TREE-RING-BASED MASS BALANCE RECONSTRUCTION AT EASTON GLACIER, WASHINGTON, USA

by

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A thesis submitted to the Faculty of the University of Delaware in partial fulfillment of the requirements for the degree of Master of Science in Geography

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ABSTRACT

Tree-rings were used to reconstruct a 150-year mass balance record for Easton Glacier in the North Cascades Range of Washington State, USA. An annual ring-width chronology was developed from climate-sensitive Mountain Hemlock trees sampled at high-elevation stands on the south flank of Mount Baker. Dendroclimatic analyses revealed that tree growth was significantly correlated to spring temperature and seasonal snowpack, which provides the basis for the mass balance reconstruction. Linear regression analysis was applied to the ring-width chronology to reconstruct the mass balance history at Easton Glacier and to estimate regional mass balance variations for the period AD 1865-2014. The reconstructions show intervals of positive mass balance between AD 1865 to 1883 and AD 1956 to 1980, and intervals of negative mass balance between AD 1885 to 1895, 1930 to 1950, and the mid 1980s to 2010. The generalized periods of positive and negative mass balance correspond well with known periods of advance and retreat for Easton Glacier, as well as independent proxy records developed for the Pacific Northwest. Our results highlight the influence of the Pacific Decadal Oscillation (PDO) and provide the framework for studies of long-term climate and glacier variability in our study area.

Chapter 1

INTRODUCTION

Glacier fluctuations, primarily driven by changes in climatic conditions, are manifested as changes in mass, area, and length over annual to century-long timescales. Because relationships between glacier geometry and climate are often complex, direct measurements of mass input (accumulation) and mass loss (ablation) provide the best means of evaluating responses to high-frequency climate variability. However, not all glacier variations should be regarded as a direct result of climate change. A glacier's response to climate is also influenced by inherent physical characteristics, such as aspect, elevation, and hypsometry. Thus, no single glacier is representative of all regional climatic conditions and the behavior of many well-monitored glaciers is needed to determine a regional climate signal.

The North Cascades Range of Washington state is the most heavily glacierized area in the contiguous United States (Post et al., 1971). In this region, many glaciers are continually studied as high-profile indicators of regional and global climate change (Dyurgerov and Meier, 2000; Oerlemans, 2005). Much of the currently available literature focuses on the synchronous response of North Cascade glaciers to centennial-scale climate forcings, in terms of extensive retreat since the recent advances of the Little Ice Age, or short-term fluctuations influenced by annual-to-decadal climate variability (Long, 1955; Thomas et al., 2000; Kovanen, 2003; O'Neal, 2005).

The majority of research regarding North Cascades glacier variations relies on geochronologic and historical records that provide information about changes in glacier length and areal extent (Long, 1955; Heikkinen, 1984; Harper, 1993; O'Neal, 2005).

Typically, glacial geologic records only represent snap-shots of slow integrations of climate and have smoothed out the high-frequency climate variability that drives advances and retreats. Geomorphic records of margin positions are traditionally discontinuous, difficult to date, and suffer from the fact that an advance can obscure evidence of earlier margin positions (Luckman, 2000). A more complete record is necessary for analyzing glacier responses to climate perturbations at annual-to-decadal timescales.

When attempting to infer the climatic forcings responsible for past glacier advances or retreats, mass-balance records are invaluable. Annual mass balance is defined as the difference between total ice accumulation and total ice ablation on glacier during a given year. Because accumulation and ablation are functions of annual precipitation and melt-season temperature, mass-balance records reflect the direct response of glaciers to climate change without waiting for the slow integration. However, due to the extensive cost and labor associated with field measurements, mass balance records are limited to a small number of easily accessible glaciers. A handful of North Cascades glaciers have been monitored for the last few decades (Pelto and Riedel, 2001), but the only glacier with a substantial long-term record is South Cascade glacier, which has direct measurements collected since 1959. Because long records of more than a few decades of mass-balance data are exceedingly rare, we frequently look for proxy records of mass balance to supplement the direct measurements.

Series of tree-ring measurements have been used to construct proxy records of glacier mass balance in western North America extending back several hundred years (Larocque and Smith, 2005; Watson and Luckman, 2004; Wood and Smith, 2013; Malcomb and Wiles, 2013). Tree growth and glacier mass balance are expected to be inversely correlated because: (1) warmer spring/summer temperatures enhance tree growth and accelerate mass-balance loss; (2) colder temperatures and deep, persistent snowpack retard tree growth and increase glacier mass balance (Gedalof and Smith, 2001; Peterson and Peterson, 2001; Larocque and Smith, 2005; Malcomb and Wiles, 2013).

This study uses nine tree-ring series collected from high-elevation stands on the southern flank of Mount Baker, Washington, USA, to construct a 150 year-long proxy record of mass balance for Easton Glacier. The objectives of this analysis are to: (1) develop a standardized site chronology and verify dendroclimatic relationships, (2) derive the first mass-balance proxy record for Easton Glacier, (3) create a regionalized mass balance reconstruction using principal components analysis with other glacier mass balance records, and (4) discuss the mass balance record for Easton Glacier in the context of regional and global climate variability.

1.1 Study Area

The North Cascades mountain range is a smaller section of the larger Cascades Range, which extends from northern California to southern British Columbia, Canada. Located east of Washington State's Puget Sound, the northernmost section of the Cascades Range is characterized by rugged topography, glaciated peaks, and deep, U-shaped valleys. Elevations in this area range from <100 m to 3,300 m, with most of the peaks under 3,000 m in elevation. However, local relief is often over 1,800 m.

Mount Baker is the largest stratovolcano of the North Cascades Range. At 3,286 m, it is the third-highest mountain in Washington State and the fifth-highest in the Cascades Range. It is located roughly 90 km inland from the Pacific and approximately 20 km south of the Canadian border with the United States. A snowcap that encircles the summit provides part of the accumulation zone for 12 glaciers that extend to as low as 1320 m a.s.l., covering 38.6 km^2 (Pelto and Brown, 2012). Easton Glacier, the focus of interpretations presented in this study, is located on the south flank of Mount Baker at 48.75° N and 121.83° W (Figure 1). Positioned between 1680 m and 2900 m a.s.l., Easton is a large valley glacier that covers an area of 2.87 km^2 (Pelto and Brown, 2012). The glacier has created two prominent lateral moraines, Railroad Grade to the west and Metcalfe Moraine to the east. Terminus positions and mass balance measurements show Easton Glacier has been retreating since the end of its most recent advance in the late 1980s (O'Neal, 2005; Harper, 1993; Pelto and Brown, 2012).

