Author response to Anonymous Referee #1

We thank the referee for providing valuable feedback on our manuscript.

Specific comments The meaning of 'Glacier-wise' is unclear, but unfortunately used quite oftern throughout the paper. Please change to a more intuitive term.

Good suggestion, this term is unclear. We have removed all instances of 'glacier-wise'. See line 302 for an example of a revision:

***"Glacier averaged snow density from snow pits and cores for spring is 457 ± 48 kg m-3…"*.**

Use 'our glaciers' should be changed to something less possessive like, 'glaciers in this study'.

We agree with this suggestion and have changed instances of 'our study glaciers' and 'our glaciers' to more appropriate phrasing such as: 'glaciers in this study', 'studied glaciers', and 'these glaciers'.

279: Assuming that the exposed old firn occurs in the ablation zone, can you please provide an explanation as to how the overlying snow/firn/ice has ablated away without filling up the available pore space of the 'old firn' and leading to more internal accumulation than is accounted for in this study? this needs to be addressed as it also applies to your discussion on internal accumulation (L415-419) where it is similarly dismissed as insignificant.

This is a fair criticism. We now state in section 2.2.3:

"Firn meltwater retention and densification are neglected in our study."

We also added a discussion of meltwater retention renaming our section 4.1.2 Firn compaction, to 'Firn and internal processes'

"Internal accumulation within firn is not not incorporated into our annual balance estimates as it is not component of surface mass balance and is not measured within geodetic or glaciological surface mass balance studies. At Haig Glacier, firn meltwater retention has not been measured, but meltwater retention in the supraglacial snowpack is a negligible contributor to mass balance, though it does create an effective "energy sink", that should be accounted for in mass balance modeling (Samimi and Marshall, 2017). For glaciers in Svalbard, coupled energy balance and snow/hydrology models have been used to quantify the effects of meltwater freezing and retention on glacier mass balance (Van Pelt et al., 2012; Van Pelt and Kohler, 2015). Rates of meltwater retention are decreasing for Svalbard glaciers (Van Pelt and Kohler, 2015), and on the Devon Ice Cap (Bezeau et al., 2013), due to decreasing firn area and in particular, warmer temperature. Like at our glaciers, melt-freeze cycles form thick 'summer surface' layers on these Svalbard glaciers and Devon Ice Cap, which could act as a barrier for vertical water transport and is likely to promote near-surface lateral water flow, limiting deep firn water storage (Gascon et al., 2013; Van Pelt and Kohler, 2015)." L417-426.

Introduction doesn't justify this work well enough. Need to elaborate on the recent trends experienced by glaciers in western Canada as per menounous et al, 2018., and the potential impacts of declining contributions to stream flow post 2040ish as per Clarke et al, 2015. Contributions from glacier melt to sea level rise are of secondary importance from this region as it is poorly quantified as to how much actually makes it to tidewater and how much is taken up enroute through groundwater storage and human usage.

We agree with the reviewer that the importance of glacier mass change is more about water resources and much less about sea level rise. We revised the introduction to emphasize the importance of mass change on water resources.

L28-29 re: 'Measurement of seasonal mass change provides...' - I assume your talking about in situ mass balance measurements? if so, then should be specific about it - seasonal balances can be derived from more than just in situ meausrmeents –as you indicate below.

A fair point. Sentence is now modified to, *"Measurement of seasonal mass change via in situ and geodetic methods provides a means to assess the importance of meteorological drivers of glacier nourishment and melt". L36-38.*

L37-41: poorly written paragraph.

We refined this poorly written paragraph.

L50-51. "The climate of the Columbia Mountains is transitional between maritime and continental (Demarchi, 2011), with a strong maritime influence (Hägeli and Mc-Clung, 2003)." - So its more maritime then than continental? Would inner montane better describe the climate type here?

We have changed the sentence to: *"The climate of the Columbia Mountains is transitional between maritime and continental (Demarchi, 2011). L72.*

L55-56: Please give average snowfall rates and specify the source and what elevations they were measured at. This is probably the aspect of the climate that is the most important for this study.

An excellent suggestion. Please see the next comment response for average winter precipitation.

L64-65: please quantify differences - ie. average temp, snow precip, total precip, etc. describing the differences between climate regimes as "colder and drier.." is not very informative.

We now refer to climate normals from 1981-2010 for two nearby weather stations in the Columbia and Rocky Mountains. The text now states:

"From 1981-2010, Rodgers Pass, located in the center of the Columbia Mountains (Figure 1), at an elevation of 1330 m, has an average annual temperature of +1.9˚C, and an average winter (December-February) temperature of -8.0˚C, and experiences 1056 ± 49 mm w.e. of precipitation through the accumulation season (October-April) (Environment Canada, 2019)". L77-80.

