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A Holocene Glaciolacustrine Record of the Lyman Glacier and Implications for Glacier Fluctuations in the North Cascades, Washington

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A Holocene glaciolacustrine record of the

Lyman Glacier and implications for glacier

fluctuations in the North Cascades, Washington

by

Harold N. Wershow

Accepted in Partial Completion
of the Requirements for the Degree
Master of Science

Kathleen L. Kitto, Dean of the Graduate School

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Harold Wershow

June 7, 2016
A Holocene glaciolacustrine record of the

Lyman Glacier and implications for glacier

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A Thesis
Presented to
The Faculty of
Western Washington University

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

by
Harold N. Wershow
June 2016
Abstract
The Holocene glacial history of the North Cascades is poorly understood, in part because most existing records rely on moraine remnants and are therefore discontinuous. To develop a more complete record of Holocene fluctuations of North Cascades glaciers, we collected and analyzed glaciolacustrine sediments (i.e., rock flour) deposited over the past ~7800 years in Lyman Lake by the upstream Lyman Glacier. We combined these results with equilibrium-line altitude (ELA) reconstructions and glacier-climate modeling to quantify the climatic conditions that drove these fluctuations. Finally, we compared the Lyman Glacier’s continuous fluctuation record to existing glacier and climate records of the North Cascades.

Our results indicate that the Lyman Glacier was absent in the early Holocene, from before 7.8 ka until ~4.9 ka, when it experienced an early Neoglacial advance that persisted until at least ~3.8 ka. Following an extended non-glacial interval, the glacier experienced significant advances between ~2.6 – 2.25 ka, ~1.8 – 1.3 ka and ~1.1 – 0.9 ka. An advance starting ~0.8 ka (1150 CE) culminated at the glacier’s maximum Holocene extent between ~0.6–0.5 ka (~1350 – 1450 CE), from which it retreated and disappeared entirely by ~0.35 ka (~1600 CE). After ~200 years with no significant glacier presence in the cirque, the glacier reformed and rapidly advanced to its maximum Holocene extent (~1800 – 1900 CE). Following this event, the glacier retreated steadily throughout the 20th and early 21st centuries and as of 2014, has approached its minimum viable extent. Paleo-ELA reconstructions of the glacier’s maximum Holocene extents suggest that summers were ~2.6 °C cooler than modern (1981 – 2010 CE); alternatively, glacier-climate modeling indicates that annual temperatures ~1.5 °C cooler than modern would result in maximum glacier extents.
Combining these new results with existing North Cascades glacial records indicates that: 1) the earliest Neoglacial advances in the region (starting ~6 ka) occurred asynchronously, with higher latitude and more maritime sites experiencing earlier advances; 2) Neoglacial advances remained small, infrequent and asynchronous until the last millennium; 3) Beginning at ~1.0 ka, glaciers throughout the North Cascades advanced synchronously, signaling the onset of the Little Ice Age (LIA); 4) North Cascades glaciers reached their maximum Holocene extents during the 15th and early 16th centuries (~0.55 – 0.45 ka), followed by apparent regional retreat and a final smaller 19th century (~0.15 – 0.05 ka) re-advance. The asynchronous early-to-mid Neoglacial fluctuations followed by synchronous LIA behavior suggests that local climate factors drove glacier fluctuations until the regional climate signal became strong enough to induce synchrony ca. 1.0 ka. Although the inferred regional retreat remains uncertain, the disappearance of the Lyman Glacier in the mid-LIA (~0.45 – 0.15 ka) is consistent with the precipitation record at Castor Lake (~100 km to the east), which indicates unusually dry winter conditions between ~1450 – 1850 CE (~0.5 – 0.1 ka).
Acknowledgements

As science advances by the collaboration of dedicated individuals, so this project came to completion by the outstanding efforts of many, many people. I am deeply grateful to everyone who lent their valuable time and effort to this endeavor. A small army of field assistants served as human mules, hauling hundreds of pounds of coring equipment and ancient mud over Spider Gap in the quest for a glacial history book. Thanks, Tien Devin Abe Katrina Mike Becky Adam Nigel Doug Geoff and Dick. Invaluable data were shared by Scott Buehn of Chelan County Public Utility District, Jim Vallance and Michael Clynne of the United States Geological Survey, Alan Hidy and Tom Guilderson of Lawrence Livermore National Laboratory and Andy Bunn of Western Washington University. Too many people shared their ideas and knowledge with me to properly cite; I’m particularly grateful to Joanne Egan and Mike Zawaski. The scientists at LacCore, including Amy Myrbo, Jessica Rodysill, Anders Noren and Jessica Heck, lent their technical expertise for the duration of the project. After spending months with department technician Ben Paulson battling an unruly grain-size analysis instrument, John Clague of Simon Fraser University graciously made their instrument available, free of charge. I am most thankful for the years of energetic assistance that Nigel Davies of Eastern Washington University (and previously of WWU and Pomona College) has happily bestowed upon me. Mitchell Plummer patiently worked through hour upon hour of troubleshooting code with me as we collaborated on a glacier–climate model. Robert Mitchell and Scott Linneman lent their experience and expertise as members of my committee. Finally, a heartfelt thanks to my advisor, Doug Clark, for presenting me with the world’s most beautiful field site, for his unflagging generosity, and for supporting me in my every endeavor, glacial or otherwise.

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Comprehensive Introduction
This thesis manuscript, “A Holocene glaciolacustrine record of the Lyman Glacier and implications for glacier fluctuations in the North Cascades, Washington” has also been submitted for review and potential publication in Quaternary Science Review. In addition to the primary author (Harold Wershow), Doug Clark and Mitch Plummer are listed as second and third authors, respectively. Wershow and Clark defined the project and selected the field site, Wershow mapped the field area, Wershow and Clark collected sediment cores from the lake and surrounding locales, Wershow processed the samples, and Wershow and Clark analyzed the results. Plummer and Wershow adapted Plummer’s glacier-climate model to the site and assessed the results. Wershow prepared the manuscript, and Clark reviewed and edited it.

This study reports the first continuous Holocene glacier fluctuation record for the North Cascades by synthesizing original data from the Lyman Glacier with existing North Cascades glacial records. To develop a fluctuation record for the Lyman Glacier, we use lake sediment cores from Lyman Lake to constrain rock flour flux, a proxy for glacier extent. Rock flour flux is quantified by a suite of established proxy measurements (magnetic susceptibility, loss-on-ignition, dry bulk density, and grain-size distribution). We calibrate the relationship between Lyman Glacier’s extent and the rock flour flux by comparison to documentation of the glacier’s historic extent. Age controls on the Holocene record are provided by AMS radiocarbon dating, Pb-210 analysis and tephrochronology. The resulting fluctuation record of the Lyman Glacier, when combined with existing North Cascades glacier fluctuation records, reveals unexpected asynchronous early Neoglacial behavior followed by an abrupt
transition to synchronous behavior during the Little Ice Age. Equally unexpected, the resulting North Cascades record suggests a period of regional glacier retreat between ~1500 – 1800 CE that has not previously been recognized.

To assess the amount of climate change necessary to induce the transition to synchronous glacier behavior, we reconstructed the paleoclimate of the Lyman Glacier’s maximum extent using ELAs and glacier-climate modeling. We reconstruct paleo-ELAs using the Area x Altitude Balance Ratio method and calculate the climate change necessary to restore these ELAs. As an alternative method, we updated Plummer’s energy balance – iceflow model with climatologic and glaciologic inputs from the field site. The results directly indicate the amount of cooling from modern climate conditions to those of the glacier’s maximum Holocene extent, and can be used to infer the amount of cooling necessary for the North Cascades to exhibit synchronous glacier advances.
1.0 Introduction
Paleoclimate records offer researchers a means to understand the temporal and spatial nature of natural climate variability. Glaciers leave records such as moraines that reflect their advances and retreats (i.e., fluctuations) in response to changing climate. Alpine glaciers offer insight into temporal climate variability by recording a time-averaged climate signal that dampens stochastic weather variability and permits paleoclimate interpretation on decadal to centennial time scales. Moreover, glacier fluctuation records can be compared within and between regions, providing insight into spatial climate variability (Davis et al., 2009). Whereas some locales in western North America have well-developed Holocene histories, such as the southern Coast Mountains of British Columbia (e.g., Menounos et al., 2009; Mood and Smith, 2015), the North Cascades of Washington lack a robust Holocene history of glacier fluctuations.

Glacier fluctuation histories can be reconstructed using terrestrial records (e.g., moraines) as well as lacustrine records in the form of proglacial lake sediments. The latter method is based on the relationship between glacier extent and clastic sediment delivery (Karlén, 1976); larger glaciers erode more bedrock, which is reflected in downstream deposition of fine-grained glacigenic sediments known as “rock flour”. A succession of researchers have refined the relationship between glacier activity and lacustrine sediments (e.g., Bakke et al., 2005; Dahl et al., 2003; Karlén, 1981; Leonard, 1986; MacGregor et al., 2011; Osborn et al., 2007), with compelling evidence from varved lake sediments in the Canadian Rockies (Leonard, 1997). Because proglacial lakes are effective sediment traps, they offer the potential of capturing a continuous record of even minor fluctuations from upstream glaciers (Dahl et al., 2003; Leonard and Reasoner, 1999). In contrast, moraines are deposited
episodically and are susceptible to being overridden and eroded, thus preserving a discontinuous record of glacier fluctuations. However, because moraines record absolute paleo-extents whereas glaciolacustrine sediments only record relative glacier activity, a synthesis of both records offers a more robust glacier fluctuation history than either record in isolation (Osborn et al., 2007).

A primary goal of many glacier reconstruction studies is to connect past glacier fluctuations to the climatic perturbations that drove them. Equilibrium-line altitudes (ELAs), which mark the elevation of the transition between a glacier’s accumulation and ablation zones, have been used by many researchers to link glacier fluctuations to regional climate change (Benn and Ballantyne, 2005; Leonard, 1989; Matthews and Karlén, 1992; Osmaston, 2005; Sagredo et al., 2014). Because ELAs can be quantitatively linked to climate parameters such as winter precipitation and summer temperature, reconstructions of paleo-ELAs offer insights into the magnitude of past climate change (Leonard, 1989). Although climatic interpretations of paleo-ELA reconstructions are relatively robust for large valley glaciers that are primarily driven by regional climate (Leonard, 1989), interpretations become progressively more complex for small, sheltered cirque glaciers. Site-specific variables such as aspect, concavity, avalanching and snowdrift significantly affect accumulation and ablation dynamics for cirque glaciers (Graf, 1976), progressively decoupling the glacier’s ELA from the regional climate (Benn and Lehmkuhl, 2000).

To address the limitations of paleo-ELAs in relating cirque glacier fluctuations to climate change, energy balance - ice-flow models are emerging as another method to quantify
paleoclimate (e.g., Harrison et al., 2014; Plummer and Phillips, 2003). Rather than using reconstructed ELAs as an intermediary step between glacier extent and paleoclimate, these models directly build a simulated glacier from climatologic and glaciologic inputs. The models are calibrated by recreating present-day glacier extents from modern climate data. If past extents are well documented, then the paleoclimate inputs can be adjusted until the model produces a glacier that reasonably approximates its paleo-extent. The result is more rigorous constraints on the climate parameters that determined the former extent of the glacier.

Our present understanding of Holocene climate variability in the North Cascades is limited by the relatively sparse and discontinuous nature of the few existing records. Major questions include timing of early Neoglacial and Little Ice Age (LIA) activity and the regional synchrony of these fluctuations. In this study, we present the most comprehensive continuous Holocene glacier fluctuation record for the North Cascades, based on a reconstruction of the Lyman Glacier from circa (ca.) 7,800 years ago to 2014. Our first objective was to reconstruct the glacier’s fluctuation history by using multiple rock flour proxies to assess relative glacier extents while constraining absolute paleo-extents via preserved moraines. The Lyman Glacier is well suited to this purpose: 1) it is a small, discrete alpine glacier with a well-preserved Holocene moraine record; 2) rock flour from the glacier is deposited in a bedrock-dammed lake below the outermost late-Holocene moraine, preserving a continuous lake sediment record (Dahl et al., 2003); 3) historical documentation of the glacier began as early as 1890 CE (Pelto, 2009); and 4) meteorological data have been recorded at a nearby Snow Telemetry site (SnoTel) since 1979 (Buehn, 2014). Our second
objective was to reconstruct the Little Ice Age climatic conditions related to the Lyman Glacier’s maximum Holocene extent. Finally, we synthesize our results with existing Holocene glacier records from the North Cascades to establish a continuous glacier fluctuation record for the region.

2.0 Study Area
The Lyman Glacier is located in the central North Cascades, 17 km northeast of Glacier Peak (Figure 1). The North Cascades is home to hundreds of cirque glaciers (Granshaw and Fountain, 2006) that have a history of observations by mountaineers and formal monitoring by scientists throughout the 20th century (e.g., Long, 1955; Pelto and Riedel, 2001). Following O’Neal et al.’s (2015) definition of the North Cascades, we limit our scope to the mountainous region on either side of the Cascade crest, extending south from the Canadian border for ~120 km. Glaciers here have steadily retreated from their Little Ice Age maximum extents, and many have already disappeared or will soon melt entirely (Pelto, 2008). The region experiences heavy precipitation in response to orographic uplift of Pacific Ocean air masses. Consequently, the east side of the North Cascades exhibits a strong rain shadow. Although the region’s topography is dominantly derived from terrane accretion, it also contains two prominent volcanic peaks, Mt. Baker and Glacier Peak, both of which have been active in the Holocene.

2.1 Lyman Lake catchment basin
The two-tiered Lyman Lake catchment basin contains the Lyman Glacier in its upper-basin and Lyman Lake in its lower-basin (Figure 2). The north-northwest oriented catchment is defined by a high (~ 2400 m) ridgeline on the southern and western margins, forming an “L”
that shelters the glacier from southwest exposure. Although no other glaciers are currently present, there are snowfields in the upper-basin that formed glacierets in the past. The upper-basin is mostly bedrock or recently deglaciated terrain covered by lodgement and ablation till. Beyond the outermost late-Holocene terminal moraine, meadows and stands of conifers proliferate. The lower-basin, in contrast, is mostly forested, although meadows with grüss exposures occupy some of the higher elevations.

Lyman Lake, with an area of 0.28 km$^2$, comprises 3.5% of the total catchment area and occupies the catchment’s outlet at an elevation of 1710 m (Figure 2). The bedrock-dammed lake is split into two sub-basins by a submerged ridge. The upper, southern sub-basin (0.16 km$^2$) reaches a depth of ~11 m but is mostly much shallower due to the prograding delta, whereas the lower, northern sub-basin (0.12 km$^2$) is over 13 m deep with steep walls.

Surface runoff in the catchment basin is dominated by meltwater from the Lyman Glacier, which flows through a series of shallow proglacial lakes before cascading into the lower-basin and creating a substantial delta in Lyman Lake (Figure 2). The only other significant fluvial source enters the lake on the western shore, forming a much smaller delta on top of the submerged ridge.

### 2.2 Lyman Glacier

In the past century, the Lyman Glacier has retreated into its NE-oriented cirque (elevation ~1800 - 2400 m) and in 2014 occupied an area of ~0.42 km$^2$, or 5% of the Lyman Lake catchment basin (Figure 3). The glacier separated into two ice masses sometime before 1963, with its upper lobe avalanching mass to the lower lobe (Post, 1963). At its late-
Holocene maximum, geomorphic evidence indicates a glacier extent of \( \sim 2.7 \text{ km}^2 \), or 32\% of the lake’s catchment. Rapid retreat began between 1901 (Figure 4), when the outer ice margin was near the terminal moraine, and 1915, when a photograph records significant retreat (Freeman, 1941). Historical observations indicate that retreat has been continuous to the present (Pelto, 2009). Notably, the glacier continued to retreat between 1950 – 1976, when regional cooling caused many North Cascades glaciers to pause or re-advance (Pelto and Hedlund, 2001). Annual mass-balance measurements of the glacier’s lower lobe from 1984 to 2008 indicate a yearly mean loss in surface elevation of 1.0 m/a (Pelto, 2009). If this rate is maintained, the glacier is predicted to disappear before 2040 (Pelto, 2009). However, the upper cirque is well sheltered and may receive substantial snowdrift from the southwest ridgeline, potentially allowing the upper lobe to persist.