Regional climate in the North Cascades Range is influenced by moist, cool air masses from the Pacific and the orographic effect of the mountains. Generally, winters are cool and wet, and summers are warm and dry, but seasons also vary greatly by location and elevation, with climate zones that range from temperate rainforests in the low valleys to alpine conditions on the high peaks. The Cascade Range experiences moderate temperatures year-round, with average January temperatures ranging from 2°C in the valleys to -27°C at high altitudes and average July temperatures ranging from 29° C in the valleys to 7°C at high altitudes. The region experiences exceptionally high precipitation, with annual totals ranging from 150 cm to 350 cm (Ruffner, 1980). Maximum precipitation occurs as snowfall in the winter, creating a seasonal snowpack that covers the ground until temperatures warm in the spring. High elevation and substantial winter precipitation contribute to Mount Baker's heavily glaciated peak. Mt. Baker itself experienced the most snowfall ever measured in the U.S. during a single season, when the winter of 1998-1999 brought a total snowfall of 28.96 m.

Climate variations at Mt. Baker and the larger Pacific Northwest region are heavily influenced by the Pacific Decadal Oscillation (PDO) and the El Niño/Southern Oscillation (ENSO). Both climate forcings produce variations in sea surface temperatures, sea level pressures, and atmospheric circulations over the upwind regions of the Pacific Ocean (Mantua and Hare, 2002). ENSO events are centered in the tropical Pacific Ocean and last between 6 to 18 months. In contrast, PDO events are centered in the North Pacific and occur on a much longer, 20- to 30-year cycle. PDO and ENSO events respectively explain up to 35% and 20% of Pacific Northwest climate variability (Bitz and Battisti, 1999). However, the forces and mechanisms driving PDO events are less well described and understood compared to those driving ENSO events.

Climate research classifies PDO events as either positive or negative, indicating whether sea surface temperatures are higher or lower than average, respectively, in the coastal Pacific Ocean immediately offshore from the Pacific Northwest (Mantua et al., 1997). In the Cascades, negative-phase PDO winters are typically wetter than normal, such as the climate pattern that governed the region between 1945 and 1976. Positive-phase PDO winters are typically drier than normal, such as those experienced from 1977 through the mid- 1990s (Pelto, 2008). Given the long timeframe of the PDO cycle, the current phase of the PDO (mid-1990s-present) cannot be determined (CIG).

Chapter 2

METHODS

2.1 Tree-Ring Chronology

Mountain Hemlock (*Tsuga mertensiana*) was chosen as the predictor species for this study because of its abundance at high elevations, its sensitivity to variations in temperature and snowpack, and its previously demonstrated dendroclimatic utility (Gedalof and Smith, 2001; Peterson and Peterson, 2001; Larocque and Smith, 2005)., Tree cores were collected from high-altitude stands of Mountain Hemlock near Easton Glacier on Mount Baker in 2014 (Figure 1). The climate/tree-growth relationship varies regionally where trees growing at the edge of their ecological range are more likely to record a strong climate signal (e.g., high elevations, rocky substrates, and steep slopes) (Speer, 2010). To maximize the potential climate signal in the study area, near-treeline stands were sampled to target trees at the edge of their ecological amplitude (Speer, 2010). Sampling sites were located along the ridge of the Railroad Grade moraine and adjacent meadow between 1628 m and 1695 m elevation (48.731° N, 121.842° W). Two cores were extracted at breast height from 10 trees using an increment borer.

Samples were glued to slotted mounting boards and sanded with progressively finer grades of sandpaper to enhance annual ring boundaries. Ring-widths were measured using a Velmex movable stage, optical linear encoder, and stereozoom microscope with a video capture system (see Figure 2). Measurements were recorded to the nearest 0.005 mm. Each core was measured by three independent workers in order to ensure accurate and consistent measurements.

Individual ring-width series were crossdated and standardized to create a master chronology. Crossdating was performed using standard dendrochronological techniques

(Stokes and Smiley, 1996) and verified with the program COFECHA (Holmes, 1983). The program ARSTAN was used for detrending and transforming raw measurements to standardized ring-indices (Cook and Holmes, 1996). A cubic smoothing spline with a 50 percent cutoff frequency at two-thirds of the series length was fit to each ring-width chronology. This detrending method has been used to reduce the impact of biological factors related to the life-cycle of the tree (e.g., age-related growth trends, stand disturbances, competition) while preserving low-frequency climate variability (Cook and Kairiukstis, 1990; Larocque and Smith, 2005; Wood and Smith, 2013). Ring indices were calculated by dividing the ring-width measurement by the spline curve value at each year. Individual series were averaged using a biweight robust mean to create a standardized site chronology.

Descriptive statistics were calculated to describe the quality of the site chronology, including series intercorrelation and mean sensitivity. The series intercorrelation measures the similarity among individual cores and is calculated as the mean correlation coefficient between each individual series and the site chronology. Series intercorrelation can range from -1.0 to +1.0, with higher values indicating a stronger common signal. The critical value of series intercorrelation for chronology reliability is 0.4 (NOAA Paloclimatology, 2008). Mean sensitivity is a measure of year-to-year ring-width variability ranging from 0 to 1. Smaller mean sensitivity corresponds to highly complacent rings, which show little difference from year to year. Larger mean sensitivity corresponds to highly sensitive rings, which alternate between very wide and extremely narrow rings. Generally, a mean sensitivity around 0.2 indicates that a series is sensitive enough for climate reconstruction, while less than 0.1 is too complacent and greater than 0.4 is too difficult to date (Speer, 2010).

2.2 Climate Data

To identify climatic parameters associated with tree growth and mass-balance, climate and snowpack records were obtained for long-term monitoring stations located closest to the study area. Monthly temperature and precipitation data were obtained from the Concrete PPL Fish Station for the period AD 1926 to 2014 (NOAA National Climatic Data Center). Means of consecutive monthly variables were used to construct seasonal and annual climatic variables. A record of seasonal snowpack was derived from monthly snow water equivalent (SWE) data collected from Snow Course/Aerial Marker sites at Marten Lake and Schreibers Meadow for the period AD 1959 to 2014 (USDA-NRCS). Due to infrequent monthly measurements, only April was used to represent seasonal snowpack because it had the most complete record and provided the best estimation of annual snow accumulation. Locations and descriptions of climate datasets are presented in Table 1 and Figure 3.