"From 1981-2010, Lake Louise, located in the center of the southern Canadian Rockies (Figure 1), at an elevation of 1524 m, had an average annual temperature of +0.2˚C, an average winter temperature of -11.6˚C, and experienced 298 ± 9 mm w.e. of precipitation through the accumulation season. As

evidenced by comparing Lake Louise and Rodgers Pass, the Canadian Rockies are drier and colder in winter than the Columbia Mountains." L88-92.

While these two stations alone are not representative of their entire respective mountain ranges, they do quantitatively demonstrate the metrological differences between the two climate regimes.

L68-71: please link glaciers to the Columbia and Rocky mountain ranges (described above) more clearly. An outline or some indication of the extent of each of the major mountain ranges in Figure 1 would be useful

We have added the following sentence to clarify the glacier locations relative to the Columbia and Rocky Mountains:

"Haig Glacier is in the Rocky Mountains, whereas the other five glaciers lie in the Columbia Mountains". L99-100.

The major mountain ranges, the Columbia and Rocky Mountains are now labeled in Figure 1 and described in the figure caption:

"The Columbia and Rocky Mountains are separated by the Rocky Mountain Trench (RMT)".

L77: indicate swath widths for each instrument/altitude.

We added swath widths for each instrument:

"The VQ-580 and Q-780 were respectively flown at flying heights of around 500 and 2500 m above the terrain that yielded swath widths of 500-1000 m and 2000-3000 m". L105-106.

L81: is there a systematic bias in error of the laser shots as a function of off-nadir angle? Ie., does accuracy of z degrade towards the swath margins?

A good point raised by the referee. Yes, off-nadir laser shots do have larger positional errors than nadir ones. We designed flight lines to have 53% overlap, and this overlap would tend to reduce off-nadir bias. Any bias introduced by this sampling should be captured in the height uncertainty that we calculate for the stable terrain. We have expanded lines 106-108 to clarify this:

"We planned laser surveys with 53% overlap between flight lines, to yield return point densities that averaged 1-3 laser shots m $^{-2}$ ***(Table 2) with an effective sampling diameter of 10-20 cm per laser shot.". Point density has been added to Table 2 for all laser surveys.***

L119: It would be helpful to add a sentence or 2 here to describe what 'snow course' data is.

We have changed 'snow course' to 'snow survey' throughout the document, as this is the official name of the BC snow survey program. We have also added a reference (Weber and Litke, 2018) that details the methodology for the BC snow survey program. The data can be found at: https://catalogue.data.gov.bc.ca/dataset/705df46f-e9d6-4124-bc4a-66f54c07b228. We now introduce the snow surveys as 'manual snow survey measurements' and have added further description of these surveys:

"These snow surveys are conducted as part of the BC snow survey program eight times per year, with most sites located between 1000 and 2000 m asl". L147-149.

L282-283: 'Excluding this site, the remaining study glaciers in the Columbia Mountains had an AAR of 0.45 with 0.15 exposed multi-year firn cover and 0.40 bare glacier ice.' - The way this is written it implies Haig is in the Columbia mtns, it is not.

Sentence is now modified to:
"The study glaciers that lie in the Columbia Mountains had an AAR of 45% with 15% exposed multi-year firn cover and 40% bare glacier ice." L314-315.

L282: I presume you mean the average AAR of the remaining glaciers in the Columbia mtns? If so, please edit.

See the above comment for the clarified sentence.

Line 279-283: Line 279-280 indicates firn/glacier ice extents as percentages (13% and 49%) while the same are expressed as ratios on 281 – 282. Need to be consistent.

We have switched all ratios in the paragraph to percent for consistency.

L373: 'In western Canada, onset of snow melt is occurring earlier on average relative to 1970-2006'. Please clarify for what period the onset of earlier snow melt is occurring.

We have removed this sentence from the manuscript.

L387-389: the statement 'We also chose not to apply a firn correction since it requires glaciological measurements that we purposely withheld in order to evaluate the feasibility of measuring seasonal balance without surface observations from the glaciers.' Is vague. Please be more specific.

This statement has been removed from the manuscript. We initially chose to produce geodetic winter balance estimates only using the snow survey density to evaluate the feasibility of measuring seasonal balance without surface observations from the glaciers. However, we then used our in-situ densities to produce a separate geodetic winter balance estimate for each glacier to assess the impact of using in-situ versus regional density values (Table 3). The statement was attempting to convey that firnification models require an estimate of accumulation zone balance, which geodetic measurements, without correcting for ice dynamics, cannot provide.

