and alluvium

Mapping Guidelines: In aerial photographs, the valley floor landform can be identified by large expanses of heavily forested valley floors, which often have surface slopes that are less than 20°. Upper extents of the landform are hard to identify, thus it is common to use dashed lines to approximate contacts.
Appendix C – Surficial Geologic Map

Preliminary Surficial Geologic Map Green River Lakes Drainage

Nigel Davies and Douglas H. Clark  Western Washington University  EMDAP Submission  9/22/2010

Map Projection: NAD 1983 UTM Zone 12N

Lakes
Glaciers
Meadow
LIA Till
Outwash
Debris Avalanche
Rock Glacier
Colluvium
Debris Cone
Debris Cone Apron
Valley Floor
Valley Wall
Bedrock
Appendix D – Wells Creek Glaciers

I performed a detailed Equilibrium Line Altitudes (ELA) analysis in the Wells Creek drainage for Mammoth, Minor and Baby glacier. While the data is presented in the results, a sensitivity analysis based on methodology was also performed. The results are presented in this appendix.

Over the past 150 years also document the retreat of the Wells Creek drainage glaciers.

Different methodology has resulted in multiple average ELA inflations in the past 150 years: AAR = 38 m, MEG = 10 m and THAR = 35 m (Table 7). Over the past 35 years ELA variability is such none of the methodology is consistent, at times AAR reports depressed ELA and MER and THAR report rising ELA. Extracting Minor and Baby Glaciers from the calculation (leaving just Mammoth Glacier) elevates the AAR ELA from LIA to modern only 30 m. Other methods for Mammoth Glacier, MEG and THAR inflate the LIA to modern 47 m and 58 m (Table 7).

The AAR method, an established norm for ELA measurement, tends to correlate well with the THAR 0.4 method (Appendix C). The THAR method lowers the ELA (with respect to the AAR) consistently by 10 -25 m, a conservative approach. The linear trend similarly increases ELA over the past 150 (eg: Mammoth Glaciers AAR= 0.22 m/year THAR = 0.26 m/year). Conversely, the AAR method compared to MEG method delivers a weak correlation. The ELA’s are inflated by upwards of 100 m compared to AAR methods. The MEG method for Minor Glacier elevates the ELA whereas both THAR and AAR show depression in the ELA. Thus THAR method, which requires far less time to complete than AAR, would be a quick proxy.
The areas of the three named glaciers (Mammoth, Baby and Minor) in the Wells Creek drainage over the past 150 years.
A method comparison for Baby Glacier for AAR, MEG and THAR over the past 150 years
A method comparison for Minor Glacier for AAR, MEG and THAR over the past 150 years
A method comparison for Mammoth Glacier for AAR, MEG and THAR over the past 150 years
Scott Lake bathymetry (5 ft contours) based on depth sounding (black dots) work complete in the summer of 2008.
Upper Green River Lake bathymetry with the location of the core sites from summer 2008.
Bathymetric map of the Lower Green River Lake and surrounding topography compiled by Wyoming Fish and Game Pinedale office in 1964.
Lower Green River Lake bathymetry with the location of the core site (LGL-1) from summer 2008.