2.3 Glacier Mass-Balance

Reconstructions were calibrated to historical mass balance records obtained from the World Glacier Monitoring Service (Zemp et al., 2013). Mass balance data for Easton Glacier (Easton Bn) were collected annually by the North Cascades Glacier Climate Project for the period AD 1990 to 2012 (Pelto and Brown, 2012). Additional mass balance records were obtained for six North Cascades glaciers located within a 75 km radius of Easton Glacier (Table 2; Figure 3). Principal components were derived for the mass balance datasets using factor analysis. The first principal component was used to construct a regional mass balance index (Regional Bn).

The standardized site chronology was correlated to Easton Bn and Regional Bn to verify mass balance-tree growth relationships. Pending significant correlations, linear regression was used to generate mass balance models from annual ring-indices. Due to the limited length of historical records, entire mass balance data sets were used for model calibration. Goodness-of-fit of the models was evaluated using coefficients of correlation and determination, p-values, and standard errors of the estimate. The Durbin-Watson d statistic was examined to test for autocorrelation in the residuals.

To assess model validity, instrumental Bn data were compared to an independent dataset generated using the "leave-out-one" method, where observations are left out one at a time and calibrations are made on the remaining sub-samples (Watson and Luckman, 2004; Gordon, 1982). Verification statistics included Spearman's rank correlation (r_s) and the more rigorous reduction of error statistic (RE). RE provides a sensitive measure of model reliability and ranges from negative infinity to +1.0, which indicates a perfect estimation. Any positive value of RE suggests that the model has some skill and the reconstruction is of some value (Fritts, 1991). For additional verification, model results were compared to dated moraine sequences and independent mass-balance reconstructions.

Chapter 3

RESULTS

3.1 Chronology Descriptive Statistics

A 150-year site chronology was developed for the near-tree line stands of Mountain Hemlock (see Figure 4). Ages of individual samples ranged from approximately 80 to 400 years. A cut-off year of AD 1865 was determined for the master chronology, above which there was sufficient sample depth and reliable cross dating for time-series reconstruction. The series intercorrelation coefficient was 0.499 and the mean sensitivity was 0.225, which indicates a common stand-level signal and dendroclimatic utility. These values were comparable to those reported in other studies for Mountain Hemlock in this region (Peterson and Peterson, 2001; Gedalof and Smith, 2001).

3.2 Climate/Growth Correlations

Dendroclimatic relationships were examined by comparing our tree-ring chronology to historical climate and snowpack data. Product moment correlations (r) calculated between annual ring-width and seasonal temperature, precipitation, and SWE are presented in Table 3. The results of our analysis revealed that ring-growth is positively correlated with spring temperature (0.44) and negatively correlated with snowpack (-0.42) and precipitation (-0.37). Inverse relationships were observed for glacier mass balance, which is highly correlated to winter precipitation (0.71) and seasonal snowpack (0.67).

3.3 Mass-Balance Reconstructions

Our 150-year tree-ring chronology was used as the predictor variable to reconstruct Easton Bn and Regional Bn from AD 1865 to 2014. The models explain 33%

(Easton) and 29% (Regional) of the variance in the instrumental mass balance records (r^2 values, Table 4). These results are comparable to the variance explained by other net mass balance reconstructions in the Pacific Northwest (e.g., 50% explained by Wood and Smith, 2013 and 22% explained by Marcinkowski, 2012).

Analysis of calibration and verification statistics determined that the models of Easton Bn and Regional Bn are are statistically significant and have reconstruction utility (Table 4). Comparisons over calibration periods in Figure 5 reveal remarkable continuity between estimated and instrumental Bn data. The latter result is corroborated by Spearman rank correlations of 0.625 and 0.578 for Easton Bn and Regional Bn, respectively. Both models passed the sign test and the more rigorous RE test, indicating that the reconstructions have predictive skill. Results of the Durbin-Watson test revealed no statistical evidence of positive autocorrelation in the residuals.

The mean estimated Bn was negative for Easton Glacier (-236.7 mm w.e./yr) and marginally positive for the regional index (0.11 standardized units/yr) over the full 150-year reconstruction period. Intervals of positive/negative mass balance based on a 10-year running means are shown in Figure 6. Both reconstructions exhibit positive mass balance until the year AD 1883, which is followed by a decade of negative mass balance. For the period AD 1895 to 1925, Easton Bn was weakly negative and Regional Bn was weakly positive. Both models experience strong negative mass balance in the AD 1930s to 1940s, proceeding a period of positive mass balance in the AD 1950s to 1970s. The AD 1980s mark the beginning of a predominately negative mass balance period.

Chapter 4

DISCUSSION

Our strong correlations between annual ring-width and temperature, precipitation, and SWE agree with other studies, which suggest that tree-ring radial growth in Mountain Hemlocks is primarily controlled by the length of the growing season (Gedalof and Smith, 2001; Peterson and Peterson, 2001). For this high-elevation species, the start of the growing season is determined by seasonal snowpack depth and melt-out date (i.e., when soil temperatures are conducive to metabolic activity) (Peterson and Peterson, 2001; Gedalof and Smith, 2001). Warmer spring temperatures enhance melt-rates and initiate a longer growing season, but subsequently accelerate glacier ablation. Colder temperatures and a deep, persistent snowpack negatively impact radial growth, but positively influence glacier mass-balance. Therefore, the climate variables that control radial growth are nearly the same as those that influence glacier mass-balance, albeit with different coupling strengths. This serves as our rationale for using a ring-width chronology developed from high-elevation Mountain Hemlock as the predictor variable for a glacier mass-balance model.

Our mass-balance reconstructions also reflect the importance of precipitation on mass-balance variability of Easton and other maritime glaciers. Roe and O'Neal (2009) indicate that the mass balance of Mount Baker glaciers is 1.5 to 2 times more sensitive to precipitation variability than to temperature variability. This greater sensitivity of mass balance to precipitation is manifested in the advances observed during the negative PDO regimes (AD 1890-1924 and AD 1947-1976) (Figure 6).