Local climate conditions are recorded by a nearby Snow Telemetry (SnoTel) site at a similar elevation (1823 m) to the glacier, although the aspect is opposite (south-facing). Between 1989 – 2010, the mean ablation season (May 1\textsuperscript{st} – September 30\textsuperscript{th}) temperature was 7.6 °C and cumulative winter (October 1\textsuperscript{st} – April 30\textsuperscript{th}) precipitation averaged 1.66 m (Buehn, 2014). Because the Lyman Glacier is not at equilibrium (Pelto, 2009), nor has it been since mass-balance measurements began in 1984 (Pelto and Riedel, 2001), it is not possible to establish a true “steady-state” equilibrium-line altitude (ELA). Field observations from late summer of 2014 noted a firn line at 2300 m, only 100 m below the ridgeline.

The Lyman Glacier’s cirque occupies a contact between the Tertiary Cloudy Pass pluton and the early Paleozoic Swakane biotite gneiss (Cater and Crowder, 1967). Both units are locally
heavily altered and commonly intruded by sills and dikes. There is no discernible change in erosive style from one unit to the next. Of importance to radiocarbon dating, no previous workers have observed carbonate rocks in the basin.

Of further importance to lake sedimentation, the slopes surrounding the lake lack landslide deposits. However, gruss exposures are abundant northwest of the lake in the upper meadows of the lower-basin. Streams with headwaters in these meadows transport coarse clastic grains to the small delta on the western flank of the lake (Figure 2).

### 3.0 Methods
To reconstruct a record of the Lyman Glacier’s Holocene fluctuations, we mapped the catchment basin, dated moraines, collected and analyzed lacustrine sediment cores, and dated the stratigraphy. Our sediment analysis goal was to quantify the flux of glacier sediments into Lyman Lake by measuring the volume magnetic susceptibility, loss-on-ignition, dry bulk density and grain-size distribution of the sediment cores. To estimate paleoclimate from these records, we used a glacier-climate model (Plummer and Phillips, 2003) in conjunction with paleo-ELA reconstructions.

#### 3.1 Field mapping and $^{10}$Be dating
Geomorphic mapping of glacial and non-glacial landforms in the catchment basin provides context for calibrating the lacustrine sediment record and subsequent paleoclimate modeling. The mapping also helped supplement historic photographs that display multiple lobes feeding the Lyman Glacier in the early 20th century. In the catchment’s lower-basin, we sought to identify other possible (non-glacial) sources of clastic sediments to the lake (e.g., landslides
and talus). We also mapped bedrock contacts in detail to assess their potential relationship to differential glacier erosion.

To constrain the age of the outermost late-Holocene moraine, we collected surface samples from four boulders for beryllium-10 cosmogenic radionuclide (CRN) exposure dating (Gosse and Phillips, 2001). Following accepted CRN sampling methods, we sampled large (>2 m diameter) flat-topped boulders that protruded above their surroundings and were close to the front of the terminal moraine. Samples were processed and analyzed at Lawrence Livermore National Laboratory’s (LLNL) Center for Accelerator Mass Spectrometry (CAMS), and final model ages (including geometric shielding) were calculated using the CRONUS $^{10}\text{Be}-^{26}\text{Al}$ exposure age calculator (version 2.2, G. Balco, 2009; http://hess.ess.washington.edu).

### 3.2 Lake coring

Before coring Lyman Lake, we recorded a bathymetric survey using a hand-held sonar and differential GPS. For coring, we targeted the northeastern end of the lake’s northern sub-basin because: 1) the ridge that bisects the lake prevents transport of coarse sediments past the southern sub-basin (Figure 2); 2) there are no landslides or talus lobes projecting into the northern sub-basin; 3) the outlet is bedrock-dammed, thus stabilizing the lake’s water level; and 4) underwater mass movements originating from the mid-lake ridge would not impact the upslope coring site (Figure 2). In total, we collected five cores from Lyman Lake: three long Livingstone piston cores (Wright, 1967) and two shorter Glew gravity cores (Glew, 1991, 1988). Two paired Livingstone cores of ~4 m length were recovered at 5.5 m lake depth, as well as a ~4 m core from 8.8 m lake depth. We collected Glew cores of 37 and 24 cm length from near the paired Livingstone coring sites in order to preserve the poorly
consolidated near-surface sediments. In order to assess any possible disturbance to the cores during subsequent transport, we measured magnetic susceptibility (MS) in the field using a Bartington MS2-C. The cores were backpacked from the site, measured again for MS with a Bartington MS2-C at Western Washington University (WWU), and shipped to LacCore (University of Minnesota), a lacustrine core analysis facility.

3.3 Laboratory analysis

Following LacCore’s Initial Core Description procedure, cores were split and imaged on a Geotek Geoscan-III and MS was measured every 0.5cm using a Bartington MS2E point sensor with 0.2 cm resolution. Visual stratigraphy was logged, noting color, cohesion and clast size. Smear slide samples were assessed for the presence of tephra, diatoms, organic material, and clastic sediments. At WWU, the cores were sampled for loss-on-ignition (LOI), dry bulk density (DBD) and grain-size distribution (GSD) using standard procedures (e.g., Davies, 2011). Samples were collected at 4 cm intervals and on either side of major visual changes in the stratigraphy. Samples were extracted from the core using a 1 cm³ syringe (e.g., Bakke et al., 2013), desiccated for density measurements, and then burned at 550 °C for 4 hours to measure organic loss (Heiri et al., 2001). GSD samples were pre-treated with hydrogen peroxide and nitric acid to remove organic matter (Triplett and Heck, 2013). Because diatoms were observed in smear slides, biogenic silica was removed with sodium hydroxide. This reaction was neutralized with hydrochloric acid to prevent digestion of siliciclastics. Grain-size analysis was conducted by laser diffraction with a Malvern Mastersizer 2000 at WWU (using an autosampler) as well as at Simon Fraser University (manual sampling).
3.4 Sediment dating and age-depth modeling

We developed a chronology for the sediment cores using Accelerator Mass Spectrometry (AMS) $^{14}$C dating, $^{210}$Pb dating and tephrochronology. Radiocarbon analyses were conducted at CAMS (LLNL). Because only three terrestrial macrofossils (TMFs) were found in the cores, we also analyzed eleven bulk organic samples. Radiocarbon ages were calibrated using CALIB (version 7.0.4) (Stuiver et al., 2005) and the IntCal13.14c curve. Near-surface sediments from a Glew core (top 12 cm) and a Livingstone core (top 8 cm) were measured for excess $^{210}$Pb using WWU’s Canberra GL 282R Ge gamma spectrometer (Appleby and Oldfield, 1978). Four tephra intervals were sampled at WWU and then identified via electron microprobe analysis at Washington State University’s Geochemistry lab. Two more tephras were tentatively identified based on visual characteristics and relative stratigraphy.

To develop an age-depth model, we used Bacon (Bayesian accumulation), an R-sourced statistical method that incorporates multiple age constraints and uncertainties to develop statistically robust envelopes of accumulation rates (Blaauw and Christen, 2011). Because tephras represent essentially instantaneous deposition, we manually removed visible intervals of airfall tephra from the age-depth model (e.g., MacGregor et al., 2011). For model age constraints, we input un-calibrated radiocarbon dates that were then automatically calibrated by Bacon. Lead-210 dates and calibrated tephra dates were input as calendar dates with symmetrical errors. Because Bacon does not (as of 2015) incorporate asymmetric error distributions, we forced asymmetric calibrated tephra date errors to symmetry. Uncalibrated tephra dates are not always available, as the reported dates are often derived from dated
material above and below the tephra layer, with the actual tephra age interpolated based on sedimentation rates (e.g., Foit et al., 2004).

Although tephras from different volcanoes were distinctive, sequential eruptions from the same source were often visually and chemically un-differentiable. Published tephra ages suffer from the same issue. We used principles of relative stratigraphy to assign published ages to tephra layers in our lake cores, but we acknowledge that tephra layers may be missing from the dated site or from our lake cores. In this study, we assume that the relative stratigraphies are in agreement and that the published ages are reasonably accurate.

### 3.5 Paleoclimate reconstruction

We pursued two methods to quantify the magnitude of climate change related to the Lyman Glacier’s retreat from its maximum Holocene extent to its modern extent: 1) We reconstructed the paleo-equilibrium-line altitude (ELA) associated with the outermost late-Holocene moraines using the Area x Altitude Balance Ratio (AABR) methodology, which explicitly accounts for a paleo-glacier’s hypsometry as well as varying net mass-balance with distance from the ELA (Osmaston, 2005); 2) We adjusted a spatially-distributed energy balance and ice-flow model (Plummer and Phillips, 2003; hereafter referred to as the “glacier-climate model”) to match the maximum Holocene limits of the Lyman Glacier. Both methods require inputs from the modern glacier-climate relationship. The Lyman SnoTel provides nearby climate data that were subsequently adjusted to ELA conditions with temperature (T) and precipitation (P) elevation lapse rates (Table 1). These lapse rates were calculated using monthly T/P data from PRISM 30-year climate normals (800 m grid cell).
overlain on a 10 m DEM (PRISM Climate Group, 2015; see Appendix 3 for Lapse Rate Sensitivity Analysis). Because measurements for some glacier-climate model parameters were not available, we used model-derived data sets as well (Table 1).

Table 1: Climate parameters

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Reference</th>
<th>Use</th>
<th>Notes</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Relative Humidity</td>
<td>(Livneh et al., 2013)</td>
<td>Glacier-climate model</td>
<td>~6 km grid cells, 3-hr time-steps,</td>
<td>modeled</td>
</tr>
<tr>
<td>Wind Speed</td>
<td>(Livneh et al., 2013)</td>
<td>Glacier-climate model</td>
<td>~6 km grid cells, 3-hr time-steps,</td>
<td>modeled</td>
</tr>
<tr>
<td>Temperature</td>
<td>Lyman SnoTel (Buehn, 2014)</td>
<td>Glacier-climate model, ELA conditions</td>
<td>Daily average from 1989 – 2010,</td>
<td>measured</td>
</tr>
<tr>
<td>Precipitation</td>
<td>Lyman SnoTel (Buehn, 2014)</td>
<td>ELA conditions</td>
<td>Daily average from 1989 – 2010,</td>
<td>measured</td>
</tr>
<tr>
<td>Temperature Lapse Rate</td>
<td>(PRISM Climate Group, 2015)</td>
<td>Glacier-climate model, ELA conditions</td>
<td>800 m grid cells, monthly climate normals from 1981-2010</td>
<td>modeled</td>
</tr>
<tr>
<td>Precipitation Lapse Rate</td>
<td>(PRISM Climate Group, 2015)</td>
<td>Glacier-climate model, ELA conditions</td>
<td>800 m grid cells, monthly climate normal from 1981-2010</td>
<td>modeled</td>
</tr>
<tr>
<td>Cloudiness</td>
<td>(Hahn, and Warren, 2007)</td>
<td>Glacier-climate model</td>
<td>~555 km grid cells</td>
<td>modeled</td>
</tr>
</tbody>
</table>

3.5.1 PALEO-ELA RECONSTRUCTIONS

To estimate the maximum Holocene ELA via the AABR method, we first calculated the Lyman Glacier’s modern AABR according to the equation (Rea, 2009):

\[
\text{AABR} = \frac{Z_{acc} A_{acc}}{Z_{abl} A_{abl}}
\]

where \(Z\) is the area-weighted mean altitude, \(A\) is area, and acc / abl refer to accumulation and ablation, respectively. This calculation required the modern glacier’s hypsometry and an
estimation of the modern ELA. For the hypsometry, we used a 10 m DEM derived from aerial photos taken in 1984 (U.S. Geological Survey, 1988). Because the glacier is not at equilibrium, it does not have a steady-state ELA. Therefore, we used the 2300 m firn line observed in late August of 2014 as a proxy for the “modern” ELA. Calculation of the paleo-ELA using AABR requires reconstructing the paleo-hypsometry of the glacier at its maximum Holocene extent. To accomplish this, we used our mapped glacier margins, paleo-flow indicators (e.g., fluted till and striations), and a 1901 photograph showing the glacier’s terminus near its maximum LIA extent (Figure 4). We constructed paleo-surface contours perpendicular to mapped flow indicators, with a concave profile near the headwall and a convex profile near the toe. Our calculation of accumulation area for AABR includes steep rock surfaces of the cirque headwall because they shed effectively all snow onto the glacier before significant ablation occurs. We combined the modern AABR and the paleo-hypsometry in a freely available spreadsheet to iteratively calculate the paleo-ELA (Osmaston, 2005). Modern climate conditions at the paleo-ELA, as calculated by the site-specific lapse rate, can thereby be compared to the “modern” ELA conditions as well as conditions at 32 other modern glacier ELAs in order to estimate climate change (Leonard, 1989; Figure 5).

3.5.2 GLACIER-CLIMATE MODELING

In an effort to develop a more rigorous paleoclimate reconstruction, we used a spatially-distributed energy mass balance model linked to a 2D finite difference ice-flow model (Plummer and Phillips, 2003) to determine the conditions necessary to “grow” a glacier equivalent in size to the maximum Holocene extent of the Lyman Glacier. The energy balance model uses site-specific climate parameters (Table 1) and a 30 m DEM to calculate
energy fluxes and resultant Snow Water Equivalent (SWE) at every 30x30 m grid cell. The ice-flow model uses the SWE results and the DEM to calculate ice-flow between each grid cell and to produce estimates of steady-state ice thickness. We calibrated the energy balance model to the modern glacier-climate relationship by varying constants (Table 2) until positive SWE values were limited to the uppermost cirque (i.e., near the “modern” ELA). The resultant ice-flow approximated the extent of the modern glacier, providing support for our calibrated parameters. To further evaluate our calibration, we compared energy flux outputs from the energy balance model to measurements at the USGS Hut above South Cascade Glacier from 2002 - 2007 (Anslow et al., 2008; Bidlake et al., 2010, 2007, 2005, 2004), an analogous site to the Lyman SnoTel (Figure 6). We chose to model paleo-temperature change because warming has driven the Lyman Glacier’s modern retreat (Pelto, 2009) and because temperature tends to be more regionally consistent than precipitation (Huybers and Roe, 2009). See Discussion (5.5) for further details.