L407-409: Re: 'Our field operations have been impacted by the melting out of crevasses: as strongly negative years are becoming the norm, and glacier flux is likely decreasing, crevasses are exposed for longer periods of time, and slower to close.
L408: ' Please define the 'melting out of crevasses'.
L408: What are you basing the assumption that flux is decreasing? Decreasing velocity or surface mass balance? Or both? If these assumptions are based on velocity changes, please indicate the sources used.
L409-411: re: 'This means that the total void area of crevasses is increasing due to ablation, which we have observed on Conrad, Zillmer, Nordic, and Haig glaciers, which could possibly increase their influence on Bw.' Can you expand on how this was observed? Was it measured? If so, how was it

measured and over what period of time?

The lines referenced in the above four referee comments have been removed from the manuscript. The authors feel that these lines add confusion and are a distraction. We have now added to the sentence leading into these lines which now reads:

"Despite the small influence of crevassing on Ba_geod observed in this study, additional studies should quantify the magnitude of this bias in greater detail".

What we intended to convey was that our visual field observations indicate that crevasses are being exposed (snow cover melted off) for a greater duration of the melt season than previously experienced. This extended exposure, tends to melt the sidewalls of the crevasses, widening the crevasses. After several years or decades of increased melt, many crevasses are merging to form ice-falls or serac fields that are difficult or impossible to navigate. This has implications for the safety and feasibility of travel during field work, but also for geodetic studies, as this likely increases the void area of crevasse fields, if not crevasse field extent. Ablation within crevasses is typically not captured by field studies, and may not be adequately captured in geodetic studies, depending on resolution and other factors.

As the length of the above explanation demonstrates, including these lines is a distraction from the goals of the manuscript, and while of scientific interest, our study has not taken steps to quantify these observations. Our primary goal was to highlight an area of uncertainty that future studies should tackle in greater detail, which the revised line above now does, without introducing a speculative discussion that we can add little to.

L415-419: Methods to measure internal accumulation include repeat shallow ice cores and ground penetrating radar (Bezeau et al., 2013; Gascon et al., 2013). As the issue of internal accumulation has not properly been addressed in western Canada, particularly over the larger icefields where this process has potential to be significant, it is worth highlighting as an important knowledge gap concerning glacier mass balance in this region.

We agree with this comment. We have now highlighted this important knowledge gap and added discussion of this process in section 4.1.2 as detailed in a previous comment. See lines 417–426 for the added material.

L426-427: re: 'Our glaciological measurement densities ranged from 0.5 to 18.5points km-2 (Table2), whereas our ALS data had around one million points km-2.' This is an unfair concluding statement as the datasets have different limitations that are not fully discussed.

We concur with the referee here and have removed this statement from the manuscript.

L433-434 specify, ' as the melt season progresses: : :' ice layers may form as internal storage 'within the snowpack'

Amended as suggested.

Technical corrections
Questionable use of hyphens throughout the paper.

Thank you for highlighting this issue. We have double-checked (sorry for the pun) our use of hyphens and corrected as requested.

L281: lower accumulation, no hyphen.

Hyphen removed.

L282: 0.06 add ' km2'

This was a ratio, and now is expressed as a percent.

Figures: text is of variable font and size – this should be standardized for all figures.
Text is so small it is unreadable on figures 7 and 4

We standardized our figure text font and size, so text in Figures 4 and 7 is legible.

References: Bezeau, P, Sharp M, Burgess D, and Gascon G (2013) Firn profile changes in response to extreme 21st century melting at Devon Ice Cap, Nunavut, Canada, J. Glaciol., 59(217), 981-991 (doi:10.3189/2013JoG12J208).

Gascon G, Sharp, M, Burgess D, Bezeau P, and Bush ABG (2013) Changes in accumulation-area firn stratigraphy and meltwater flow during a period of climate warming: Devon Ice Cap, Nunavut, Canada. J. Geophys. Res.-Earth, 118, 2380-2391

Thank you for these references, in addition to (Van Pelt et al., 2012; Van Pelt and Kohler, 2015), these were very informative to read and better discuss firn processes.

Author response to Anonymous Referee #2

We thank the referee for their valuable comments and time spent on evaluating our manuscript.

DTM-related uncertainties:

Which interpolation algorithms have been used for DTM production? Please state in one paragraph!

Laser survey point density has been added to Table 2 for all ALS surveys. We added the following lines:
"We post-processed point clouds and exported finished LAS files into LAStools (https://rapidlasso.com/lastools/) from which we used las2DEM to create 1 m resolution DEMs. Las2dem triangulates ground classified ALS points from las/laz files into a temporary triangulated irregular network (TIN). A DEM is then created from this using nearest neighbor interpolation. Given an average point density of greater than 2 points m^{-2} (Table 2), little interpolation was required." **L111-115.**

The authors use the mean, median and normalized median absolute deviation (NMAD) of the DTM-differencing over selected stable terrain as "systematic" uncertainty (Biash). It is unclear through the manuscript, how overall hdDEM was determined? Please describe in more detail! Furthermore, is hdDEM the same as Biash? If so, please streamline through the manuscript!