Our mass balance reconstructions correspond well with independent mass-balance proxy-records developed for North America (e.g., Watson and Luckman, 2004 for Peyto

Glacier, Larocque and Smith, 2005 for the Mt. Waddington area, Wood et al., 2011 for Place Glacier, and Malcomb and Wiles, 2013 for Blue and South Cascade Glacier). For all of these records, common intervals of positive mass balance occur between AD 1860 and 1880 and between AD 1960 and 1980. Likewise, common intervals of negative mass balance occur between AD 1930 and 1950 and between ca. AD 1980 and 2010 or earlier for a few shorter records. Additionally, nearly all proxy records experience a sharp drop in mass balance near the end of the 19th century.

Several studies have presented past ice-margin positions for Easton Glacier based on ice-contact landform ages, personal accounts, and imagery (e.g., Long, 1955; Harper, 1993; O'Neal, 2005; Zemp et al., 2013). These studies suggest a history of Easton Glacier's terminal position as: (1) reaching its Little Ice Age (LIA) maximum in the mid-19th century, (2) retreating significantly at the beginning of the 20th century, (3) advancing in the AD 1960s through the 1980s, and (4) retreating from AD 1990 to the present. Although it is difficult to interpret high-frequency mass-balance records in terms of glacier termini changes, the 10 year running mean of our mass-balance (Figure 7) corresponds well with the known advances and retreats of Easton Glacier. The lag time of terminus advance/retreat with respect to mass balance changes, as observed in Figure 7, is consistent with the results of Harper (1992), who showed glacier/climate response times ranged from 13 to 17 years for Easton Glacier. Although historic glacial records support our model validity, the timing and magnitude of terminus response to climate forcing is an integration of the mass balance record translated via the dynamics of each individual glacier. The dynamics in turn depend on a multitude of factors (e.g., hypsometry, elevation, surface area) and are highly variable between glaciers.

It is also important to recognize that our model predictions will only be as good as the field observations of mass-balance used to calibrate our reconstruction. Errors in these mass balance measurements will be equally manifested in our results. Additionally, the models may be biased to the climatic conditions of the calibration period. Because mass balance measurements were collected during a primarily positive

PDO regime, mass balance anomalies related to the negative-phase PDO (i.e., positive Bn) may be weakened in the reconstruction.

Chapter 5
SUMMARY

We developed a 150 year tree-ring chronology near Easton Glacier that demonstrated strong stand-level coherence. Dendroclimatic utility was demonstrated by significant correlations among annual ring-width and mass balance with temperature, precipitation, and snowpack. Our mass balance reconstruction for Easton glacier corresponds well to independent mass balance reconstructions (Watson and Luckman, 2004; Larocque and Smith, 2005; Wood et al., 2011; Malcomb and Wiles, 2013). More importantly, an integration of our mass-balance reconstruction for Easton Glacier over time corresponds well to records of glacier terminus fluctuations. Furthermore, we found that multidecadal intervals of positive and negative mass balance tendency coincide closely with opposited phases of the PDO, fitting the expectation that negative PDO corresponds with positive mass balance in the Cascades, and vice versa.

Although our work was limited to a single location, our techniques can be used as proof-of-concept for future work to temporally and spatially extend the reconstruction. Future investigations could include developing multiple chronologies that are spatially and species-diverse, as well as, increasing chronology sample depth. While this is a reasonable goal, it should also be noted that tree ring correspondence with climate is always improved by being near the limits of a species' ecotone. A consequence of our being near treeline on the lateral moraine of an active glacier is that no other species but Mountain Hemlock was available in sufficient number to make a statistically reliable chronology. Very few climatic zones have glaciers below treeline, so glacier mass balance reconstructions based on tree rings will always rely on serendipitous availability of trees.

The decadal to century scale climate fluctuations in the Cascades most likely have multiple influences in the past century, including the PDO, greenhouse gas influences, and regional land use changes (O'Neal et al., 2010). In this study, based on the last 150 years, the clearest evidence is that PDO phases correspond to mass balance phases. Our study reinforces that winter precipitation variability has a greater influence on regional mass balance than summer temperature, which is opposite of the importance of temperature versus precipitation in other regions (Hanson, 1987). Because of the importance of precipitation changes to mass-balance variability in this region, greenhouse-gas-induced climate changes will not unambiguously cause glacier terminus retreat.

TABLES

Table 1: Summary of climate data

Site Name	Station ID	Lat/Lon	Elev	Period	Source[a]
Concrete PPL Fish	#USC00451679	48.54º, -121.74º	59.4 m	1926-2014	GHCN-Monthly
Schreibers Meadow	#21A10	48.70º, -121.82º	1036 m	1959-2014	USDA-NRCS
Marten Lake	#21A09	48.76º, -121.72º	1097 m	1959-2014	USDA-NRCS

[a] GHCN=Global Historical Climatology Network (NOAA); USDA=U.S. Department of Agriculture; NRCS=Natural Resources Conservation Service

Table 2: Glacier mass-balance datasets used to construct a regional mass balance index (Regional Bn) for the North Cascades

Glacier	**Location**	**Lat/Lon**	**Elev Range**	**Length**	**Source**[a]
Easton	Mt. Baker	48.75º, -121.83º	1680-2900m	1990-2012	NCGCP
Rainbow	Mt. Baker	48.80º, -121.77º	1310-2200m	1984-2012	NCGCP
Sholes	Mt. Baker	48.80º, -121.78º	1610-2110m	1990-2012	NCGCP
Lower Curtis	Mt. Shuksan	48.83º, -121.62º	1650-1950m	1984-2012	NCGCP
Yawning	Magic Mtn.	48.45º, -121.03º	1880-2010m	1984-2012	NCGCP
North Klawatti	Primus Peak	48.57º, -121.12º	1740-2400m	1993-2012	NPS
South Cascade	Sentinel Peak	48.37º, -121.05º	1640-2300m	1953-2012	USGS

[a] NCGCP=North Cascades Glacier Climate Project (Pelto and Riedel, 2001); NPS=National Park Service (Riedel and Larrabee, 2011); USGS=U.S. Geological Survey

Table 3: Correlations (r) between climatic variables with the annual ring-width chronology and mass-balance datasets used for model calibration. Annual variables represent one hydrologic year (October of previous year through September of current year). All values are statistically significant ($p < 0.05$).