Table 2: Glacier-climate model parameters

<table>
<thead>
<tr>
<th>Energy Balance Model Parameters</th>
<th>Ice-flow Model Parameters</th>
</tr>
</thead>
<tbody>
<tr>
<td>Latitude (decimal degrees)</td>
<td>48.1758</td>
</tr>
<tr>
<td>High albedo value for new snow</td>
<td>0.8</td>
</tr>
<tr>
<td>Low albedo value for old snow</td>
<td>0.4</td>
</tr>
<tr>
<td>Snow emissivity</td>
<td>0.99</td>
</tr>
<tr>
<td>Basin emissivity</td>
<td>0.94</td>
</tr>
<tr>
<td>Ground heat flux (W/m²)</td>
<td>0.04</td>
</tr>
<tr>
<td>Wind coefficient for elevation change</td>
<td>0</td>
</tr>
<tr>
<td>Bulk transfer coefficient</td>
<td>0.0025</td>
</tr>
</tbody>
</table>
4.0 Results

4.1 Moraine sequence of the Lyman Glacier

The Lyman Lake catchment basin contains two distinct sets of moraines from Lyman Glacier: a fresh set directly downvalley of the modern glacier and a substantially more weathered set in the lower-basin and outside of the catchment to the east (Figure 2). The younger set is defined by a terminal moraine with steep, unstable slopes lacking in vegetation or soil development. Within the terminal boundary are numerous small but well-defined recessional moraines. No tephras were observed inside the terminal moraine. Conversely, Mazama ash was observed in exposures directly outside the terminal moraine underneath well-developed soil and vegetation. Berrylium-10 exposure analyses of two prominent boulders in the terminal moraine (Balco et al., 2008) indicate a weighted mean age of 1770 CE (Table 3); a third boulder (5934 $^{10}\text{Be}$ years) appears to reflect prior exposure. The two younger boulder ages are consistent with historical observations of the glacier near the terminal moraine in the 1890s (Jumpponen et al., 1998), and confirm emplacement late in the Little Ice Age.

The set of older moraines includes a terminal moraine that bisects Lyman Lake, creating a prominent underwater ridge. This ridge protects the lake’s northern sub-basin from the inlet stream’s bedload, limiting deposition to suspended sediments. On the western flank of the lake and just east of the lake’s catchment, Mazama ash (~7.7 ka, Bacon and Lanphere, 2006) overlies deposits within similarly weathered moraines, suggesting a Lateglacial deposition for the entire set of older moraines. Although a less likely source, the observed 8.2 ka event in British Columbia cannot be precluded (Menounos et al., 2009).
In the upper reaches of the catchment’s upper-basin, small moraines frequently occur beneath well-shaded snowfields (Figure 3). Historical photographs show glacierets above these moraines, indicating probable deposition during the Little Ice Age.

Although the Lyman Glacier’s cirque occupies a bedrock contact, there is no discernible change in erosive style from one unit to the next. Of further importance to lake sedimentation, the slopes surrounding the lake lack landslide deposits. However, grüss exposures are abundant northwest of the lake in the upper meadows of the lower-basin. Streams with headwaters in these meadows transport coarse clastic grains to the small delta on the western flank of the lake (Figure 2).
### Table 3: $^{10}$Be data

<table>
<thead>
<tr>
<th>Sample Name</th>
<th>Lat. (°N)</th>
<th>Long. (°E)</th>
<th>Elev. (m)</th>
<th>Shielding</th>
<th>$^{10}$Be concentration (atoms/g)</th>
<th>Uncertainty (atoms/g)</th>
<th>Exposure age$^a$ (yr)</th>
<th>Uncertainty (yr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>LYMN14-NEO-01</td>
<td>48.184</td>
<td>-120.901</td>
<td>1813</td>
<td>0.981</td>
<td>116908</td>
<td>2312</td>
<td>5819$^b$</td>
<td>519</td>
</tr>
<tr>
<td>LYMN14-NEO-02</td>
<td>48.18361</td>
<td>-120.902</td>
<td>1805</td>
<td>0.985</td>
<td>5604</td>
<td>235</td>
<td>280$^c$</td>
<td>27</td>
</tr>
<tr>
<td>LYMN14-NEO-04</td>
<td>48.18249</td>
<td>-120.904</td>
<td>1802</td>
<td>0.966</td>
<td>4315</td>
<td>262</td>
<td>220$^c$</td>
<td>23</td>
</tr>
</tbody>
</table>

$^a$ Exposure ages calculated with the CRONUS $^{10}$Be-$^{26}$Al exposure age calculator (version 2.2, G. Balco, 2009; [http://hess.ess.washington.edu](http://hess.ess.washington.edu)), using Nishiizumi et al.'s (2007) nominal isotope ratios and standards and the time-dependent scaling scheme for spallation of Lal, (1991) and Stone (2000). Erosion was assumed to be negligible and samples were collected in 2014.

$^b$ Outlier presumably reflects an inheritance age

$^c$ Weighted average: 244 ± 18, or 1770 CE
4.2 Lake sediment stratigraphy

The three piston cores retrieved from Lyman Lake have consistent stratigraphy, although there are minor variations in thickness. By matching prominent stratigraphic layers, we constructed a 3.50 m composite sediment record from below the Mazama climactic (Mc) ash fall to the surface sediment. Overall, the sediments are well-laminated but remarkably heterogeneous; composition varies from dark, low-density gyttja to light-toned coarse tephras to bluish-gray fine-grained siliciclastic intervals. There is no evidence of sublacustrine mass wasting, depositional hiatuses, post-depositional bioturbation or diagenesis.

Tephra intervals total ~80 cm in thickness, with identified deposits from Mt. Mazama, Mt. Saint Helens and Glacier Peak (Table 4). Mazama climactic (Mc) is tan colored, glass-rich, well sorted and ~4 cm thick. This silty layer is notable for having low MS values. Above Mc, numerous pulses of heterogeneous coarse-grained tephras from Glacier Peak total ~58 cm and include Layer D (GP D) and the Dusty Creek interval (GP DC). Mt. Saint Helens’ Yn tephra (MSH Yn) is ~3 cm thick and dominantly coarse sand with a distinctive salt and pepper appearance. It is glass-poor with abundant cummingtonite and hornblende. A second MSH tephra (Py) is poorly sorted, coarse-grained (silt to sand), and less than 1.5 cm thick. Finally, ~ 10 unnamed coarse-grained co-ignimbrite ash fall deposits from Glacier Peak (GP CAD) vary from <1 cm to 2 cm thick over an interval of ~ 25 cm. Because these ashes have similar physical properties to rock flour, we removed them from our analysis to reduce the “noise” in the rock flour signal (MacGregor et al., 2011). Their removal produces a second composite sediment record of 2.73 m.
## Table 4: Tephra chronology

<table>
<thead>
<tr>
<th>Tephra</th>
<th>Depth (cm)</th>
<th>Identification method</th>
<th>(^{14})C age (yr BP)</th>
<th>± yr</th>
<th>Calibrated age range (cal yr BP)</th>
<th>Source</th>
<th>Constraints</th>
</tr>
</thead>
<tbody>
<tr>
<td>GP CAD</td>
<td>81.5</td>
<td>relative stratigraphy</td>
<td>2016(^d)</td>
<td>16</td>
<td>1950 - 1998</td>
<td>J. Vallance, USGS, p.c.</td>
<td>2215 ± 25 yr BP (^{14})C maximum from lahar beneath upper-most flow</td>
</tr>
<tr>
<td>MSH Py</td>
<td>118.5</td>
<td>microprobe</td>
<td>NA</td>
<td></td>
<td>2576 – 2704(^e)</td>
<td>J. Vallance, USGS, p.c.</td>
<td>2377 – 2737 cal yr BP, minimum (N. Foit, WSU, p.c.); 2700 – 2958 cal yr BP, maximum (Mullineaux, 1996);</td>
</tr>
<tr>
<td>MSH Yn</td>
<td>162.5</td>
<td>microprobe</td>
<td>3442(^f)</td>
<td>62</td>
<td>3635 – 3825</td>
<td>M. Clynne, USGS, p.c.</td>
<td>3510 ± 80 yr BP (^{14})C maximum from basal charcoal (Crandell et al., 1981)</td>
</tr>
<tr>
<td>GP DC</td>
<td>236.5</td>
<td>microprobe</td>
<td>NA</td>
<td></td>
<td>5710- 5820 (upper), 5760 – 5880 (lower)</td>
<td>Foit Jr et al., 2004</td>
<td>Interpolated from TMF ([5195 \pm 40, (^{14})C]) found 5cm below</td>
</tr>
<tr>
<td>GP D</td>
<td>239.5</td>
<td>relative stratigraphy</td>
<td>NA</td>
<td></td>
<td>5990 - 6100</td>
<td>Foit Jr et al., 2004</td>
<td>Interpolated from TMF ([5195 \pm 40, (^{14})C]) found 2cm above</td>
</tr>
<tr>
<td>Mc</td>
<td>267.5</td>
<td>microprobe</td>
<td>6845(^g)</td>
<td>50</td>
<td>7593 - 7787</td>
<td>Bacon &amp; Lanphere, 2006</td>
<td></td>
</tr>
</tbody>
</table>

*Note that only bolded dates were used as inputs for age-depth modeling*

\(^a\) Composite sediment depth of Lyman Lake, with airfall tephra removed  

\(^b\) Microprobe identification by N. Foit at WSU; Relative stratigraphy identification by authors;  

\(^c\) Radiocarbon ages calibrated with CALIB (version 7.1) (Stuiver et al., 2005)  

\(^d\) Weighted average of two \(^{14}\)C dates (2010 ±20 and 2025 ± 25 yr BP) from log found within pyroclastic flow  

\(^e\) Age range calculated by J. Vallance (p.c.) from series of 5 dates found above, between and below two distinct P layers on Mt. Rainer. Upper P layer is presumed to be Py, which is the youngest of the MSH P set.  

\(^f\) Age calculated by M. Clynne (p.c.) from weighted average of eight dated samples  

\(^g\) Age calculated by Bacon and Lanphere (2006) from weighted average of published dates
Rock flour proxies (MS, LOI, DBD, GSD) reveal a trend from lower values in the early Holocene to higher values in the last millennium. LOI and DBD closely correspond, and MS generally varies in agreement with them, although notable exceptions occur near tephra intervals and in the top 30 cm of the core (Figure 7). Clay-sized clastics show no clear trend, but instead maintain a 15-20% base level with narrow peaks surpassing 40% (Figure 8). Fine silt (FS) exhibits a similar trend to MS, LOI and DBD; less than 40% of clastic sediments are FS in the early Holocene, but the FS component rises above 50% by mid-Holocene and peaks during the last millennium at 60% synchronously with MS, LOI and DBD. Coarse silt fluctuates throughout the sediment record, exhibiting no trend. However, peaks in coarse silt (>30% of clastic sediment) commonly occur immediately below and above FS peaks. The sand-sized component of clastic sedimentation is usually negligible, but irregular broad (up to 25 cm) peaks above 10% occur. However, as with clay and coarse silt, sand does not covary with other rock flour proxies.

Our analysis suggests three discrete clastic depositional styles in the cores: significant rock flour, negligible rock flour, and tephra-dominated. The first style is characterized by fine-grained bluish-gray intervals that lacked gyttja and tended to have MS > 120 SI*10^{-5}, LOI < 3% and DBD ~1 g/cc. In contrast, rock flour poor intervals are characterized by red to brown organic-rich sediments of varying grain-size with MS < 80 SI*10^{-5}, LOI > 5% and DBD < 0.8 g/cc. Despite removing visible airfall tephra, persistently high MS values above the airfall layers indicate that slopewash processes continued to transport significant tephra into the lake, overwhelming the background sedimentation signal. These tephra-dominated
intervals are well-sorted, clastic-rich and have MS values greater than the sediments below the airfall tephras.

### 4.3 Chronology of lake sediments

Our radiocarbon analysis produced dates for three terrestrial macrofossils (TMFs) and ten bulk organic samples (Table 5). Unfortunately, a consistent mismatch between the bulk ages and those of known tephra layers and TMFs indicates a significant reservoir effect, with the bulk samples up to 2000 years too old (Figure 9). The source of this reservoir effect is unclear, because there are no known sources of “old” carbon in the drainage or nearby.

Tephrochronology was based on the six known ash layers encountered in the sediment cores (Table 4). We used the most recent constraints available for each tephra and applied them to the most likely correlative portion of each tephra in our composite sediment record. Uncertainties in these assignments typically represent less than 100 years of sedimentation (Appendix 4).

The $^{210}$Pb analysis indicates a near-surface sediment accumulation rate of 6.1 ± 0.3 cm for the last 100 years, with an $R^2$ value of 0.9876. Anthropogenic cesium-137 decay counts demonstrate a sedimentation rate of 6.7 cm/100 yrs since 1952 CE, in close agreement with $^{210}$Pb sedimentation rates. Combining the terrestrial macrofossils, tephras and near-surface dates in Bacon (Blauuw and Christen, 2011) produced a probability distribution age-depth model with sedimentation rates ranging from 1.7 cm/100 yrs in the early Holocene to 7.1 cm/100 yrs in the late Holocene (Figure 10). Although the overall trend of sedimentation rates corresponds with increasing glacier activity, peaks in sedimentation from ~6000 – 5100
cal yr BP and from ~2700 – 2600 cal yr BP are associated with increases in tephra slopewash and coarse-grained clastic sediments, respectively.

**Table 5: Radiocarbon data**

<table>
<thead>
<tr>
<th>Radiocarbon sample</th>
<th>Lab code</th>
<th>Depth (cm)</th>
<th>Material</th>
<th>$^{14}$C age (yr BP)</th>
<th>± yr</th>
<th>Median age (cal yr BP)</th>
<th>2σ age range (cal yr BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>HW-01</td>
<td>169080</td>
<td>12</td>
<td>Bulk</td>
<td>1725</td>
<td>30</td>
<td>1635</td>
<td>1562-1705</td>
</tr>
<tr>
<td>HW-02</td>
<td>169089</td>
<td>16.5</td>
<td>Bulk</td>
<td>1695</td>
<td>30</td>
<td>1598</td>
<td>1538-1634</td>
</tr>
<tr>
<td>HW-03</td>
<td>169083</td>
<td>31.5</td>
<td>Bulk</td>
<td>1950</td>
<td>40</td>
<td>1901</td>
<td>1822-1989</td>
</tr>
<tr>
<td>HW-05</td>
<td>169082</td>
<td>49</td>
<td>Bulk</td>
<td>2860</td>
<td>35</td>
<td>2977</td>
<td>2868-3075</td>
</tr>
<tr>
<td>HW-06</td>
<td>169087</td>
<td>57.5</td>
<td>Bulk</td>
<td>2925</td>
<td>35</td>
<td>3072</td>
<td>2962-3170</td>
</tr>
<tr>
<td>HW-08</td>
<td>169086</td>
<td>82.5</td>
<td>Bulk</td>
<td>2640</td>
<td>35</td>
<td>2761</td>
<td>2727-2799</td>
</tr>
<tr>
<td>HW-16</td>
<td>169091</td>
<td>95</td>
<td>Twig</td>
<td>2330</td>
<td>100</td>
<td>2378</td>
<td>3845-4012</td>
</tr>
<tr>
<td>HW-10</td>
<td>169081</td>
<td>96</td>
<td>Bulk</td>
<td>3635</td>
<td>40</td>
<td>3950</td>
<td>4420-4584</td>
</tr>
<tr>
<td>HW-17</td>
<td>169090</td>
<td>109.5</td>
<td>Wood</td>
<td>2535</td>
<td>30</td>
<td>2624</td>
<td>2135-2714</td>
</tr>
<tr>
<td>HW-15</td>
<td>169085</td>
<td>135</td>
<td>Bulk</td>
<td>4035</td>
<td>35</td>
<td>4498</td>
<td>2494-2597</td>
</tr>
<tr>
<td>HW-19</td>
<td>169084</td>
<td>186.5</td>
<td>Bulk</td>
<td>5055</td>
<td>35</td>
<td>5819</td>
<td>5722-5906</td>
</tr>
<tr>
<td>HW-20</td>
<td>169092</td>
<td>206</td>
<td>Wood</td>
<td>5520</td>
<td>220</td>
<td>6311</td>
<td>5753-5826</td>
</tr>
<tr>
<td>HW-25</td>
<td>169088</td>
<td>243</td>
<td>Bulk</td>
<td>6305</td>
<td>30</td>
<td>7228</td>
<td>7166-7279</td>
</tr>
</tbody>
</table>

| Note: | | | | | |
|-------|---------|---------|---------|---------|---------|---------|
| a     | Radiocarbon analysis conducted at LLNL’s CAMS facility | | | | |
| b     | Composite sediment depth of Lyman Lake, with airfall tephra removed | | | | |
| c     | Radiocarbon ages calibrated with CALIB (version 7.0.4) (Stuiver et al., 2005) | | | | |

**4.4 Paleoclimate reconstructions**

**4.4.1 PALEO-ELA RECONSTRUCTIONS**

We estimate the Lyman Glacier’s ELA at its maximum Little Ice Age (LIA) extent as ~1995 m, using a modern AABR of 5.66 combined with our reconstructed hypsometric profile (Osmaston, 2005). This AABR is relatively high; Rea (2009) calculated an average AABR of 2.38, with a maximum AABR of 4.44, for a selection of mountain and valley glaciers in the western North American Cordillera. Higher values signify high ablation gradients below the ELA and low accumulation gradients above the ELA. The Lyman Glacier’s high AABR
may relate to its hypsometry; as a small cirque glacier, its accumulation area is shaded by steep headwalls, but its lower, unsheltered extents likely experience high ablation rates. A preserved lateral moraine on the east side of the upper-basin indicates that the glacier’s LIA paleo-ELA was greater than 1954 m, consistent with our AABR results.