$\bar{\mathbf{h}}_{\mathbf{dDEM}}$ is the same as Bias$_{\Delta h}$ and should not have been included. We have removed all references to $\bar{\mathbf{h}}_{\mathbf{dDEM}}$. Bias$_{\Delta h}$ is *"…the mean height difference over stable terrain between two DEMs after coregistration".* L129.

Elevation change uncertainty ($\sigma h_{\Delta DEM}$) was calculated as a 2σ uncertainty using Supplemental eqns. 1 and 2: "*Elevation change uncertainty is derived from the σ of height change over stable terrain (σh) after correction for effective sample size (N_{eff})*". Lines 29-35 in S2.

Values of $\sigma h_{\Delta DEM}$ for both Ba and Bw can now be found in Table S3.

The authors have stated not to use their gridded data for correction of their sample sizes for spatial autocorrelation (L103-104). Please state what was used instead (point cloud data?) and, if so, have the point cloud datasets also been co-registered beforehand? To calculate uncertainty in ALS-derived height change, the authors account for spatial correlation, assessed over stable terrain based on semi-variogram analysis as described in Rolstad et al. (2009). How did the authors determine h and what are the values of hdDEM Please describe in more detail and present the values in a Table (also possible in Supplement).

Our wording was not clear. We use our gridded data to calculate effective sample size while accounting for spatial autocorrelation. The lines detailing this are in the supplemental information (S2), lines 28-37:

"Stable terrain generally covered 10-20 km2. We determined L by plotting semivariance (Figure S3) for randomly selected coordinate pairs (n=10,000) against distance for ten separate simulations and defined L as the distance at which semivariance becomes asymptotic (5% change threshold). Decorrelation length averaged 0.75 km and varied from 0.5 to 1.3 km".

We have also further clarified our treatment of spatial autocorrelation directly in the manuscript: "*To calculate uncertainty in ALS-derived height change, we also account for spatial correlation as assessed over stable terrain based on semivariogram analysis (Figure S3) as described in Rolstad et al. (2009)." L246-247.*

Values of $\sigma h_{\Delta DEM}$ for both Ba and Bw can now be found in Table S3.

Specific comments

L19: delete aiborne, redundant to ALS

Removed

L26ff: I recommend to align the references based on the date of appearance. Streamline through the manuscript.

The standard for The Cryosphere (Copernicus Publications) is alphabetical, as reflected throughout the manuscript.

L28: "'Measurement of seasonal mass change provides: : :" , you mean in situ measurements or do you refer to all methods?

This was also pointed out by referee #1. We refer to both methods as either produces results relevant to assessing meteorological drivers of glacier nourishment and melt. Revised:

"Measurement of seasonal mass change via in situ and geodetic methods provides a means to assess the importance of meteorological drivers of glacier nourishment and melt". L36-38.

L35: Please rephrase sentence!

Corrected.

L37-41: This paragraph could be improved, giving a bit more substance!

Similar to the concerns raised by Referee #1, we substantially revised the introduction of our paper.

L51: abbreviation CBR has to be introduced earlier

This acronym is now omitted since we only used it once in the paper.

L68-71: Please indicate the extent of the major mountain ranges in Figure 1.

The major mountain ranges, the Columbia and Rocky Mountains are now labeled in Figure 1 and described in the figure caption: *"The Columbia and Rocky mountains are separated by the Rocky Mountain Trench (RMT).*

L69: try to omit redundant information, which improves readability: eg. rephrase sentence to (1) Zillmer Glacier (5.4 km2) in the Cariboo Mountains, (2) Nordic Glacier (3.4 km2) and (3) Illecillewaet Glacier (7.7 km2) in the Selkirk Mountains, (4) Conrad Glacier (11.5 km2) and (5) Kokanee Glacier (1.8 km2) in the Purcell Mountains, and (6) Haig Glacier (2.6 km2), which straddles the continental divide in the Rocky Mountains.

Excellent suggestion. We have reorganized as suggested.

L77: swatch change to swath

Corrected.

L82: Please state in one paragraph, which interpolation algorithms you used!

Addressed in an earlier comment.

L117: It would be helpful to describe what 'snow course' data is.