Climate Variable	RW Chronology	Easton Bn	Regional Bn
Temperature			
Spring (Mar-May)	+0.44	-0.59	-0.59
Summer (Jun-Aug)	+0.21	-0.49	-0.60
Annual (Oct-Sep)	+0.41	-0.53	-0.60
Precipitation			
Winter (Nov-Mar)	-0.37	+0.70	+0.71
Annual (Oct-Sep)	-0.37	+0.63	+0.63
Snowpack[a] (SWE)			
SM (April)	-0.47	+0.56	+0.53
ML (April)	-0.42	+0.64	+0.67

[a] SM=Schreibers Meadow; ML=Marten Lake

Table 4: Calibration and verification statistics for glacier mass balance models

Calibration							Verification		
Model	Period	r	r^2	D-W[a]	p-val	Equation[b]	r_s[c]	Sign test[d]	RE[e]
Easton Bn	1990-2012	0.578	0.334	1.928	0.004	$y = -3272.0x + 2972.8$	0.625	17+5-	0.28
Regional Bn	1993-2012	0.540	0.292	1.365	0.014	$y = -2.955x + 3.012$	0.578	16+3-	0.22

[a]The Durbin-Watson *d* statistic
[b]y = estimated mass balance; x = standardized ring-width chronology
[c]Spearman's rank correlation coefficient ($p < 0.01$)
[d]Number of years in the calibration period where the mass-balance estimate correctly (+) or incorrectly (-) tracks the direction of change of the observations (signifcant at the 95% confidence level)
[e]Reduction of error statistic

FIGURES

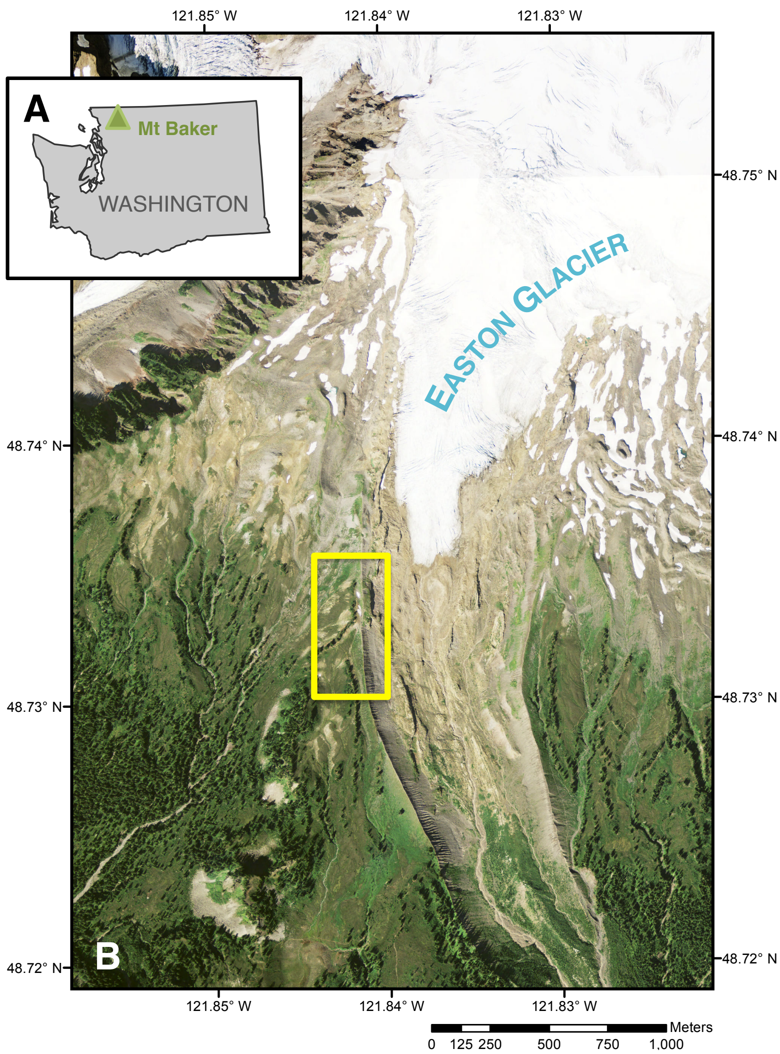

Figure 1: (A) Map of Washington State showing the location of Mount Baker. (B) NAIP 2013 imagery displaying Easton Glacier and the tree-ring sampling site (yellow box) located along the ridge and adjacent terrains of Railroad Grade moraine between ca. 1625 m to 1695 m a.s.l..