Comparing modern climate conditions between the Lyman Glacier’s “modern” ELA and our calculated LIA ELA suggests a difference of ~1.2 °C and 11 cm precipitation (Figure 5). However, because the “modern” ELA does not represent a steady-state condition, this difference is a minimum estimate of climate change since the LIA. Comparing conditions at the LIA ELA to conditions at modern glacier ELAs suggests that conditions at LIA maximum may have been ~2.6 °C cooler in the summer or 140 cm wetter in the winter than present conditions.

4.4.2 GLACIER-CLIMATE MODELING

Results from the glacier-climate model (Plummer and Phillips, 2003) suggest that cooling of 1.5 °C (annual average daily temperature) would produce a glacier of approximately the same extent and thickness as our reconstruction of the Lyman Glacier at its LIA maximum (Figure 11a). At this temperature depression, the modeled glacier flows slightly past the terminal moraine and exceeds 125 m (height of lateral moraine above valley) in its thickest region. Temperature depressions of less than 1.5 °C produce modeled glaciers that are too thin, whereas cooling greater than 1.5 °C produces extents much too large. However, the glacier on the western side of the upper-basin is unreasonably large at 1.5 °C, as it flows past its high-elevation LIA moraine and joins the trunk glacier past the terminal moraine (Figure 11a).
Adjusting the glacier-climate model to the modern glacier-climate relationship results in reasonable energy balance model outputs, but subsequent ice-flow modeling produces a glacier (Figure 11b) with a slightly different extent than mapped extents from either 1984 (U.S. Geological Survey, 1988) or 2014 (this study). This discrepancy may be because basal topography influences modeled ice-flow direction and the input DEM includes the relatively uniform 1984 glacier surface instead of the actual topography of the bedrock floor. Furthermore, the glacier is actively retreating and is not in equilibrium, implying that even if modern climate were to stabilize the glacier would continue to shrink and change its extent.

5.0 Discussion

5.1 Interpretation of the rock flour signal

In order to quantify the activity of the Lyman Glacier from its lake sediment record, the sediment signal must be calibrated to known glacier extents. The Lyman Glacier’s extent is documented by direct observations since 1890 CE (e.g., Freeman, 1941; Jumpponen et al., 1998; Lyman, 1909), when it was less than 30 m short of its outer-most Holocene moraine. These observations allow us to calibrate the recent lake sediment record between minimum (present-day) and maximum (1890 CE) glacier extents.

The rock flour signal in near-surface sediments peaks between 1869-1900 CE (8.5–6.5 cm sediment depth, Figure 12), broadly consistent with commonly-cited mid-19th century glacial maxima for the Cascades (Osborn et al., 2012; Pelto and Hedlund, 2001). Because this interval corresponds with the Lyman Glacier’s presence near the outer-most Holocene moraine (Jumpponen et al., 1998), we infer that the rock flour signal from 8.5–6.5 cm
characterizes the maximum Holocene rock flour production of the Lyman Glacier. The firm, bluish-gray sediments in this interval have moderate MS values (90-100 SI * 10^{-5}), but reach maximum measured values for DBD (1.34 g/cc) and FS (61%) and minimum measured for LOI (2%), while coarse silt is low (10.5%) and clay is moderate (23%). The rock flour signal declines synchronously with the glacier’s unabated retreat since 1915 CE; late-20th century sediments record MS < 70, LOI at 3.6%, DBD < 0.7 g/cc, and FS comprising only 48% of clastic sediments. Because the glacier is on the verge of disappearing, we equate its modern extent to its minimum viable extent. Thus, the Lyman Glacier’s rock flour signal is defined for its maximum and minimum extents. We employ this interpretative framework to reconstruct paleo-extents of the Lyman Glacier in Section 5.2.

In some glaciolacustrine settings, volume-normalized magnetic susceptibility is a sensitive proxy for rock flour (e.g., Bowerman and Clark, 2011). At Lyman Lake, MS co-varies with other rock flour proxies for most intervals, except during the late 19th century rock flour peak (Figure 7). The cause of the relatively low near-surface MS is unclear. Historical photographs clearly indicate the glacier was near its Holocene maximum when the low MS, rock flour-rich sediments were deposited. In other settings (e.g., the Enchantment Lakes Basin), MS trends inversely correlate with glacier extent because the source bedrock is lower in ferromagnetic mineral content than the forefield bedrock (Bilderback, 2004). Other locales exhibit an inverse correlation between mass-normalized MS and rock flour, as quantified by fine-grained sediments (Rosenbaum et al., 2012). Such relationships should remain consistent from advance to advance, which is not the case in the Lyman Lake record. Thus, the low MS in late 19th century rock flour remains unexplained.
Glacial deposition in Lyman Lake’s northern sub-basin should be restricted to the suspended fraction of clastic sediments (i.e., silt and finer), because the moraine that crosses the lake should prohibit bedload deposition. Sand intervals in the cores must therefore represent periods of increased colluvial activity on the proximal slopes of the lower-basin, where grüss exposures provide the most likely source of coarse clastic sediment. Such enhanced slope activity may result from periods characterized by intense summer precipitation. In contrast, intervals enriched in clays (but not silts) likely indicate periods of enhanced weathering but low mobilization energy. Intervals with bimodal grain-size distribution may thus reflect periods of high humidity combined with intense summer precipitation.

Although large glaciers can produce a broad distribution of grain sizes (e.g., Bakke et al 2005, 2010, 2013), the Lyman Lake record indicates that periods of enhanced rock flour production are associated with a relatively narrow grain-size distribution, concentrated in fine silt (Figure 8). This difference may reflect the small size of the Lyman Glacier; even at peak extents, its meltwater flux would be relatively small and incapable of transporting coarse grains to the lake’s northern sub-basin. In the Lyman Lake catchment basin, the association of narrow grain-size distribution with glacial maxima is similar to that observed in small cirque glaciers in Glacier National Park (Munroe et al., 2012).

Peaks in rock flour generally coincide with glacial maxima on centennial time scales (Leonard, 1997), but this relationship does not always hold at decadal scales. For some large glaciers, rock flour peaks appear to correspond to rapid retreats, presumably due to the
increased availability of glacigenic sediments for fluvial reworking and transport (Bakke et al 2005, Menounos et al 2009, Osborn et al 2007). In the case of the Lyman Glacier, the rock flour peak (1869 – 1900 CE) coincides with the documented maximum glacier extent (~1890 CE). Therefore, it appears unlikely that there is a significant lag between glacier extent and rock flour deposition.

5.2 Holocene fluctuations of the Lyman Glacier

Using the rock flour calibration established above (5.1), we reconstruct a 7,800-year record of the Lyman Glacier’s fluctuations. Similar efforts to calibrate such relationships have relied on several temporal control points where both the glacier extent and sediment signal are well-known (e.g., six for Bakke et al., 2013). For the Lyman Glacier, we have two definitive control points (maximum extent and minimum viable extent) as well as two inferred points (no glacier and an intermediate-size glacier; Table 6). Although this small number of control points limits the precision with which we quantify Holocene extents, it provides the most comprehensive continuous record yet established for the North Cascades (Figure 13, Table 7). Given the various uncertainties in our age-depth model, we round calibrated ages to the nearest hundred years.

Table 6: Rock flour signal

<table>
<thead>
<tr>
<th>Glacier Extent</th>
<th>Size (km²)</th>
<th>MS (SI*10⁻⁵)</th>
<th>LOI (%)</th>
<th>DBD (g/cc)</th>
<th>FS (%)</th>
<th>Appearance</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum</td>
<td>2.7</td>
<td>&gt; 100⁵</td>
<td>2</td>
<td>1.34</td>
<td>61</td>
<td>Bluish-gray, firm</td>
</tr>
<tr>
<td>Minimum</td>
<td>0.4</td>
<td>&lt; 70</td>
<td>3.6</td>
<td>&lt; 0.7</td>
<td>48</td>
<td>Olive-green, soft</td>
</tr>
<tr>
<td>Intermediate</td>
<td>2.7 – 0.4</td>
<td>100 – 70</td>
<td>2 – 3.6</td>
<td>1.34 – 0.7</td>
<td>60 – 50</td>
<td>Varies</td>
</tr>
<tr>
<td>Negligible</td>
<td>&lt;&lt; 0.4</td>
<td>&lt;&lt; 70</td>
<td>&gt; 3.6</td>
<td>&lt;&lt; 0.7</td>
<td>&lt; 48</td>
<td>Dark brown</td>
</tr>
</tbody>
</table>

* MS typically >100 when other proxy values are elevated

MS = magnetic susceptibility (SI * 10⁻⁵), LOI = loss-on-ignition (% of mass), DBD = dry bulk density (g/cc), Fine Silt = clastic sediment between 3.9 μm - 15.6 μm (% of sample);
Table 7: Lyman Glacier’s Holocene fluctuation history

<table>
<thead>
<tr>
<th>Time Period(^a) (cal yr BP)(^b)</th>
<th>Extent(^c)</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>7800 – 4900</td>
<td>No glacier</td>
<td>Tephra obscuration from 6000 – 5500 cal yr BP</td>
</tr>
<tr>
<td>4900 – 4200</td>
<td>Intermediate</td>
<td>Onset of Neoglacial</td>
</tr>
<tr>
<td>4200 – 3800</td>
<td>Minimum</td>
<td>Decreasing rock flour</td>
</tr>
<tr>
<td>3760 – 3400</td>
<td>Unsure</td>
<td>Tephra obscuration</td>
</tr>
<tr>
<td>3400 – 2900</td>
<td>Unsure</td>
<td>Bimodal grain size</td>
</tr>
<tr>
<td>2900 – 2600</td>
<td>No glacier</td>
<td>Negligible rock flour flux</td>
</tr>
<tr>
<td>2600 – 2500</td>
<td>Minimum</td>
<td>Increasing rock flour</td>
</tr>
<tr>
<td>2500 – 2250</td>
<td>Intermediate</td>
<td>May have persisted until 2100</td>
</tr>
<tr>
<td>2250 – 2100</td>
<td>Unsure</td>
<td>Tephra obscuration</td>
</tr>
<tr>
<td>2100 – 2000</td>
<td>No glacier</td>
<td>Negligible rock flour flux</td>
</tr>
<tr>
<td>2000 – 1800</td>
<td>Unsure</td>
<td>Tephra obscuration</td>
</tr>
<tr>
<td>1800 – 1300</td>
<td>Intermediate</td>
<td>May have advanced prior to 1800</td>
</tr>
<tr>
<td>1300 – 1200</td>
<td>Minimum</td>
<td>Decreasing rock flour</td>
</tr>
<tr>
<td>1200 – 1100</td>
<td>Unsure</td>
<td>Coarse-grained sediment signal</td>
</tr>
<tr>
<td>1100 – 900</td>
<td>Intermediate</td>
<td>May have advanced prior to 1100</td>
</tr>
<tr>
<td>900 – 800</td>
<td>Minimum</td>
<td>Decreasing rock flour</td>
</tr>
<tr>
<td>800 – 600</td>
<td>Intermediate</td>
<td>Increasing rock flour</td>
</tr>
<tr>
<td>600 – 500</td>
<td>Maximum</td>
<td>Peak rock flour</td>
</tr>
<tr>
<td>500 – 350</td>
<td>Intermediate</td>
<td>Decreasing rock flour</td>
</tr>
<tr>
<td>350 – 150</td>
<td>No glacier</td>
<td>Negligible rock flour flux</td>
</tr>
<tr>
<td>150 – 50</td>
<td>Maximum</td>
<td>Increasing rock flour; MS does not increase</td>
</tr>
<tr>
<td>50 – Present</td>
<td>Minimum</td>
<td>Decreasing rock flour</td>
</tr>
</tbody>
</table>

\(^a\) Ages from age-depth model rounded to nearest 100 years

\(^b\) Calibrated radiocarbon years (cal yr) before 1950 CE (BP)

\(^c\) Extents and time periods match graphical representation in Figure 13
The sediment record from Lyman Lake during the early Holocene (~7800 – 4900 cal yr BP) is characterized by low rock flour proxies, indicative of negligible glacier activity. Because of substantial tephra deposition related to Glacier Peak’s Dusty Creek eruptive interval, we cannot preclude the possibility of minor glacier advances during the prolonged period of tephra slopewash following the event (~6000-5500 cal yr BP).

The earliest significant period of Neoglacial activity in the Lyman cirque began at ~4900 cal yr BP, with the glacier reaching an intermediate (0.4 – 2.7 km$^2$) extent by ~4800 cal yr BP. It maintained this extent for nearly 700 years until it retreated to a minimum extent by ~4200 cal yr BP, where it persisted for at least another 400 years. The deposition of the MSH Yn tephra effectively masks the sediment signal between ~3760 – 3400 cal yr BP. From ~3400 – 2900 cal yr BP, the sediment signal contains a bimodal grain-size distribution that is not associated with glacier activity, although we lack a modern analog in the near-surface calibration record. The sediment signal from ~2900 – 2600 cal yr BP is distinctly non-glacial, indicating negligible glacier activity and the likely disappearance of the Lyman Glacier by this time.

A subsequent period of Neoglacial activity began at ~2600 cal yr BP and reached an intermediate extent by ~2500 cal yr BP. It maintained this extent for at least 250 years, at which time (~2250 cal yr BP) the sediment signal was overwhelmed by a series of co-ignimbrite ash fall deposits from Glacier Peak. The glacier may have persisted until ~2100 cal yr BP, when the tephra effects diminish enough to discern non-glacial sedimentation. Thus this mid-Neoglacial interval is relatively short, ranging from 350 – 500 years.
Intermittent eruptions from Glacier Peak obscure the sediment signal from ~2000 – 1800 cal yr BP, when a relatively strong rock flour signal indicates the Lyman Glacier was already at an intermediate extent. The glacier maintained an intermediate position until ~1300 cal yr BP, when it retreated to a minimum extent for another hundred years. Between ~1200 – 1100 cal yr BP the sediment signal is obscured by coarse-grained sediments of uncertain but likely non-glacial origin.