We have changed 'snow course' to 'snow survey' throughout the document, as this is the official name of the BC snow survey program. We have also added a reference (Weber and Litke, 2018) that details the methodology for the BC snow survey program. The data can be found at: https://catalogue.data.gov.bc.ca/dataset/705df46f-e9d6-4124-bc4a-66f54c07b228. We now introduce the snow surveys as 'manual snow survey measurements' and have added further description of these surveys:

"These snow surveys are conducted as part of the BC snow survey program eight times per year, with most sites located between 1000 and 2000 m asl". L147-149.

L256: 'glacier-wise' is an unclear term, but used quite often throughout the manuscript. Is it possible to change to another more intuitive term?

Good suggestion, this term is unclear. We have removed all instances of 'glacier-wise'. See line 302 for an example of a revision:

"Glacier averaged snow density from snow pits and cores for spring is 457 ± 48 kg m-3…".

L281: omit hyphen in lower-accumulation

Removed.

L282: Which unit does 0.06 have? Think km2? I thought Haig glacier is not in the Columbia Mountains?

The value 0.06 was unitless as it is the ratio of firn area to total glacier area, but we now use 6% for clarity, and have switched all ratios in the paragraph to percent for consistency.

We have added the following sentence to clarify the glacier locations relative to the Columbia and Rocky Mountains*:*

"Located in the Rocky Mountains, Haig Glacier is the easternmost site in our study and is in a lower accumulation environment. It has lost nearly all its firn cover over the last 20 years, with firn area at 6% in 2015. The study glaciers that lie in the Columbia Mountains had an AAR of 45% with 15% exposed multi-year firn cover and 40% bare glacier ice." L312-315.

L331: hdDEM is not in Table 3, see comment above

$\bar{\mathbf{h}}_{\mathbf{dDEM}}$ is the same as $\mathrm{Bias}_{\Delta h}$ and has thus been removed from the manuscript. $\mathrm{Bias}_{\Delta h}$ is in Table 3, and we have added Table S3 which contains elevation change uncertainty (σhΔDEM), calculated as a 2σ uncertainty using Supplemental eqns. 1 and 2.

L373: 'In western Canada, onset of snow melt is occurring earlier on average relative to 1970-2006'. For what period is the onset of earlier snow melt occurring? Please give detrails.

We have removed this sentence from the manuscript.

L407-411: 'Our field operations have been impacted by the melting out of crevasses: as strongly negative years are becoming the norm, and glacier flux is likely decreasing, crevasses are exposed for longer periods of time, and slower to close. This means that the total void area of crevasses is increasing due to ablation, which we have observed on Conrad, Zillmer, Nordic, and Haig glaciers, which could possibly increase their influence on Bw.' This part is a bit unclear! What is meant with melting out of crevasses, please clarify. Which flux decreases (surface velocity or mass balance or both)? Give references of the source your assumption is based on! How was the increase of the void area of crevasses due to ablation observed? Can you detail this?

These lines have been removed from the manuscript. The authors feel that these lines add confusion and are a distraction. We have now added to the sentence leading into these lines which now reads:

"Despite the small influence of crevassing on Ba_geod observed in this study, additional studies should quantify the magnitude of this bias in greater detail". L457-458.

What we intended to convey was that our visual field observations indicate that crevasses are being exposed (snow cover melted off) for a greater duration of the melt season than previously experienced. This extended exposure, tends to melt the sidewalls of the crevasses, widening the crevasses. After several years or decades of increased melt, many crevasses are merging to form ice-falls or serac fields that are difficult or impossible to navigate. This has implications for the safety and feasibility of travel during field work, but also for geodetic studies, as this likely increases the void area of crevasse fields, if not crevasse field extent. Ablation within crevasses is typically not captured by field studies, and may not be adequately captured in geodetic studies, depending on resolution and other factors.

As the length of the above explanation demonstrates, including these lines is a distraction from the goals of the manuscript, and while of scientific interest, our study has not taken steps to quantify these observations. Our primary goal was to highlight an area of uncertainty that future studies should tackle in greater detail, which the revised line above now does, without introducing a speculative discussion that we can add little to.

L426: The statement “Our glaciological measurement densities ranged from 0.5 to 18.5 points km-2 (Table2), whereas our ALS data had around one million points km-2” is a bit of comparing pears with apples. Please discuss in more detail or omit!

Complete agreement here. This statement is unfair and has been removed. To discuss the relative strengths and weaknesses of each method, which this statement fails to do, is not the purpose of this section.

Figures: text is of variable font and size within figures. Especially on figures 4 and 7 the text is hardly readable

We have standardized our figure text font and size, and now text in Figures 4 and 7 is legible.

Interactive comment on The Cryosphere Discuss., https://doi.org/10.5194/tc-2019-30, 2019.