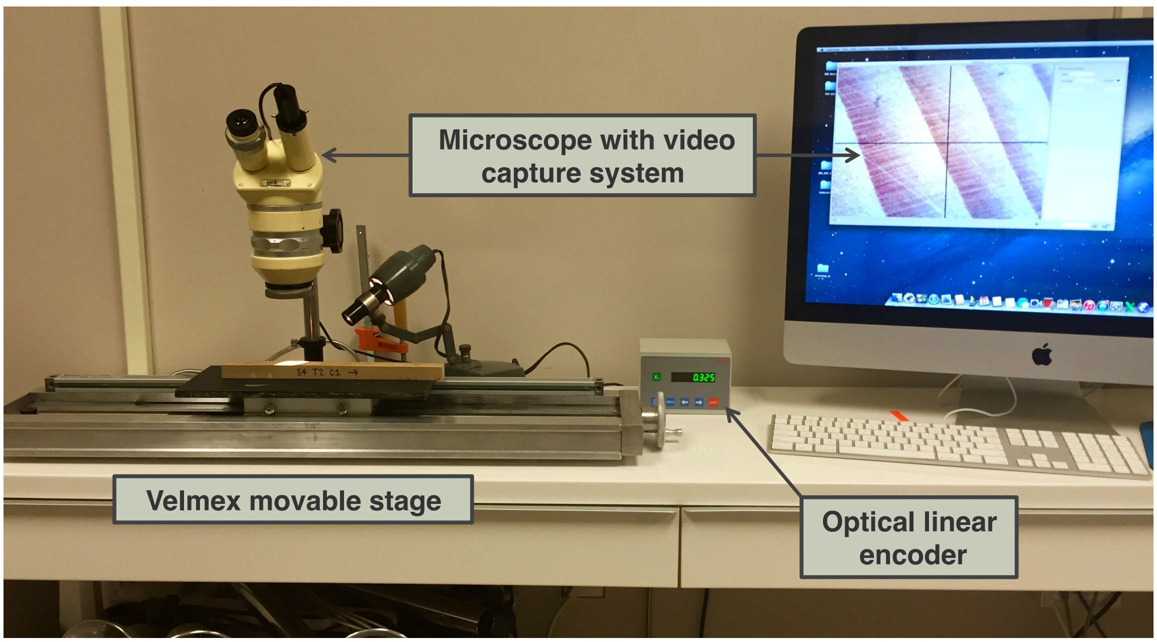

Figure 2: Ring-width measuring system

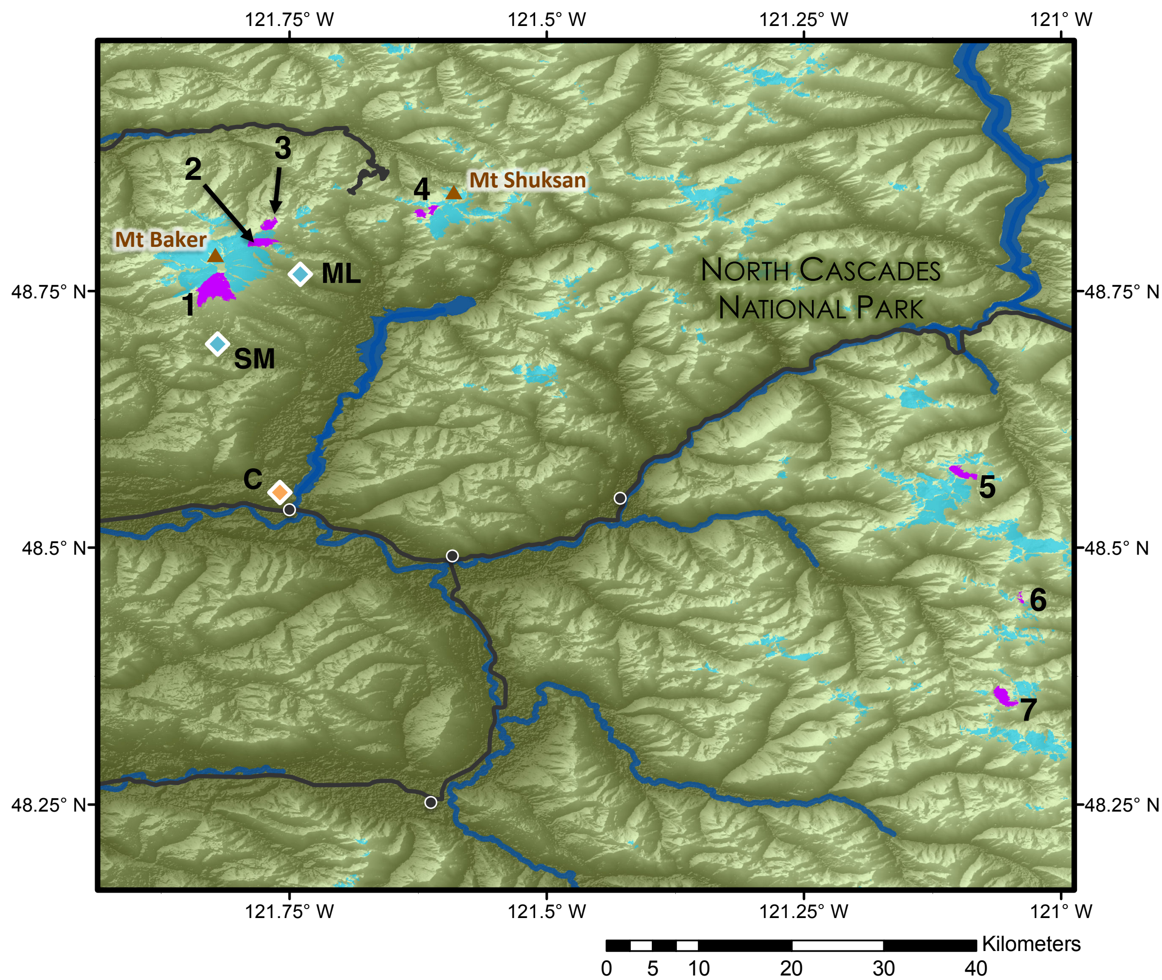

Figure 3: Map showing locations of glaciers used in the regional index calculation (glacier area shown in purple; 1= Easton, 2= Rainbow, 3= Sholes, 4= L. Curtis, 5= N. Klawatti, 6= Yawning, 7= S. Cascade), as well as the locations of long-term monitoring climate and snow stations (C= Concrete PPL Fish Station, SM= Schreibers Meadow, ML= Marten Lake). All other North Cascades glaciers are depicted by the light blue areas. ASTER Global DEM was used for the basemap (ASTER GDEM is a product of METI and NASA).