Directly following the coarse-grained interval (~1100 cal yr BP), the rock flour signal indicates an intermediate glacier extent. It is possible that the glacier never disappeared between ~1300 – 1100 cal yr BP and thus this interval is a continuation from the previous period. Regardless, at ~900 cal yr BP the rock flour flux waned and the glacier retreated to its minimum viable extent, where it remained for less than one hundred years before re-advancing. At ~800 cal yr BP, increasing rock flour indicates glacier growth to an intermediate extent for ~200 years, after which peak rock flour flux indicates expansion of the Lyman Glacier to its maximum Holocene extent by ~600 cal yr BP. The glacier maintained this extent for less than one hundred years before a decreasing rock flour flux indicates retreat to an intermediate position by 500 cal yr BP. The rock flour signal continued to wane, and by ~350 cal yr BP it had decreased to negligible levels, indicating that the glacier may have disappeared entirely. Thus, the glacier fluctuated from its minimum viable size to its maximum Holocene extent to a negligible presence in less than 500 years.
Following a ~200 year hiatus, the Lyman Glacier rapidly expanded beginning at 150 cal yr BP, with the rock flour signal strengthening from negligible to maximum in less than 100 years. The strength of the rock flour signal, as well as the corroborating historical record, demonstrate that the glacier reached its maximum Holocene extent, although it once again persisted for only a short duration before retreating. The historically documented retreat began circa 1900 CE and persists to the present-day (2014 CE), at which time the glacier is at its minimum viable extent. In less than 200 years, the glacier has oscillated from non-existent to maximum to minimum.

From the onset of Neoglacial to historic retreat, the fluctuations of the Lyman Glacier simultaneously increase in magnitude while decreasing in duration (Figure 13). The well-preserved stratigraphy and absence of significant bioturbation or unconformities indicate that this trend is real and not a result of preservation bias or increased resolution in younger sediments.

### 5.3 Holocene glacier activity and climate in the North Cascades

The timing of Holocene climatic transitions for most of western North America is generally poorly understood, and appears to vary significantly from region to region (Davis et al., 2009). Existing records of Holocene glaciation in the North Cascades (Figure 1) are limited to moraine records from Glacier Peak (Beget, 1984), Dome Peak (Miller, 1969) and Mt. Baker (e.g., Osborn et al., 2012), and glaciolacustrine records from the Enchantment Lakes Basin (Bilderback, 2004). Even at these sites, the process of obliterative overlap has removed nearly every moraine deposited between the Lateglacial and the Little Ice Age.
Continuous glaciolacustrine sediment records, such as in the Enchantment Lakes Basin and at Lyman Lake, offer a means for identifying non-glacial periods and thus providing a more complete picture of Holocene glacier fluctuations in the North Cascades. To supplement these glacier records, we also use reconstructed climate histories from lake sediments (Nelson et al., 2011; Prichard et al., 2009; Spooner et al., 2008, 2007; Steinman et al., 2012).

To enable comparison between records, we have standardized the reported dates. We calibrated original radiocarbon ages (CALIB version 7.1, Stuiver et al., 2005) and report the median probability calibrated age as kilo calibrated years before 1950 CE (ka). Dates from the last millennium, many of which are dendrochronologic, are reported in Common Era calendar years (CE). For lake sediment records, dates were interpolated according to the worker’s age-depth model, so we report them with no modification.

5.3.1 EARLY TO MID-NEOGlacIAL RECORDS OF THE NORTH CASCADES

The onset of Neoglacial in the North Cascades is poorly constrained but appears to vary spatially (Figure 14). The earliest documented advance is of the Easton Glacier on Mt. Baker circa (ca.) 6 ka (Osborn et al., 2012). The South Cascades Glacier advanced from ca. 5.71 ka through 5.4 ka, as recorded by tree kills in the glacier forefield (Miller, 1969). At Glacier Peak, undated till deposits from the Gamma Peak advance of the Dusty Glacier are underlain by the Dusty Creek tephra (5.7 ka) and overlain by MSH Yn tephra (3.71 ka) (Beget, 1984). We report an advance of the Lyman Glacier ca. 4.9 ka, which persists until at least 3.8 ka. In the Enchantments Lake Basin, the first Neoglacial advance is not until ca. 3.3 – 2.8 ka, about 1600 years after the Lyman Glacier’s earliest Neoglacial advance (Bilderback, 2004). All
early Neoglacial advances were significantly smaller than later Little Ice Age advances, suggesting that glacial conditions were substantially less favorable.

During the mid-Holocene, glaciers in the North Cascades advanced infrequently and at different times. Overridden trees preserved in the lateral moraines of the Deming Glacier indicate advances at ca. 2.09 ka and ca. 1.76 ka, as well as an advance ca. 1.47 ka that may be a continuation of the previous advance (Osborn et al., 2012). Meanwhile, the Lyman Glacier advanced from ca. 2.6 – 2.3 ka and again from ca. 1.8 – 1.3 ka before retreating and possibly disappearing, followed by another advance ca. 1.1 ka. The Lyman Lake record indicates that the glacier was absent from the landscape from ca. 2.9 to 2.6 ka, and again from ca. 2.1 – 2.0 ka. The sediment signal from the Enchantments also indicates a lack of glacier activity from ca. 2.8 ka through 1.0 ka (Bilderback, 2004). Although no other advances are preserved, it is likely that evidence (i.e., moraines) has been removed from the landscape by obliteratorial overlap, possibly allowing for unrecorded synchronous advances. However, a direct comparison of the two continuous lake sediment records demonstrates marked differences (Figure 14), i.e. asynchronous behavior. In summary, North Cascades glacier activity during the early to mid-Neoglacial was spatially heterogeneous, largely asynchronous, and featured prolonged non-glacial intervals.

5.3.2 LITTLE ICE AGE GLACIER RECORDS OF THE NORTH CASCADES

By the 2nd millennium Common Era (CE), glaciers throughout the North Cascades were advancing to their Holocene maxima, signifying the local onset of the Little Ice Age (Figure 14). In the Enchantments, a major advance began ca. 950 – 1150 CE with no definable retreat before emplacement of the maximum extent moraines after 1480 CE (Bilderback,
Mt. Baker’s Deming Glacier matched this behavior, advancing to its largest extent thus far (ca. 1037 CE) and then superseding it by ca. 1487 CE to reach its greatest Holocene extent (Osborn et al., 2012). The nearby Easton Glacier was advancing toward its Holocene maximum by ca. 1440 CE and probably achieved it by 1500 CE. On the north side of Mt. Baker, the Coleman Glacier began advancing ca. 1020 CE and overran a forest sometime after 1300 CE, reaching its Holocene maximum by the early 1500s CE (Heikkinen, 1984). Meanwhile, on Glacier Peak, the Streamline Ridge advance of the Chocolate Glacier is constrained between ca. 950 CE and 1650 CE (Beget, 1984). Just to the north, on Dome Peak, the Chickamin Glacier reached its Holocene maximum sometime before 1300 CE, and then re-advanced to nearly the same extent during the late 1400s CE (Miller, 1969). The Lyman Glacier kept pace as well, with an advance beginning ca. 1100 CE that matched the glacier’s maximum Holocene extent by ca. 1350 CE and sustained it into the 1400s CE. Shortly thereafter (1500s CE), three glaciers at Dome Peak (South Cascades, Le Conte and Dana) occupied their Holocene maxima as well (Miller, 1969). Thus, nearly every glacier record from the North Cascades reports maximum Holocene extents during the 1400s to early 1500s CE.

Glacier records from the second half of the Little Ice Age in the North Cascades (~1500 – 1900 CE) are sparse but suggest widespread retreat (Figure 14). At Mt. Baker, the Coleman and Easton glaciers lack dated deposits from their ca. 1500 CE maxima until the mid-1700s and mid-1800s, respectively (Heikkinen, 1984; Osborn et al., 2012). Evidence of a retreat of the Deming Glacier is less clear, but there are no dated deposits during the 1700s. At Glacier Peak, numerous smaller moraines from the Dusty and Chocolate glaciers are presumed to be
from the last few centuries, but have not been dated (Beget, 1984). The Chickamin Glacier on Dome Peak may have retreated between its late 1400s maximum and an advance in the 1600s, and definitively lacks dated moraines post-1600s until the late 1800s CE (Miller, 1969). Meanwhile, the other three Dome Peak glaciers all exhibit ~300 year gaps between their mid-millennium maxima and the mid-1800s. The Enchantment Lakes Basin was inundated with the Wn tephra from Mt. Saint Helens at 1480 CE, preventing comparison (Bilderback, 2004). We report the Lyman Glacier’s retreat beginning ca. 1450 CE and absence between ca. 1600 – 1800 CE, followed by re-advance to its Holocene maximum extent in the late 1800s. The Little Ice Age in the North Cascades peaks in the 1800s with many glaciers re-advancing after two to three hundred years of diminished presence.

Glacier activity in the North Cascades during the Little Ice Age is striking for its regional uniformity. Unlike the sporadic onset of Neoglacialiation, the LIA onset was marked by regionally synchronous advances. North Cascades glaciers reached their maximum Holocene extents within a ~100 year span (~1400 – early 1500s CE). The region then experienced roughly synchronous retreats, with nearly every studied glacier either disappearing or lacking dated deposits from the mid-1500s to the early 1800s CE. The final LIA advance was also broadly synchronous, featuring advances that nearly matched maximum Holocene extents by the late 1800s CE, at which time regional retreat began.

5.3.3 HOLOCENE CLIMATE RECORDS OF THE NORTH CASCADES

Lacustrine climate records from lower elevations in the North Cascades and locations further north and east (Figure 1) give additional context to the North Cascades Neoglacial record. At Thunder Lake (elev. 421 m) (Spooner et al., 2007), Ridley Lake (elev. 957 m) (Spooner et
al., 2008), and Panther Potholes (elev. 1100 m) (Prichard et al., 2009), climate histories were reconstructed using pollen assemblages. Just across the international border from the North Cascades, July temperature anomalies were reconstructed based on chironomid records from four subalpine lakes (Gavin et al., 2011). Approximately 120 km to the east, oxygen isotope records from Castor Lake (elev. 594 m) were used to reconstruct precipitation cycles (Nelson et al., 2011) and winter precipitation (Steinman et al., 2012).

Beginning with warm and dry conditions in the early Holocene, each site transitions towards cool, wet conditions. Between ca. 7.0 – 6.0 ka, the Canadian subalpine sites cool by ~1.5 °C, while lower elevation Ridley Lake (just south of the border) is interpreted to have had modern, cold-adapted forests since 7.0 ka. Further south, Thunder Lake does not see such forests till 6.0 ka, while the higher elevation Panther Potholes becomes cooler much earlier, ca. 7.7 ka. Neoglacial climate signals begin to emerge in correspondence with the first advance of the Lyman Glacier (ca. 4.9 ka), as Canadian sites exhibit a further 1 °C drop from ca. 5.0 – 4.0 ka, and pronounced cooling is inferred at Panther Potholes beginning ca. 4.5 ka. Notably, no evidence of a distinct Neoglacial cooling is found at Thunder and Ridley Lake. These records suggest that the Neoglacial climate signal emerges at an earlier date dependent on latitude and elevation.

Climate records indicate that conditions became cooler and wetter through the mid-Holocene (ca. 4.5 – 2 ka) (Gavin et al., 2011; Prichard et al., 2009). The Lyman Lake record indicates that the glacier was absent from the landscape from 2.9 – 2.6 ka, and again from 2.1 – 2.0 ka. This recorded absence suggests that unfavorable local conditions overwhelmed the regional
climate signal. Regional cooling became even more pronounced around 2.0 ka with the arrival of Alaska yellow-cedar, a species well adapted to cold, wet conditions (Prichard et al., 2009). Shortly thereafter, July temperature reconstructions exhibit a Holocene minima at ca. 1.5 ka (Gavin et al., 2011). Yet, despite glacially-favorable regional conditions, glacier advances are mostly absent from the record during this time, either due to obliterative overlap or unfavorable local conditions.

The dramatic glacier fluctuations observed during the Little Ice Age are consistent with the isotope records from Castor Lake. The onset of the Little Ice Age, and the resultant regional glacier advance, coincided with unusually wet winters from ca. 900 – 1300 CE (Steinman et al., 2012). As the LIA progressed, glacier fluctuations increased in magnitude and approached centennial intervals while drought and pluvial cycles increased in magnitude and lengthened from decadal to centennial periods (Nelson et al., 2011). The inferred late-LIA regional retreat (ca. mid-1500s to early 1800s CE) corresponds with unusually dry winters at Castor Lake from ca. 1450 – 1850 CE. As has been observed elsewhere in the world (Clague et al., 2009), climate during the Little Ice Age in the North Cascades changed with greater frequency and intensity than at any other time during the Holocene.

5.3.4 PATTERNS IN GLACIER BEHAVIOR

Spatially variable characteristics such as latitude and precipitation may influence the relative timing of glacier activity within a region. In the North Cascades, later onsets of Neoglacial advances are associated with lower latitudes and increasingly continental climate conditions. The earliest Neoglacial advance occurred ~85 km west of the Cascade crest on a Mt. Baker
glacier, whereas the latest Neoglacial onset is at the Enchantment Lakes Basin, ~60 km east of the crest and ~150 km to the south of Mt. Baker. Intermediate onsets of Neoglacialiation occur amongst glaciers adjacent to the crest and in the central North Cascades. A similar pattern is seen in lacustrine climate records, where the northern-most lakes record earlier cooling than lakes further south.

The two continuous glacier records from the North Cascades further demonstrate the importance of latitude and continentality (Figure 14). The Lyman Glacier advanced 1600 years before the earliest Neoglacial activity in the Enchantments, and subsequent Lyman Glacier advances between 2.6 ka and 1.0 ka were not matched by activity in the Enchantments. The Enchantments are ~120 km further south than the Lyman Glacier and are ~60 km further east of the Cascade crest than the Lyman Glacier. Together, these spatial factors create a less favorable glacial climate in the Enchantment Lakes Basin. Modern conditions lend additional support; as of 2014, none of the Enchantments’ cirques support active glaciers, whereas the Lyman Glacier is still flowing.

The more favorable glacier conditions at higher latitude and maritime sites correspond to larger glaciers (e.g., Mt. Baker’s Easton, Coleman and Deming Glaciers). Because larger glaciers tend to be more closely tied to regional climate conditions as opposed to local orographic effects (Leonard, 1989), climate changes such as mid-Holocene cooling are reflected first in larger glaciers. In contrast, smaller cirque glaciers (as are common in the non-volcanic North Cascades) are dominantly influenced by local variables such as aspect, concavity, avalanching and snowdrift (Graf, 1976). As a result, a greater magnitude of
climate deterioration was required for these cirques to transition from marginal to favorable glacier conditions and thus trigger Neoglacial advances. The observed patterns in Neoglacial behavior suggest that glaciers throughout the North Cascades were influenced by different signals during the early to mid-Neoglacial, but that a dominant signal emerged during the last millennium that caused glacier activity to synchronize.

5.4 Glacier synchrony in response to climate change

5.4.1 Glacier-climate synchrony thresholds

Although many regional glacier studies emphasize centennial-scale Neoglacial synchrony (e.g., B. H. Luckman, 2001), some studies indicate regional asynchrony (e.g., Matthews and Karlén, 1992). Indeed, as Clague et al. (2009) note, regional summaries often lump individual, non-coincident events into single named advances, fostering the appearance of synchrony. Our analysis of the North Cascades demonstrates that although centennial-scale synchrony occurred during the Little Ice Age, glaciers fluctuated in a much more stochastic pattern during the earlier Neoglacial (6.0 – 1.0 ka). This distinct difference in glacier behavior between the two intervals suggests that synchronous fluctuations may result from regional climate surpassing a “glacier-climate synchrony threshold”.

Asynchrony of glacier fluctuations may in fact be the norm. During near steady-state climate conditions, modeling results demonstrate the potential for centennial-scale glacial asynchrony, with individual advances of magnitudes that could rival Holocene maxima (Huybers and Roe, 2009). Even when glacier mass-balance trends across a region appear to covary over many decades (Pelto and Riedel, 2001), differing mass balance magnitudes may
be compounded on a multi-decadal time scale and result in asynchronous fluctuations (Fountain et al., 2009). Fundamentally, whenever a locally variable system (e.g., weather) drives a regional system with a long-term memory (e.g., glaciers in the North Cascades), the results should reflect the stochastic nature of the inputs (Roe and O’Neal, 2009). In other words, absent a strong and persistent regional climate shift, asynchronous fluctuations would be expected among glaciers in a given region.