Multi-year Evaluation of Airborne Geodetic Surveys to Estimate Seasonal Mass Balance, Columbia and Rocky Mountains, Canada

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Abstract. Seasonal measurements of glacier mass balance provide insight into the relation between climate forcing and glacier change. To evaluate the feasibility of using remotely~~-~~ sensed methods to assess seasonal balance we completed tandem airborne laser scanning surveys (ALS) and field-based glaciological measurements over a four-year period for six alpine glaciers that lie in Columbia and Rocky Mountains, near the headwaters of the Columbia River, British Columbia, Canada. We calculated annual geodetic balance using ~~co-registered~~coregistered late~~-~~ summer digital elevation models (DEMs), and distributed estimates of density based on surface classification of ice, snow and firn surfaces. Winter balance was derived using ~~co-registered~~coregistered late~~-~~ summer and spring DEMs, and density measurements from regional snow ~~course~~survey observations and our glaciological measurements. Geodetic summer balance was calculated as the difference between winter and annual balance. Winter mass balance from our glaciological observations averaged 1.95 ± 0.09 m w.e., 4% greater than those derived from geodetic surveys. Average glaciological summer and annual balance were ~~also 4~~3% smaller and 3% greater respectively, than our geodetic estimates. We find that distributing snow, firn and ice density based on surface classification has a greater influence on geodetic annual mass change than the density values themselves. Our results demonstrate that accurate assessments of seasonal mass change can be produced using ~~airborne~~ ALS over a series of glaciers spanning several mountain ranges. Such agreement over multiple seasons, years, and glaciers demonstrates the ability of high-resolution geodetic methods to increase the number of glaciers where seasonal mass balance can be reliably measured.

1 Introduction

~~Glacier mass balance is a function of accumulation and ablation processes, responding directly to meteorological forcing at timescales of a season or more (Oerlemans et al., 1998). Mass balance observations help address research about changes of glacier runoff (Jost et al., 2012; Ragettli et al., 2016; Stahl and Moore, 2006), contributions of glacier mass loss to sea level rise (Bahr et al., 2009; Huss and Hock, 2015; Radić and Hock, 2011), and the response of glaciers to climate change (Clarke et al., 2015). Annual mass balance is the sum of accumulation and ablation throughout the balance year. Measurement of seasonal mass change provides one method to assess the importance of meteorological drivers of glacier nourishment and melt.~~

~~Unfortunately, seasonal balance is logistically and financially difficult, and these challenges lead to few seasonal mass balance records at the global scale (Ohmura, 2011). Currently, seasonal balance measurements for western Canadian glaciers are not publicly available (WGMS, 2017). Seasonal snowpack forms a critical component of glacier mass balance (Machguth et al., 2006). Knowledge of the high elevation snowpack and its change through time is limited or non-existent for most alpine regions (Barnett et al., 2005; Hamlet et al., 2005), including most glaciers within the Canadian Columbia River Basin (Brahney et al., 2017). A better understanding of the seasonal snowpack and its changes is needed to ascertain whether it is changing and whether and will offset or exacerbate increased ablation.~~

~~Several recent studies have used geodetic methods to measure seasonal snow depth on glaciers (Dadic et al., 2010; Helfricht et al., 2014; Machguth et al., 2006; McGrath et al., 2015; Sold et al., 2014) and to measure seasonal mass balance (Belart et al., 2017; Sold et al., 2013). Geodetic methods offer the ability to greatly expand the number of glaciers over which these measurements can occur (Berthier et al., 2014; Nolan et al., 2015). The primary objective of our study is to evaluate the reliability of geodetic surveys for estimation of seasonal mass change of temperate glaciers over multiple years.~~

Glaciers are in rapid retreat across western Canada (Menounos et al., 2018). Deglaciation is projected to have pronounced impacts on streamflow in western Canada (Clarke et al., 2015), with greatest reductions in August and September streamflow as glaciers shrink (Huss and Hock, 2018; Jost et al., 2012). In the Canadian Columbia River Basin, peak glacier runoff from ice wastage is either currently underway (Huss and Hock, 2018) or will occur within the next decade (Clarke et al., 2015). Improved projections of changes in glacier runoff will require refined treatment of seasonally-varying processes that nourish and deplete glaciers, namely the re-distribution of snow by wind and gravitational processes and changes in surface albedo. Seasonal mass balance records are also required to calibrate and validate these physically-based mass balance models. These records do not currently exist for the Columbia River Basin, however.