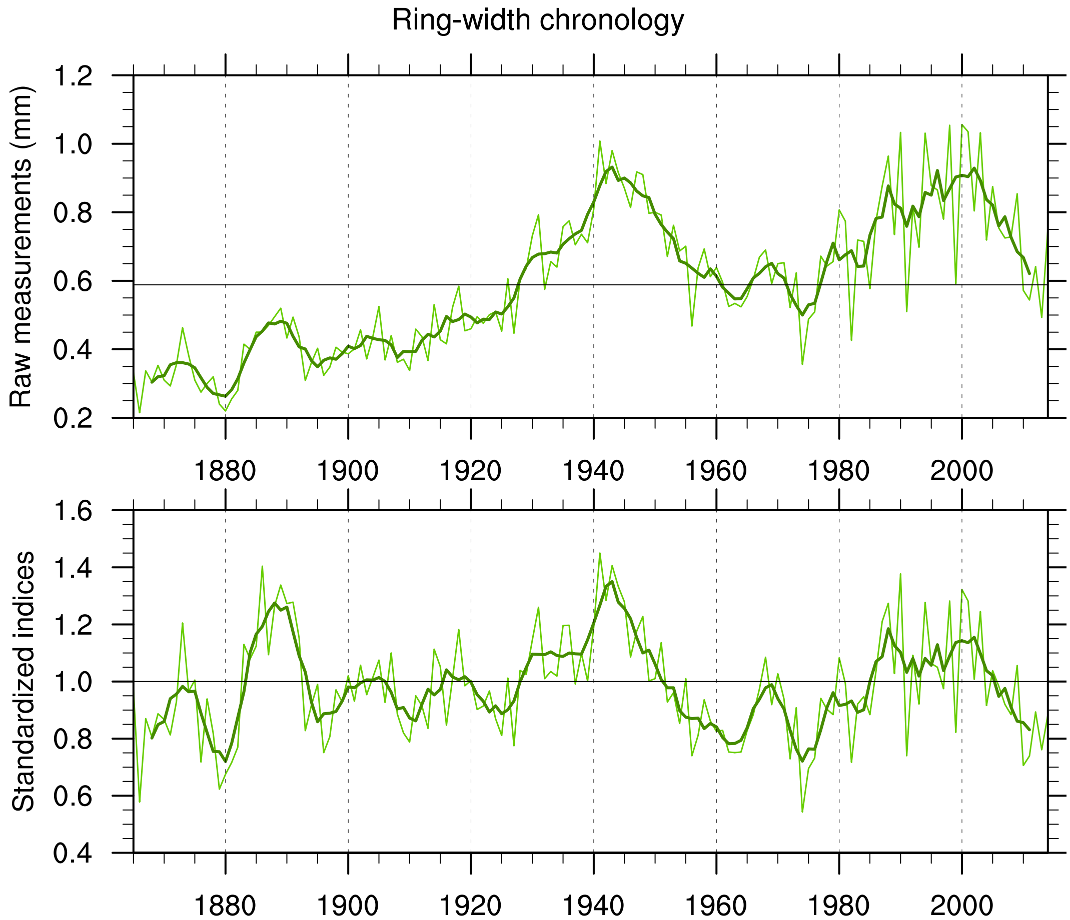

Figure 4: Time-series plots of the raw and standardized ring-width chronology developed from mountain hemlock trees near Easton Glacier on Mount Baker, Washington, USA (AD 1865 to 2014). Standardization was completed in ARSTAN using a cubic smoothing spline for detrending individual series. Thin lines represent annual values and thick lines represent 5-yr moving averages. The means for each series are delineated by a black horizontal line.

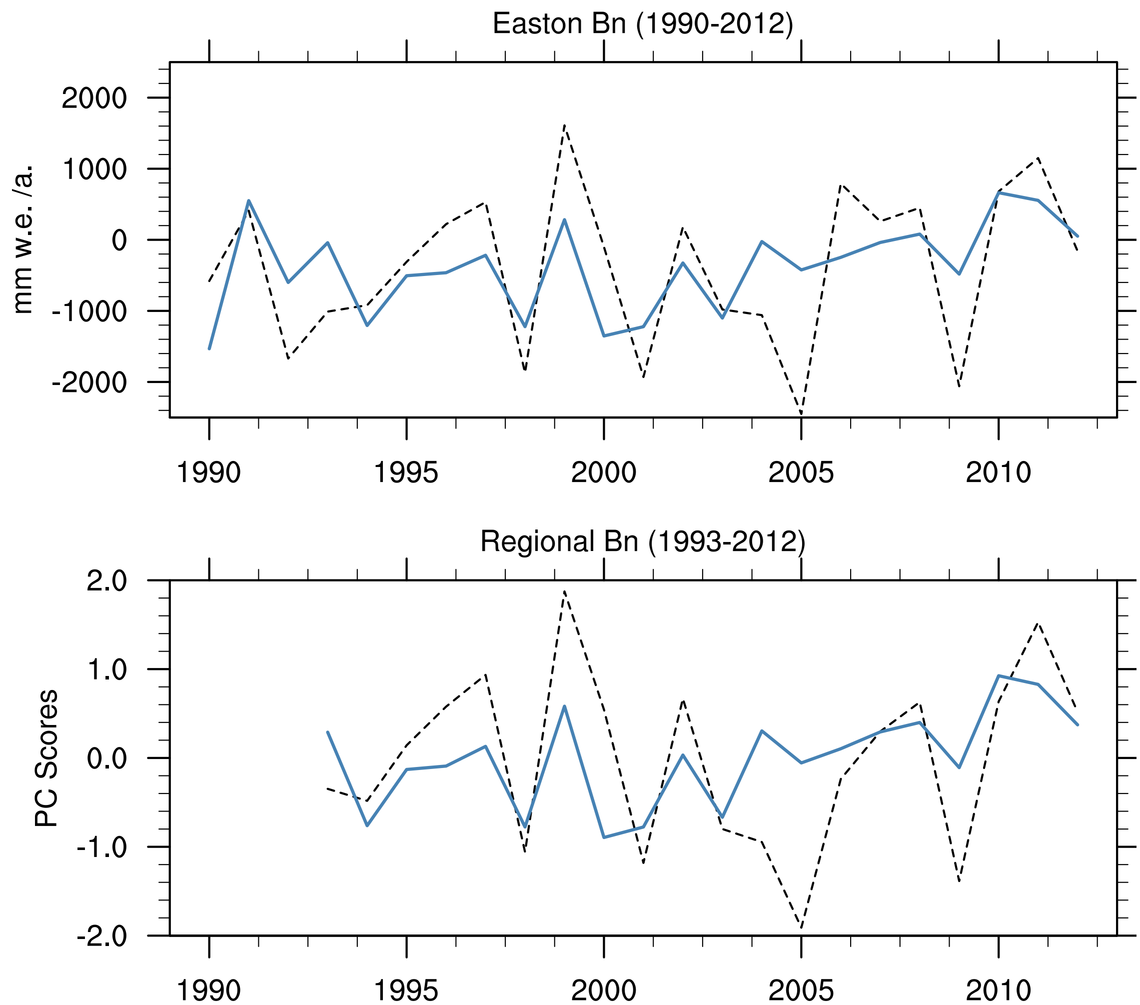

Figure 5: Observed vs. predicted mass balance over calibration periods for Easton Glacier and the regional index. Dashed lines represent observed values reported by the NCGCP and solid blue lines represent values predicted by our tree-ring-based models.