In the North Cascades, glacier synchrony (or lack thereof) appears to result from the interaction of regional climate and locally stochastic weather (particularly precipitation). During times of moderate regional climate, (e.g., 6.0 – 1.0 ka in the North Cascades), the local factors dominate, resulting in a stochastic pattern of fluctuations. It is only when the magnitude of regional climate change is sufficiently large so as to overwhelm local factors (Metcalfe et al., 2015) that the region will experience synchronous glacier fluctuations.

5.4.2 HOLOCENE SYNCHRONY THRESHOLD OF THE SOUTHERN COAST MOUNTAINS

The southern Coast Mountains of British Columbia, which adjoin the North Cascades, provide a good location to test this concept of a “glacier-climate synchrony threshold” because they have perhaps the most complete record of Holocene glacier fluctuations in the North American Cordillera. Moraine-based records in these mountains are well dated, in part because the relatively low-elevation Holocene glaciers repeatedly overrode forests (Mood and Smith, 2015). Glaciolacustrine records are also abundant and commonly agree with the moraine records (Menounos et al., 2009).
Early Holocene advances in the southern Coast Mountains appear to be asynchronous (Menounos et al., 2009), although they are not described as such. In Garibaldi Provincial Park, a glaciolacustrine record from Green Lake exhibits increased clastic sedimentation ca. 7.0 – 6.6 ka (Osborn et al., 2007), whereas nearby Black Tusk Lake records upstream glacier activity between roughly 5.9 – 4.9 ka (Cashman, 2004). Elsewhere in the Coast Mountains, lake sediments indicate glacier activity ca. 6.5 ka at lower Joffrey Lake (Filippelli et al., 2006) and between 6.0 – 5.0 ka at Klept Lake in the Kwoiek Creek watershed (Souch, 1994). Meanwhile, three other lakes in the same watershed show no evidence of early Holocene glacier activity. This period of spatially and temporally irregular glacier responses appears similar to the asynchronous early Neoglacial glacier behavior observed in the North Cascades. Such behavior suggests that the southern Coast Mountains’ climate signal was not strong enough to drive a synchronous response from its glaciers during the early Holocene.

Following this early period of episodic glacier advances, glaciers appear to synchronize ca. 3.5 ka (Menounos et al., 2009). Clastic sedimentation in Green Lake peaks ca. 3.6 – 3.4 ka while lower Joffrey Lake’s sediment signal peaks at 3.5 ka; both remain elevated for the remainder of the Holocene, punctuated by pulses of increased glacigenic sedimentation (Filippelli et al., 2006; Osborn et al., 2007). Black Tusk Lake records a clastic peak between ca. 3.8 – 3.3 ka followed by a brief decline, but then exhibits a sustained increase in clastic sedimentation by ca. 2.6 ka (Cashman, 2004). The four Kwoiek Creek lakes received a sustained increase in clastic sediments between ca. 4.0 – 3.0 ka, and again over the last two millennia. Although these lakes do not exhibit centennial-scale synchrony, Souch (1994) argues that this may be an artifact of uncertainty in the dating methods. Still, the consistency
of initial peaks in glacial sedimentation ca. 3.5 ka suggests that the region crossed its glacier - climate synchrony threshold.

Glacier fluctuations in B.C.’s southern Coast Mountains were asynchronous during the mild early Holocene but became synchronized ca. 3.5 ka (Menounos et al., 2009), suggesting that the region crossed its glacier – climate synchrony threshold. If true, this would indicate that these mountains crossed their synchrony threshold ~2500 years earlier than the North Cascades. It appears that the southern Coast Mountains, which are further north and far more heavily glaciated than the North Cascades, required less climatic perturbation to achieve synchronistic glacier behavior.

**5.5 Climate change in the North Cascades**

The onset of synchronous glacier behavior throughout the North Cascades at ca. 1.0 ka required a sufficiently strong and regionally consistent climate shift to overwhelm local variability. Whether temperature or precipitation changes were the primary drivers of this shift remains uncertain. North Cascades climate records demonstrate progressive cooling since the early Holocene (Gavin et al., 2011; Nelson et al., 2011; Prichard et al., 2009), driven in part by continuously decreasing summer insolation (Solomina et al., 2015). Precipitation in the North Cascades also increased over the early to mid-Holocene (Prichard et al., 2009), but was highly variable in magnitude during the last millennium (Nelson et al., 2011). Precipitation tends to be spatially highly variable, as suggested by modeling results of the North Cascades (Huybers and Roe, 2009). Although the high relief of the Himalaya likely results in more dramatic variance in precipitation, fivefold changes in precipitation over distances less than 10 km (Anders et al., 2006) suggest that even the North Cascades
should exhibit substantial spatial variability of precipitation. Temperature, in contrast, varies more systematically with altitude and latitude and is thus more regionally consistent (Huybers and Roe, 2009). Because the behavior of North Cascades glaciers becomes regionally consistent at ca. 1.0 ka, we infer that such synchrony was driven by a regional change (i.e., temperature decrease), as opposed to more locally influenced factors such as precipitation. Such temperature-driven synchrony is also consistent with late 20th century glacier behavior in the North Cascades, as anthropogenic warming has driven the largely synchronous retreat despite a slight increase in precipitation (Pelto, 2008).

The fluctuation record of the Lyman Glacier provides a means to quantify the cooling responsible for inducing synchronous glacier activity in the North Cascades. During the last millennium, the Lyman Glacier twice advanced to its Holocene maximum extent. If temperature was the primary driver of glacier synchrony during the last millennium, as we reason above, then by inference the Lyman Glacier’s maximum advances represent periods of maximum Holocene cooling. According to our ELA reconstruction, summer temperatures during the maximum LIA extent at the Lyman Glacier were 2.6 °C colder than present (1989-2010 CE), and our glacier-climate modeling suggests yearly mean temperatures were 1.5 °C cooler. Given that the Lyman Glacier’s two LIA advances were matched by most other glaciers in the region, we conclude that summer cooling of 2.6 °C or annual cooling of 1.5 °C was sufficient to drive regional glacier synchrony.

Although regional cooling was probably the dominant driver for the synchronous LIA advances, it cannot explain the inferred regional late-LIA retreat (ca. mid-1500s – early
1800s CE), because cool conditions persisted globally (Mann et al., 2009) and regionally (Prichard et al., 2009). Furthermore, tree-ring records from the western foothills of the North Cascades suggest that high ablation rates did not drive the late-LIA regional retreat, because these summer temperature-sensitive records exhibit no warming trends during this period ((Robertson, 2011)). Changes in precipitation, if regionally consistent and large enough in magnitude, may explain the late-LIA retreat. The oxygen isotope record from Castor Lake, which records winter precipitation, indicates an unusually dry period (ca. 1450 – 1850 CE) that coincides with the late-LIA retreat. Further analysis of ten lakes throughout western North America confirms that this dry period was regionally extensive (Steinman et al., 2014). Unlike tree-ring reconstructions of summer precipitation (Stahle et al., 2007), which can exhibit drought during glacier advances (Steinman et al., 2012), winter precipitation-sensitive records should more closely match glacier behavior.

6.0 Conclusion
In this study, we have reconstructed the Holocene fluctuation history of the North Cascades’ Lyman Glacier by measuring the rock flour flux recorded in sediments of proglacial Lyman Lake (Figure 13, Table 7). The glacier was absent from the early to mid-Holocene (~7.8 – 4.9 ka) before beginning its earliest Neoglacial advance at ~4.9 ka, which lasted until ~3.8 ka. Between ~3.8 – 2.9 ka glacier activity is uncertain, but from ~2.9 – 2.6 ka there is evidence of a distinct lack of glacier activity. Glacier activity resumed ~ 2.6 ka – 2.25 ka, but ceased again between at least ~2.1 – 2.0 ka. Tephra deposits obscure the glacial record both before and after this non-glacial interval. By ~1.8 ka, however, rock flour again indicates the Lyman Glacier was present and probably remained so until ~0.35 ka, although
the record is unclear from ~1.2 – 1.1 ka. Following a brief non-glacial interval (~0.35 – 0.15 ka), the glacier re-advanced between ~0.15 – 0.05 ka before retreating to its present-day extent. In contrast to the comparatively small advances during the early to mid-Neoglacial, the glacier reached its maximum Holocene extent during the latter two (LIA) advances at ~0.6 – 0.5 ka (~1350 – 1450 CE) and ~0.15 – 0.05 ka (~1800 – 1900 CE). Glacier-climate modeling (Plummer and Phillips, 2003) indicates that these LIA maxima were driven by cooling of 1.5 °C relative to the late-20th century, whereas paleo-ELA reconstructions suggest a decrease in mean summer temperatures of 2.6 °C (Osmaston, 2005).

Comparison of the Lyman Glacier’s record to other published records of climate change in the region provides new constraints on Neoglaciation in the North Cascades. Notable findings include the following:

1) North Cascades glacier activity was largely asynchronous during the early to mid-Neoglacial (~6.0 – 1.0 ka)

2) During the last ~1.0 ka (the Little Ice Age), glaciers throughout the North Cascades have fluctuated synchronously, in association with increasing frequency and magnitudes of glacier and climate fluctuations

3) Many North Cascade glaciers achieved their maximum Holocene glacier extents between ~1400 – early 1500s CE (~0.55 – 0.45 ka). A subsequent near-maximum advance occurred during the 1800s CE (~0.15 – 0.05 ka) after a 200 – 300-year interval of winter drought that extinguished many glaciers.

The distinct switch from largely asynchronous to synchronous glacier behavior in the North Cascades at ~1.0 ka indicates the region crossed a glacier-climate synchrony threshold at that
time. The regional nature of this shift suggests that it was driven by temperature rather than precipitation. Rapid anthropogenic warming during the past century has also induced a synchronous response, as the glaciers of the North Cascades retreat en masse in spite of locally variable factors such as precipitation.
7.0 References


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8.0 Figures

Figure 1: Place map of North Cascades with names of referenced locations overlain on shaded terrain. Inset map shows volcanoes that have deposited prominent tephras in the North Cascades during the Holocene. Asterisks denote study sites.
Figure 2: Geomorphic map of Lyman Lake catchment basin on hillshade created from USGS 10m DEM (based on topography from 1984 aerial imagery), elevation in meters. Glacier extent indicated as of 2014. Inset map shows bathymetry of Lyman Lake, depth in meters.
Figure 3: View to southeast of Lyman Glacier from Cloudy Pass, August 12th, 2014. Photograph shows separation of glacier into upper and lower lobes. Lower lobe is fed in part by avalanching snow, ice and rock from upper lobe and cirque headwall (note debris streaks on lower lobe). Blue lines indicate outer limits of Holocene moraines.
Figure 4: Historic photograph of the Lyman Glacier, taken by W.D. Lyman in 1901, view to southeast from Cloudy Pass (Lyman, 1909). The glacier’s terminus sits on or directly behind the terminal moraine (circled), which appears as a thin dark line at base of ice. Note the continuous glacier ice flowing (arrows) from Spider Gap and Chiwawa Mountain into main lobe, in contrast to the glacieret (at right) that terminates well above the Lyman Glacier.
Figure 5: Modern climate conditions at the Lyman Glacier’s modern and Little Ice Age equilibrium-line altitudes plotted on Leonard’s (1989) envelope of modern glacial conditions. Modern ELA based on location of firn line at 2300 m in August of 2014. Little Ice Age ELA calculated via AABR method (Ostmaston, 2005). Temperature at each ELA is extrapolated from the June 1st – August 31st (1989 - 2010) daily recorded average at Lyman SnoTel, using a PRISM-derived June-Aug lapse rate. Similarly, precipitation is extrapolated from the average total SWE value (October 1st – April 30th, 1989 - 2010) recorded at the Lyman SnoTel using a PRISM-derived Oct-Apr lapse rate.
Figure 6: Comparison of measured temperature (Buehn, 2014) and modeled shortwave radiation (Plummer & Phillips, 2003) at the Lyman SnoTel site (1823 m) with measured parameters at the USGS Hut (1842 m) above South Cascade Glacier (Anslow et al., 2008; Bidlake et al., 2010, 2007, 2005, 2004). A) average monthly temperature values from water years (WY) 2002 - 2007, measured at the Hut (red line) and the SnoTel (blue line). B) measured incoming shortwave radiation (red line) at the Hut (WY 2002 - 2007) versus modeled incoming shortwave radiation (blue line) at the SnoTel. The modeled output is based on climate inputs from 1981-2010 (Table 1).
Figure 7: Composite sediment record of Lyman Lake cores, with measurements of rock flour proxies. Intervals dominated by tephra have been omitted (dashed lines) to emphasize non-volcanic clastic sedimentation within the Lyman Lake catchment basin. Proxy values are plotted so that shifts to the right indicate greater rock flour flux and shifts to the left indicate decreased rock flour flux. MS data has been smoothed to a 5-point moving average (2 cm), raw data displayed as gray line. MS = magnetic susceptibility ($SI \times 10^5$), LOI = loss-on-ignition (% of mass), DBD = dry bulk density (g/cc), Fine Silt = relative component of clastic sediment between 3.9 μm - 15.6 μm (% of sample);
Figure 8: Composite sediment record of Lyman Lake cores, with measurements of grain-size distribution. Intervals dominated by tephra have been omitted (dashed lines) to emphasize non-volcanic clastic sedimentation within the Lyman Lake catchment basin. 

Clay = relative component of clastic sediment less than 3.9 μm (% of sample); Fine Silt = % of sample from 3.9 μm - 15.6 μm; Coarse Silt = % of sample from 15.6 μm - 62.5 μm; Fine Sand = % of sample from 62.5 μm - 250 μm;
Figure 9: Calibrated $^{14}\text{C}$ ages (CALIB v. 7.0.4, Stuiver et al., 2005) and tephra ages plotted vs composite sediment record of Lyman Lake cores. Tephra ages (red) include both published and unpublished sources (Table 5). Tephras and terrestrial macrofossils are fit to a $3^{rd}$ order polynomial. Bulk carbon samples are fitted to a linear regression (dashed line). Outliers are not included in trendline calculations. Error bars represent 95% confidence of calibrated ages.
Figure 10: Bayesian accumulation (“Bacon”) age-depth model (Blaauw and Christen, 2011) for composite sediment record of Lyman Lake cores. Red center line is the best fit based on weighted mean age for each depth. Darker shading corresponds to higher probability, and the bounding grey stippled line is the 95% confidence limit. Calibrated radiocarbon ages and depth ranges shown in purple (calibrated by Bacon). Green age-depths at top of stratigraphy are calendar years from $^{210}$Pb analysis, and green age-depths lower in stratigraphy are calendar years from tephrochronology (Table 5). Default settings adjusted: “memory” sensitivity decreased from 0.7 to 0.3, and “hiatus” lowered from 1000 to 10.