In addition to their use in refining estimates of future changes in glacier runoff, mass balance observations provide a valuable synopsis of a glacier's mass budget and its implications for glacier runoff (Jost et al., 2012; Ragettli et al., 2016; Stahl and Moore, 2006), water storage, regional climate (Huss et al., 2008; Radić and Hock, 2014), and contribution to sea level rise (Huss and Hock, 2018). Glacier mass balance is a function of accumulation and ablation processes, responding directly to meteorological forcing at timescales of a season or more (Oerlemans et al., 1998). Measurement of seasonal mass change via in situ and geodetic methods provides a means to assess the importance of meteorological drivers of glacier nourishment and melt. These observations can reveal trends and patterns in glacier mass evolution, and are valuable calibration and validation datasets for global (Huss and Hock, 2018; Maussion et al., 2019) and regional glacier models (Clarke et al., 2015), and for ingestion into regional hydrologic models (Schnorbus et al., 2014).

Seasonal balance is logistically and financially difficult and globally, few seasonal mass balance records exist (Ohmura, 2011). Currently, seasonal balance measurements for western Canadian glaciers are not publicly available (WGMS, 2018). Seasonal snowpack forms a critical component of glacier mass balance (Østrem and Brugman, 1991); it controls the volume and timing of runoff in the snowmelt-dominated tributaries to the Columbia River (Brahney et al., 2017). Like many regions (Barnett et al., 2005), high elevation snow and precipitation records are limited in the Columbia River Basin of Canada. Snow data are routinely only monitored at or below treeline, and much of the basin, including its glaciated terrain, exists above this elevation. Some models suggest snowpack may be increasing at high elevations (Schnorbus et al., 2014), though existing snow observations below treeline indicate decreased water equivalent through the 1980–2011 period (Brahney et al., 2017). This data gap hinders accurate estimates of alpine snowpack in the region, critical for glacier nourishment, ecosystems, hydropower, and flood forecasts (Hamlet et al., 2005).

Geodetic methods are now regularly used to measure seasonal snow depth on glaciers via surface (Helfricht et al., 2014; McGrath et al., 2015) or helicopter-borne ground penetrating radar (Dadic et al., 2010; Machguth et al., 2006; Sold et al., 2014), airborne laser scanning surveys-ALS (Helfricht et al., 2012, 2014; Sold et al., 2013), airborne photogrammetry (Nolan et al., 2015), and stereoscopic satellite imagery (Belart et al., 2017). Geodetic surveys offer the ability to greatly expand the number of glaciers over which snow depth and mass change measurements can occur (Berthier et al., 2014; Nolan et al., 2015). For hydrological applications, snow depth must be converted into snow water equivalent (SWE), and thus snow density must be known or estimated. Physical modeling of snow density is difficult (Sold et al., 2014), and in situ density measurements are sparse, and are expensive in terms of cost and effort. Often density measurements show little relation to either elevation or snow depth (Fausto et al., 2018; Machguth et al., 2006; McGrath et al., 2015), increasing the importance of in situ measurements. Density thus introduces uncertainty to geodetic winter SWE estimates which are vital to calibrate and validate hydrological modeling, and to measure seasonal mass balance (Belart et al., 2017; Sold et al., 2013). The primary objective of our study is to evaluate the reliability of geodetic surveys and density assumptions for estimation of seasonal mass change of temperate glaciers over multiple years.

1.1 Study Area

1.1.1 Columbia Mountains

The transboundary Columbia River Basin (668,000 km^2) spans seven U.S. states and British Columbia (BC), Canada. The Canadian portion of the Basin represents 15% of the watershed's total area, yet provides between 30~~-~~–40% of its total runoff, largely due to the presence of mountainous terrain with high amounts of orographic precipitation and extensive glacial cover (Cohen et al., 2000; Hamlet and Lettenmaier, 1999). There are 2,200 glaciers covering 1,760 km^2 in the Columbia Mountains (Bolch et al., 2010); these glaciers primarily exist within the Cariboo, Monashee, Selkirk, and Purcell ranges, with the highest elevations rising to over 3,000 m above sea level (asl).

The ~~climate of the~~ Columbia Mountains ~~is~~are transitional between maritime and continental ~~(Demarchi, 2011), with a strong maritime influence (Hägeli and McClung, 2003). Monthly average temperatures in the Canadian CRB~~(Demarchi, 2011). Monthly average temperatures in the Canadian Columbia River Basin (elevation range from 420 to 3700 m asl) range from -9.2˚C in January to 13.3˚C in July (Najafi et al., 2017; Schnorbus et al., 2014). General circulation is dominated by westerly flow, which brings consistent Pacific moisture, particularly in the winter months. Approximately 65% of annual precipitation falls as snow, with snowfall possible throughout the year (Schnorbus et al., 2014). The snow accumulation season in both the

Columbia and Canadian Rocky Mountains extends from October to May. The summer melt season runs from May through September. From 1981–2010, Rodgers Pass, located in the center of the Columbia Mountains (Figure 1), at an elevation of 1330 m asl, had an average annual temperature of +1.9˚C, and an average winter (December–February) temperature of -8.0˚C, and experienced 1056 ± 49 mm w.e. of precipitation through the accumulation season (October–April) (Environment Canada, 2019).