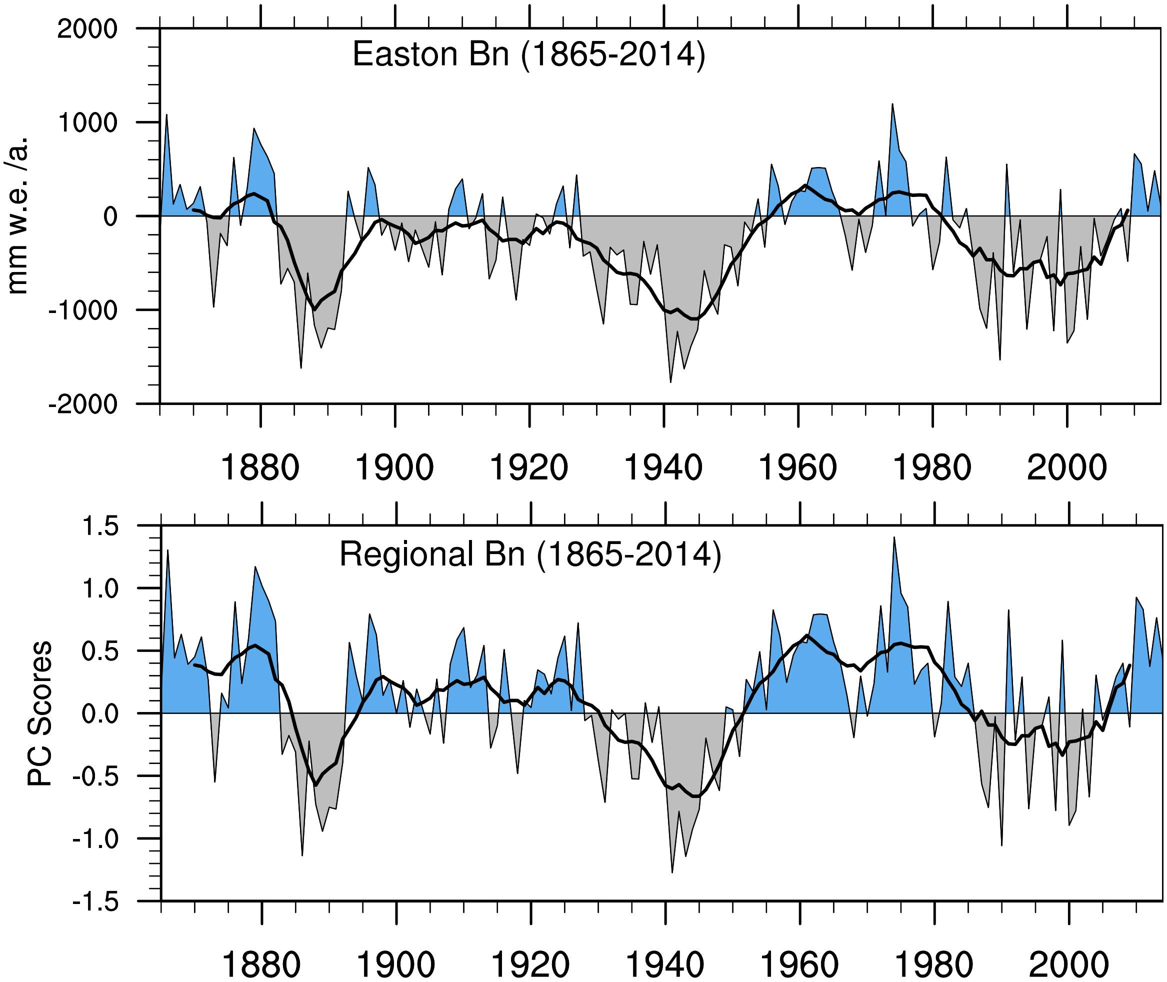

Figure 6: Net mass-balance reconstruction for Easton Glacier and the regional index. Blue shading corresponds to values greater than 0, above which mass balance is positive. Gray shading corresponds to values less than 0, below which mass balance is negative. Thick black lines represent 10-yr moving averages.

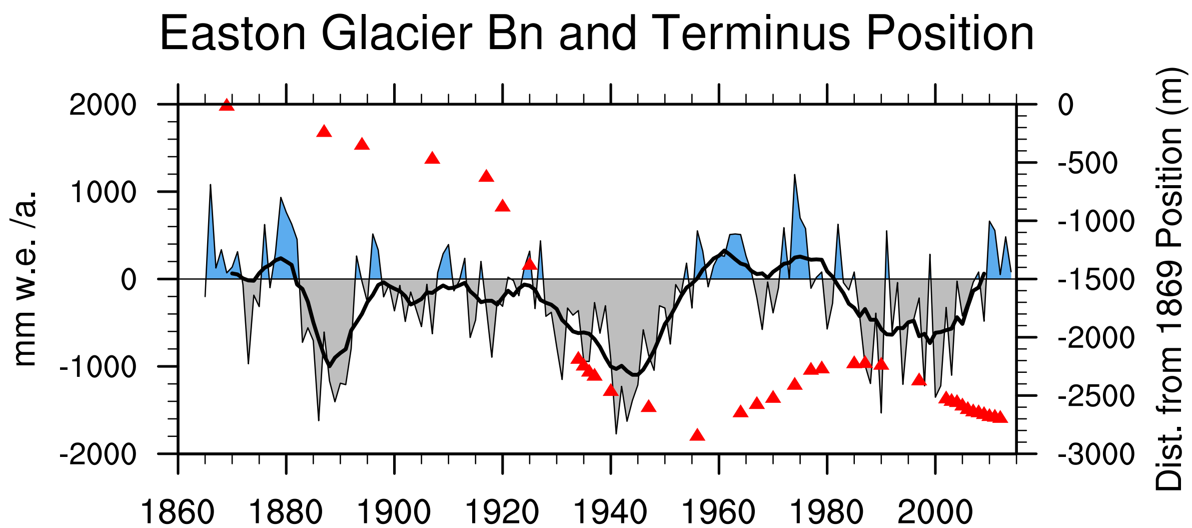

Figure 7: Comparison of mass-balance reconstruction to terminus records of Easton Glacier. Terminus positions between AD 1869 to 2012 are measured relative to the AD 1869 position (red triangles). Shading is same as described in Figure 6.

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