Blue line on secondary y-axis shows sedimentation rate (cm/100yrs), as calculated from age-depth model. Dashed blue line shows average sedimentation rate (3.8 cm/100 yrs) over entire interval. Sedimentation peak at ~2600 - 2700 cal yr BP is associated with non-glacial, non-volcanic coarse-grained sediment influx.
Figure 11: Glacier-climate model (Plummer & Phillips, 2003) reconstructions of A) the Lyman Glacier’s Little Ice Age maximum extent, driven by cooling of 1.5 °C; and B) the Lyman Glacier’s modern extent (from 1981-2010 climate data).
Figure 12: Composite sediment record of near-surface sediments from Lyman Lake cores, with measurements of rock flour proxies. Turquoise shading highlights peaks in rock flour flux.

\[ MS = \text{magnetic susceptibility (SI } \times 10^{-5}) \], \[ LOI = \text{loss-on-ignition (\% of mass)} \], \[ DBD = \text{dry bulk density (g/cc)} \], \[ \text{Fine Silt} = \text{relative component of clastic sediment between 3.9 \( \mu \text{m} \) - 15.6 \( \mu \text{m} \) (\% of sample)} \];
Figure 13: Reconstructed paleo-extent of Lyman Glacier from ~7.8 ka to present (solid blue line, km$^2$). Reconstruction derived from sediment signal in Lyman Lake cores. Also shown are periods of significant rock flour flux from Lyman Glacier (blue dots), periods lacking rock flour (green diagonals), and intervals of significant tephra slopewash (red highlights). When possible, the sediment signal underlying the slopewash is displayed as well (dots or diagonals underneath red highlights). Unresolved sediment signals are reflected with question marks. Ages are from age-depth model (Figure 10), and can be seen in Table 7.
Figure 14: Summary diagram of Holocene glacier records from the North Cascades (Figure 1 for locations). Blue shading indicates confidence of record. With the exception of the first two records (Lyman Glacier and Enchantment Lakes Basin), the remaining glacier records are discontinuous and do not show non-glacial periods. Stars denote maximum Holocene extents. Note that time scale (cal yr BP) is not linear, and contains a secondary x-axis for the last 1000 years (1000 - 2000 CE).
Appendix 1: Sediment Core Analysis

Unit descriptions

Stratigraphic divisions are based on rock flour flux and magnitude. Rock flour flux begins low during the pre-Mazama early Holocene, increases moderately during the mid-Holocene (~4900 cal yr BP), becomes consistently high after the MSH Py deposit (~2600 cal yr BP), peaks at the onset of the Little Ice Age (14th century), decreases during the middle of the LIA, and once again peaks during the 19th century before falling off during 20th century warming. Despite removing tephra layers from the analyzed data, slopewash continues to impact rock flour proxy measurements for up to 20 cm post-deposition.

Unit A (273.5 – 236.5 cm) is a dark olive-brown massive gyttja with low rock flour proxy values. Mazama tephra slopewash causes elevated rock flour proxy values for ~20cm post-deposition. It has a minor upwards trend in rock flour proxy values that ends 5cm before deposition of Layer D of the Dusty Creek eruptive interval.

Unit A Slopewash (236 – 219 cm) is a heterogeneous sequence of light yellow-tan clastic layers that grade into an olive-gray fine-grained clastic unit with frequent gyttja layers. Rock flour proxy values are strongly influenced by the Dusty Creek tephra, spiking immediately after deposition before descending towards moderate values. The tephra fall ends as FS, and so the grain-size of the slopewash begins as FS, increases to fine sand, and then descends through silt to clay at 221cm, 15cm above the tephra.

Unit B (219 – 196cm) is a massive reddish-brown gyttja with decreasing rock flour flux, although proxy values are still above Unit A levels. The interval ends with an upwards trend in LOI and DBD. Grain-size fluctuates from coarse silt to sand to FS and again to sand.

Unit C (196 – 164cm) is a laminated blue-gray coarse clastic unit in its lower extent that becomes more fine-grained up-section. Rock flour proxy values begin at their highest values yet (excluding tephra-influenced data), although MS and DBD decline at the upper end while gyttja layers increase in frequency. The unit is truncated by the Mount St Helens Yn tephra.

Unit C Slopewash (163.5 – 147cm) is a yellowish-olive well-laminated fine-grained clastic interval with frequent gyttja layers. Rock flour proxy values are elevated from Unit C, although MS decreases. The grain-size exhibits a similar decline as in Unit A Slopewash; it begins as coarse silt overlying sandy Yn tephra, and sequentially descends to FS and ends as clay at 149.5cm, 14cm above the tephra.

Unit D (147 – 111cm) is a brownish-olive well-laminated gyttja-rich interval. Rock flour proxy values decrease to Unit B levels (low) before returning to unit C levels (moderate). Coarse silt and sand match the higher rock flour levels on either end, with FS and clay corresponding to the parameter trough in the middle.

Unit E (111 – 89.5cm) is a bluish-olive poorly-laminated fine-grained interval with high rock flour proxy values that is strongly influenced by the three co-ignimbrite ash fall deposits
within it. However, proxy values are rising above Unit C levels before the first tephra fall, and remain elevated after the third fall at 95cm. The color retains its bluish hue after the 3rd fall as well. Grain-size begins fine, with clay dominating as rock flour peaks at 102.5cm. As with the proxy values and color, the fine grain-size is restored after the 3rd fall. The slopewash contribution appears to rise in the next interval.

Unit E Slopewash (89.5 – 74.5cm) is a heterogeneous interval with two more co-ignimbrite ash fall deposits that varies from olive-gray to reddish-brown to bluish-olive. Rock flour proxy values begin on the descent, but are elevated after the second deposit. Grain-size descends from sandy tephras to silt to clay at 76.5cm, signaling the end of the slopewash.

Unit F (74.5 – 37cm) is a well-laminated fine-grained interval with frequent gyttja layers and varying hues from bluish-gray to olive to brown. Rock flour proxy values are consistently elevated, although MS descends to Unit D levels. Grain-size begins with a clay-sized peak and has two prominent coarse peaks; however, the FS component stays elevated throughout.

Unit G (37 – 20.5cm) is characterized by the largest spike in rock flour proxy values of any non-tephra-influenced interval thus far. It begins as a well-laminated bluish-gray coarse-grained interval that abruptly increases in parameter values. At the rock flour peak, FS and clay dominates. Immediately afterwards, as rock flour proxy values begin to wane, the grain-size coarsens and the bluish-gray lamina are replaced by massive olive-brown deposits. Gyttja is sparse throughout.

Unit H (20.5 – 12cm) is a well-laminated reddish-brown gyttja-rich interval with very low rock flour proxy values. Although the FS component remains high, clay becomes the dominant grain-size as coarse grains disappear entirely.

Unit I (12 – 0cm) surpasses Unit G in the magnitude of its rock flour spike. Similarly, it begins as a well-laminated bluish-gray silty interval that transitions to finer grains at the peak. The rock flour decline is precipitous, and the FS declines with it, being replaced by coarse silt. Notably, the MS stays low throughout the interval, in contrast to LOI and DBD.

Sediment data

LymanLake_Stratigraphy.xlsx

Sediment cores were measured at numerous points in the process of retrieving, transporting, splitting, and imaging. Two unmodified depths are presented (column headers “Visual Section Depth” and “XYZ Section Depth”), taken from LacCore’s visual scanning instrument and from LacCore’s Bartington 2SCE instrument, respectively. These raw depths were used to compile a composite sediment record (column header “Reference Depth”), primarily from the 2nd Livingstone core sequence (Liv2-01 through Liv2-04). At core boundaries, areas of liquefaction, the top and the bottom, sections from Liv3 and Liv1 were used for data. The top 2cm of the stratigraphy were not recovered by the Livingstone cores; 210Pb analysis demonstrated that the Glew cores recovered 2cm that were younger than the uppermost layer in the Livingstone cores.
*Master_Stratigraphy* is a composite sediment record from multiple Livingstone cores that has had visible tephra intervals deleted, and additionally has had the 1cm above each tephra interval withheld. Note that it lacks the uppermost 2cm, as the Livingstone cores did not recover the upper-most sediment. This is the sediment record that is used for all rock flour proxy data (hence the withholding of tephra-influenced data). MS, LOI and DBD values are presented along with the source core. If multiple cores are used for one depth, the 2nd core’s rock flour analysis data is presented as well, signified by “Correlation” header. “XYZ” headers signify data is from LacCore’s Bartington 2SCE instrument. “Visual section depth” refers to depths derived from, and are associated with the Corelyzer imaging program.

*Surface_Stratigraphy* is the predecessor to *Master_Stratigraphy*. It contains the complete compiled sediment record from multiple cores, and all tephra layers remain included. This is the stratigraphy used for tephra analysis. *Tephra_Removal* is the intermediary step from Surface to Master Stratigraphy. Tephra intervals have been highlighted and removed, but the additional 1cm above each interval has not been withheld.

**RockFlourProxy_Data.xlsx**

Spreadsheet containing seven worksheets. Primary rock flour proxy measurement (MS, LOI, DBD, Fine Silt %) are highlighted “rock flour turquoise”. Values are compiled from multiple cores, and are presented against the master stratigraphy, usually referred to as “Depth (tephra removed)”. Note that MS at a given point is a five-point running average.

*RockFlourProxies* contains values of primary rock flour proxy values used for analysis, with values every 0.5cm where available. Note that depth and parameter values are compiled from multiple sources. *MS* and *DBD_LOI* contain every data point for their respective parameters along with the source core. *Grain_Size* contains all grain-size data, including many analyses not included with the primary rock flour parameters. Note that it is sorted via “Depth (tephra removed)”, but that all tephra grain-size data points are included as well, below the non-tephra points. *Proxy_Correlations* is intended for analysis of intradata relationships. *RockFlour_byInterval* contains an averaged value of each proxy for each described stratigraphic interval. Values are plotted after being rescaled to a 0 -1 spectrum. *RockFlour_byInterval_noTephra* has the same content, but all intervals influenced by tephra slopewash have been removed from the analysis to facilitate comparison of rock flour proxies.

**Livingstone_MS_WWU.xlsx**

After transport from the field to WWU, the Livingstone sediment cores (LL14-Liv1, LL14-Liv2, LL14-Liv3) were measured for magnetic susceptibility using a Bartington MS2-C. Results are presented for each core.

**LacCore_Data (folder)**

Raw data from LacCore is presented within this folder. MSCL files came from LacCore’s Geotek MSCL-S, which performs loop measurements (including MS and gamma density) on whole cores. XYZ files were produced by LacCore’s Geotek MSCL-XYZ, which performs point measurements with a Bartington MS2-E magnetic susceptibility sensor and uses a color spectrophotometer to record grayscale, reflectance and color data. The **Images** sub-folder
contains high-resolution jpegs for every core section. Finally, raw MS data for the Glew cores is presented as well. MS measurements were made in the field with a Bartington MS2-C loop sensor and at LacCore with the Bartington MS2-E point sensor.

There is tremendous potential for additional investigation of the Lyman Lake cores with spectral analysis (see XYZ data files). The other cores from nearby locales have not been analyzed in any way, and are rich with information. Note that individual cores have both field names and LacCore names (e.g., field LL14-Liv3-01 = LacCore LYMAN-LYM14-3A-1L-1-W). See LMAN_MSCL.xlsx for translation from “Original ID” (field name) to “SECT NAME” (LacCore name). The suffixes “A” or “W” in some data files indicate that the core has been split, and now has an Archive half and a Working half.

LYMAN_XYZ_adjusteddepth_forCorewall contains all data from LacCore’s multi-spectra instrument (“XYZ”) for the three Lyman Lake core sequences. This includes MS as well as CIE L*a*b* values (a measurement of color), Greyscale and Reflectance values. There is great potential for analysis using these measurements! The format of the data has been modified for easier display in Corewall’s Corelyzer software. When used in the master stratigraphy, depths have been adjusted first for compressions and extensions of core (“adjusted depth”) and finally to the master stratigraphy (“Reference Depth”, cross-check to RockFlourParameters in RockFlourParameter_Data.xlsx). “Core Depth” describes the measured depth by the XYZ instrument. Lyman_XYZ.xlsx and Lyman_MSCL.xlsx contain raw data from LacCore’s MSCL-XYZ and Geotek MSCL-S, respectively, for all cores including non-lake sites. The MSCL-S contains gamma density as well as MS via loop measurements. Both these files have been formatted for display in Corewall as well (LYMAN_XYZ_forCorewall.xlsx and LMAN_MSCL_forCorewall.xlsx). The folder Images contains two sub-folders, one (“jpeg”) containing raw images of every core section, and the second (“rbg”) contains digitized color values. I highly recommend downloading Corelyzer from CoreWall.org – free and remarkably useful for displaying core imagery.

MagneticSusceptibility_GlewCores.xlsx contains MS data for both Glew cores, measured in the field (loop) and at LacCore (point). Depths are presented both raw and adjusted to the master stratigraphy. Finally, comparisons with Livingstone-3 are used to compile to master stratigraphy.

Appendix 2: Age-Depth Model for Lyman Lake

Sediment_Chronologies.xlsx

14C_Data contains the raw AMS radiocarbon results from LLNL. Calibrated_14C_2sig contains calibrated ages, from CALIB. 210Pb_Glew1 and 210Pb_Liv3 contain results from the gamma spectrometer, input into a spreadsheet created by Dave Schultz (WWU) that calculates sedimentation rates. The near-surface sediment chronology is derived from 210Pb_Glew1 sedimentation rates. 210Pb_Liv3 is used to determine amount of offset between near-surface sediment record (Glew cores) and deep sediment record (Livingstone cores). MSH_Yn_14C_Ages contains tephra dates compiled and calculated for a weighted average of MSH Yn by Michael Clynne, USGS. Glass composition of four tephra samples by WSU’s Geochemical Lab is presented below.
Table 8: Tephra composition

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<th>sample 1</th>
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<th>sample 4</th>
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<td>77.5-78 cm</td>
<td>36-37 cm</td>
<td>6-7 cm</td>
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<td>12.19 (0.17)</td>
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<td>1.03 (0.04)</td>
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<td>3.79 (0.12)</td>
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<td>0.37 (0.03)</td>
<td>0.44 (0.04)</td>
<td>0.15 (0.02)</td>
</tr>
<tr>
<td>CaO</td>
<td>1.85 (0.12)</td>
<td>1.84 (0.07)</td>
<td>1.57 (0.09)</td>
<td>0.90 (0.07)</td>
</tr>
<tr>
<td>Cl</td>
<td>0.12 (0.03)</td>
<td>0.11 (0.03)</td>
<td>0.22 (0.06)</td>
<td>0.15 (0.01)</td>
</tr>
</tbody>
</table>

Total² : 100 100 100 100
Number of shards analyzed : 17 18 18 18

Probable Source/Age

- MSH Py 2500-2600 ¹³C BP
- MSH Yn 2900-3900 ¹³C BP
- Mazama Climactic 6845 + 50 ¹⁴C BP
- Glacier Peak Dusty Creek 5780-5830 cal BP

Similarity Coefficient³

<table>
<thead>
<tr>
<th></th>
<th>sample 1</th>
<th>sample 2</th>
<th>sample 3</th>
<th>sample 4</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.98</td>
<td>0.97</td>
<td>0.99</td>
<td>0.98</td>
</tr>
</tbody>
</table>

1. Standard deviations in parentheses
2. Analyses normalized to 100 weight percent
4. Foit et al., 2004
*Age_Depth.xlsx*

The age-depth modeling program *Bacon* was run with a variety of input dates to test the sensitivity of the input data. For each input file, only columns A-E were read by *Bacon*; the highlighted columns F and G are for user reference. *BaconInput_PresumedTephras* includes dates associated with six tephra layers, with varying levels of confidence in the tephra identification and subsequent relative stratigraphy. However, when *Bacon* was run with only the most robust tephra dates (*BaconInput_LimitedTephras*), the model incorrectly placed a TMF with a reservoir age inside the probability distribution, so we did not use this output. The upper co-ignimbrite ash fall deposit from Glacier Peak may be incorrectly placed in the core stratigraphy, and the same is true for Layer D from Glacier Peak. A sensitivity analysis was run for each; GP CAD was shifted 5 cm down (*BaconInput_offsetCAD*), and Layer D was shifted 2 cm down (*BaconInput_offsetD*). In both cases, ages shifted by at most 100 years. See *Output_*** worksheets for results from each model run. Columns A-E are direct *Bacon* outputs, whereas columns F-J were calculated by myself. To view or reproduce each *Bacon* run, see folder *Bacon_Input_Output*.

**Appendix 3: Paleoclimate Reconstruction**

*Climate data*

Climate data (Table 1) were necessary for estimating ELA climate conditions and as inputs for the glacier - climate model. If not already in a monthly format, parameters were aggregated from their source time-step to a monthly time-step for use in the energy balance component of the climate - glacier model. Most parameters came from modeled climate data sets based upon meteorological observations that were spatially extended to a uniform grid using physical processes such as lapse rates. The *Cloudiness* parameter contains the most uncertainty, as it was modeled on a global scale (Hahn, and Warren, 2007). Given the paucity of cloud observations available, this is unavoidable. In contrast, the PRISM Climate Group (2015) and Livneh et al. (2013) data sets were modeled on a local scale because they were able to interpolate measurements from thousands of meteorological stations throughout the region.

To assess the accuracy of the PRISM data, we compared data from the PRISM grid cell nearest to the Lyman SnoTel to the measured SnoTel data. Monthly averages of precipitation (cm) and daily mean temperature (°C) were plotted against each other. The $R^2$ correlation is very high in both cases (~0.99), although the PRISM grid cell is ~0.73°C warmer than the SnoTel site. We concluded that the PRISM data is an appropriate substitute for measured data in the Lyman Lake catchment basin.
Precipitation correlation of SnoTel (measured) vs PRISM (modelled)

\[ y = 1.039x - 0.0063 \]

\[ R^2 = 0.9895 \]

Temperature correlation of SnoTel (measured) vs PRISM (modelled)

\[ y = 0.9922x - 0.7329 \]

\[ R^2 = 0.998 \]
Lapse rates were calculated via four different methods; 1) Direct measurements at Lyman SnoTel (elev. 1823 m) and Thunder Basin SnoTel (elev. 1317 m); 2) PRISM data for Suiattle Pass DEM (~140 km²), comprising 219 data points; 3) PRISM data within the Suiattle Pass DEM but restricted to above 1800 m (108 data points); and 4) PRISM data confined within the Lyman Lake catchment basin (16 data points). The SnoTel-derived temperature lapse rate was clearly inaccurate at -0.07°C/100m, in comparison to standard temperature lapse rates of ~0.65°C/100m. The three PRISM-derived lapse rates varied by less than an order of magnitude and I was not able to discern which spatial group was the most accurate. However, the basin-limited lapse rate is the most precise spatial representation of the Lyman basin’s climate, hence its use in this study. A brief sensitivity analysis is presented (Table 8) to illustrate impacts of lapse rate on ELA climate.

Table 9: Lapse rate sensitivity analysis

<table>
<thead>
<tr>
<th>Lapse Rate Method</th>
<th>Lyman / Thunder Basin</th>
<th>PRISM, Suiattle Pass</th>
<th>PRISM, &gt;1800m</th>
<th>PRISM, Lyman Lake basin</th>
</tr>
</thead>
<tbody>
<tr>
<td>Summer T° lapse rate&lt;sup&gt;a&lt;/sup&gt;</td>
<td>-0.0007 °C/m</td>
<td>-0.0048 °C/m</td>
<td>-0.0045 °C/m</td>
<td>-0.0038 °C/m</td>
</tr>
<tr>
<td>Lapsed T° at 1995 m (LIA ELA)&lt;sup&gt;b&lt;/sup&gt;</td>
<td>7.50 °C</td>
<td>6.79 °C</td>
<td>6.84 °C</td>
<td>6.96 °C</td>
</tr>
<tr>
<td>Lapsed T° at 2300 m (modern ELA)</td>
<td>7.29 °C</td>
<td>5.33 °C</td>
<td>5.47 °C</td>
<td>5.81 °C</td>
</tr>
<tr>
<td>LIA – modern T° change</td>
<td>0.21 °C</td>
<td>1.46 °C&lt;sup&gt;c&lt;/sup&gt;</td>
<td>1.37 °C</td>
<td>1.15 °C&lt;sup&gt;c&lt;/sup&gt;</td>
</tr>
</tbody>
</table>

<sup>a</sup> Summer is defined by the ablation season at Lyman SnoTel, which is from May through September.

<sup>b</sup> Temperatures are lapsed from the average summer T° at Lyman SnoTel (7.61°C)

<sup>c</sup> Note that total variance of ELA T° from the PRISM methods is ~0.3°C

Climate_Data.xlsx

Lapse rates are a critical input for paleoclimate reconstructions by either the ELA reconstruction or climate - glacier modeling method. I experimented with four different lapse rate derivations (LapseRate_Summary) before settling on the most spatially confined data source, the PRISM grid cells within Lyman Lake basin (blue fill). Modeled T/P data from a PRISM-generated grid cell near the Lyman SnoTel was compared to SnoTel-measured T/P data for validation (PRISM_SnoTel_Validation). Precipitation values closely agreed at a monthly resolution (R² = 0.99) and exhibited no significant offset. Temperature values also agreed (R² = 0.99) but had significant offset, with the PRISM data point 0.73°C warmer. LymanBasin_ClimateData is a simple spreadsheet of the input parameters used for the energy balance model. Columns B – D are the precipitation lapse rate equation, and
columns J-L are the temperature lapse rate equation. To validate the energy balance results, I
compared measured data from the South Cascade Glacier with measured and modeled data at
the Lyman SnoTel (EnergyBalance_Validation). Note the graphs comparing temperature
and incoming shortwave radiation. Finally, Model Parameters contains the relevant constants
for the energy balance and ice-flow models if a user wanted to re-run the models.

**ELA reconstruction**

Although I employed the Area x Altitude Balance Ratio (AABR) method for my final
analysis, I also estimated paleo-ELAs via the Lateral Moraine (LM) and Accumulation Area
Ratio (AAR) methods (Table 9). Preserved lateral moraines were mapped to their highest
elevations on both sides of the valley, providing constraints on the paleo-ELA (Bowerman
and Clark, 2011; Meierding, 1982). Reconstruction of paleo-hypsometry followed standard
practices (3.5.1). Because the AAR method is so widely used, I estimated ELA at four
different AARs. In all cases, the paleo-glacier was bounded by the terminal and lateral
moraines and the cirque headwall was included as part of the accumulation area. ELA ranges
reflect enhanced sheltering on west side of valley, leading to an altitude-transgressive ELA.

**Table 10: ELA reconstruction methods**

<table>
<thead>
<tr>
<th>ELA Method</th>
<th>ELA (ft) range, E-W</th>
<th>ELA (ft), avg</th>
<th>ELA (m), avg</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>Maximum Elevation of Lateral Moraine (East-side)</td>
<td>NA</td>
<td>6410’</td>
<td>1954</td>
<td>No preserved lateral deposits above 6410’ on east-side</td>
</tr>
<tr>
<td>MELM (West-side)</td>
<td>NA</td>
<td>6550’</td>
<td>1996</td>
<td>Post-LIA deposition, therefore LIA ELA must be lower</td>
</tr>
<tr>
<td>Accumulation Area Ratio (0.65)</td>
<td>6600’-6400’</td>
<td>6500’</td>
<td>1981</td>
<td>Commonly used AAR (Bilderback, 2004; Marcott et al., 2009)</td>
</tr>
<tr>
<td>AAR (0.67)</td>
<td>6600’-6200’</td>
<td>6400’</td>
<td>1951</td>
<td></td>
</tr>
<tr>
<td>AAR (0.75)</td>
<td>6500’-6400’</td>
<td>6450’</td>
<td>1966</td>
<td></td>
</tr>
<tr>
<td>AAR (0.77)</td>
<td>6500’-6200’</td>
<td>6350’</td>
<td>1935</td>
<td></td>
</tr>
<tr>
<td>Area x Altitude Balance Ratio, ( BR = 5.66 )</td>
<td>NA</td>
<td>6545’</td>
<td>1995</td>
<td>Assumes equal mass balance gradients above and below ELA; BR derived from modern glacier</td>
</tr>
</tbody>
</table>
Glacier - climate modeling

The spatially-distributed, linked energy balance and ice-flow model (i.e., the glacier-climate model) created by Plummer and Phillips (2003) is a commonly-cited (47 citations as of April 2016) model that is under continual refinement. The iteration used herein resulted from a collaboration between the authors and Mitch Plummer of Idaho National Laboratory that recalibrated the model from the Tetons (where it was last used) to the unique glaciologic and climatologic conditions of the Lyman Glacier.

Recreating paleo-glacier reconstructions is a many-step process using the Matlab-based energy balance model and the Windows command line (CMD) ice-flow model, with Arc-GIS used to spatially transform and display cell data. Before the model can recreate paleo-glaciers, it must be tuned to the local meteorological and morphological conditions via an iterative process of calibration and validation. Although my final model configuration is unique to the Lyman Glacier, the process I followed should be applicable to any glacier.

The energy balance model requires climate parameters and spatial information as inputs. I searched for climate data sets that were as local as possible and covered a time period that overlapped with glacier extent observations (Table 9). A digital elevation model (DEM) was used, but first it had to be aggregated in ArcMap from 10 m resolution to 30 m resolution to facilitate model processing. The Matlab script requires a rectangular grid, so I clipped the DEM to a rectangle encompassing the Lyman Lake catchment basin. Upon assembly of inputs, I calibrated the model by attempting to match outputs to observed modern conditions. In theory, this meant positive Snow Water Equivalent (SWE) values should only be seen above the modern ELA. Given that the Lyman Glacier is actively retreating, I sought to limit the +SWE cells to the upper regions of the cirque headwall.

Model tuning was accomplished by altering the wind coefficient, the bulk transfer coefficient, and lapse rate derivation methods. The wind function facilitates ablation by increasing sensible heat flux across the landscape. It was found to almost entirely remove accumulated precipitation, so it was set to zero. I observed that differing lapse rates strongly effected the spatial distribution of SWE, leading to my selection of the most spatially precise data source. Once the spatial structure of SWE values looked reasonable, the magnitude of accumulation remained too large, so I reduced the bulk transfer coefficient by an order of magnitude to reduce accumulation. Thus, although I found one possible set of parameters that recreated modern accumulation conditions, there are certainly multiple solutions possible.

The energy balance model creates an elevation-based grid with SWE values for input into the ice-flow model. To do so, it calculates energy fluxes in every cell as well. These values can be used for model validation by comparison to measured energy fluxes, preferably at local sites with similar aspect and exposure. Although the Lyman SnoTel does not measure radiation, the nearby South Cascade Glacier (25 km to the northwest) has been instrumented
for such measurements for nearly two decades (USGS Weather Station 12181110), and was therefore selected for comparison (Figure 6).

After calibrating and validating the energy balance model output, I ran the ice-flow model to attempt to recreate modern glacial extents. However, because the ice-flow model operates on hydrologic flow principles, it must use a catchment-clipped extent. Therefore, I transferred the SWE grid to ArcGIS where I clipped it to the Lyman Lake catchment basin perimeter. In addition, the energy balance output SWE was in meters, whereas the ice-flow input assumes millimeters, necessitating a conversion. This issue manifested itself in the time steps necessary for the ice-flow model to run to equilibrium, which approached tens of thousands of years instead of hundreds.

Calibration of the ice-flow model was limited because the modern glacier is not at equilibrium, thus I did not expect the model to precisely reproduce it. Furthermore, the ice-flow model uses the modern glacier (embedded in the DEM) as its surface, meaning that it is simulating a glacier on top of a glacier instead of in a cirque. Initial results generated unreasonably thin ice, so I modified the deformation constant and the sliding constant (Table 2). The deformation constant had little effect on output, even after being altered by two orders of magnitude. Reduction of the sliding constant by one order of magnitude, however, resulted in appropriately thick ice.

Upon completion of glacier - climate model tuning, I altered climate in the energy balance model by iteratively reducing temperature. Each output was run through the ice-flow model, and the results were compared to the field evidence of the Little Ice Age maximum extent.

**Appendix 4: Tephra Discussion**

The Lyman basin’s proximity to Glacier Peak has resulted in frequent and voluminous tephra deposits within Lyman Lake. These deposits have similar characteristics to rock flour (high MS, high DBD, low LOI), often obscuring the glacier’s sedimentary signal. However, the repeated basin-wide tephra inundations also reveal much about basin sedimentation dynamics and the duration of tephra impacts on landscapes.

Grain-size analysis of the sediment record, with tephra layers included, reveals a consistent pattern of progressively finer-grained clastic sediments being deposited in Lyman Lake with time from tephra fall. For example, after deposition of the sandy MSH Yn tephra, the first pulse of slopewash is predominantly coarse silt, and then progresses to fine silt and eventually to clay after 17 cm of sedimentation. We interpret this pattern to reflect input of clastic sediments from a progressively greater catchment area. Because the coarser grained component of the tephra fall requires more energy to be mobilized, it is limited in its ability to be transported across the basin. In contrast, the finer grained component has a much lower threshold for transport, and therefore has the ability to traverse the catchment and reach the lake. Initial deposition, therefore, is dominated by proximal coarse grains which are quickly exhausted. The time lag between tephra fall and deposition of distally-sourced clay-sized
grains reflects transport time across the landscape, sometimes termed “residence time” (Ayris and Delmelle, 2012).

Table 11: Tephra residence time

<table>
<thead>
<tr>
<th>Tephra</th>
<th>Thicknessa (cm)</th>
<th>Tephra Grain-sizeb</th>
<th>Residence Time ( yrs)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mazama</td>
<td>4</td>
<td>Coarse silt</td>
<td>530 - 650</td>
</tr>
<tr>
<td>Dusty Creek</td>
<td>43</td>
<td>Fine silt</td>
<td>360 - 440</td>
</tr>
<tr>
<td>MSH Yn</td>
<td>3</td>
<td>Sand</td>
<td>310 – 430</td>
</tr>
<tr>
<td>MSH Py</td>
<td>1.5</td>
<td>Fine silt</td>
<td>15 - 80</td>
</tr>
</tbody>
</table>

a Estimate of airfall thickness recorded in core stratigraphy
b Dominant Grain-size at termination of airfall

Exceptions to the observed slopewash-fining pattern prove illustrative. Tephra fall from the Dusty Creek eruptive interval of Glacier Peak is mostly coarse-grained, but the final waning pulse is fine silt. The initial slopewash pulse is also fine silt, but quickly coarsens back to sand as fluvial and colluvial processes excavate the uppermost layer and mobilize the coarser tephra beneath. The pattern resumes, as sand gives way to silt and then clay over 12 cm.

It is important to note that the pattern observed, a fining of clastic sediments, does not directly indicate an increased sediment load to the lake. However, because this pattern dominates the Grain-size distribution signal, we infer that tephra slopewash is still significantly influencing the measured rock flour proxy values for the duration of the fining pattern.

The observed lag in lake deposition from tephra fall to the final clay-sized peak can be interpreted as the tephra’s residence time on the landscape. These data provide an interesting point of comparison with previously studied tephra residence times. Most studies have found residence times ranging from months to years (Ayris and Delmelle, 2012). In contrast, the tephras around Lyman Lake appear to have spent hundreds of years on the landscape (Table 11). Thickness of airfall is an obvious factor to be considered, but there is no correlation observed in the Lyman Lake record. However, the Lyman basin is a relatively barren alpine landscape dominated by snow and ice. There are few opportunities for the tephra to be sequestered into the biological landscape, in marked contrast to studies of vegetative communities around Mt. Saint Helens (Ayris and Delmelle, 2012). Even the basin’s snow accumulation, which can be effective in burying tephra, will eventually deliver the tephra to the lake as snowfall compresses to ice and flows into ablation areas. Thus, the glaciated nature of the region may artificially extend the duration of the residence time.

Tephra transport and residence time are not well-characterized, and would greatly benefit from additional study. The sedimentation record from Lyman Lake may prove helpful.