1.1.2 Rocky Mountains

The southern Canadian Rockies are located east from the Columbia Mountains (Figure 1) across the Rocky Mountain Trench and are home to 1090 glaciers covering 1350 km^2 (Bolch et al., 2010).

The eastern slopes of the Canadian Rocky Mountains experience a continental climate with mild summers and cold winters. However, winter precipitation along the continental divide is greatly influenced by moist Pacific air masses, with persistent westerly flow driving orographic uplift on the western flanks of the Rocky Mountains (Sinclair and Marshall, 2009). This combination of continental and maritime influences fosters extensive glaciation along the continental divide in the Canadian Rockies, with glaciers at elevations from 2200 to 3500 m asl on the eastern slopes. ~~The Canadian Rockies are drier, clearer, and colder in winter than the Columbia Mountains.~~From 1981–2010, Lake Louise, located in the center of the southern Canadian Rockies (Figure 1), at an elevation of 1524 m asl, had an average annual temperature of +0.2˚C, an average winter temperature of -11.6˚C, and experienced 298 ± 9 mm w.e. of precipitation through the accumulation season (Environment Canada, 2019). As evidenced by comparing Lake Louise and Rodgers Pass, the Canadian Rockies are drier and colder in winter than the Columbia Mountains.

2 Data and Methods

2.1 Study Sites

Over the period 2014~~-~~–2018 we measured seasonal mass balance of six alpine glaciers (~~Figure 1,~~ Table 1): (1) Zillmer Glacier (5.4 km^2) in the Cariboo Mountains, (2) Nordic Glacier (3.4 km^2) ~~in the Selkirk Mountains,~~and (3) Illecillewaet Glacier (7.7 km^2) in the Selkirk Mountains, (4) Conrad Glacier (11.5 km^2) and (5) Kokanee Glacier (1.8 km^2) in the Purcell Mountains, ~~(5~~and (6) Haig Glacier (2.6 km^2), which straddles the continental divide. Haig Glacier is in the Rocky Mountains, ~~and (6) Kokanee Glacier (1.8 km^2)~~whereas the other five glaciers lie in the ~~Purcell~~Columbia Mountains.

2.2 Geodetic Mass Balance

We performed repeat fixed-wing ALS surveys from late~~-~~ summer 2014 to late~~-~~ summer 2016 (Table 2) using a Riegl VQ-580 infrared (1024 micron) laser scanner with dedicated Applanix POS AV Global Navigation Satellite System (GNSS) Inertial Measurement Unit (IMU). Later surveys used the same GNSS IMU and a Riegl VQ-780 infrared (1024 micron) laser scanner.

The VQ-580 ~~was typically~~and Q-780 were respectively flown at ~~a height~~flying heights of around 500 and 2500 m above the terrain~~, whereas the VQ-780 was flown at 2600-2700 m above the terrain to enable wider swatch~~ that yielded swath widths ~~compared to the VQ-580.~~of 500-1000 m and 2000-3000 m. We planned the airborne surveys with 53% overlap between flight lines, to yield return point densities that averaged 1-~~2~~3 laser shots m^{-2} (Table 2) with an effective sampling diameter of 10~~-~~–20 cm per laser shot, and to minimize systematic bias from off-nadir laser shots.

2.2.1 ALS Post-Processing

Post-processing of the ALS survey flight trajectory data used the PosPac Mobile Mapping Suite (Applanix), with Trimble CenterPoint RTX with vertical and horizontal positional uncertainties that were typically better than ±15 cm (1σ). We ~~cleaned~~post-processed point clouds and exported finished LAS files into LAStools (https://rapidlasso.com/lastools/) ~~where~~from which we used las2DEM to create 1 m resolution DEMs. Las2dem triangulates ground classified ALS points from las/laz files into a temporary triangulated irregular network (TIN). A DEM is then created ~~1-m digital elevation models (DEM) from each survey.~~from this using nearest neighbor interpolation. Given an average point density of greater than 2 points m^{-2} (Table 2), little interpolation was required. We ~~co-registered~~coregistered all DEMs following the method detailed in Nuth and Kääb (2011). For late~~-~~ summer surveys, one master DEM was chosen and all other late~~-~~ summer DEMs were ~~co-registered~~coregistered to that DEM for stable surfaces only. Stable surfaces were identified in satellite imagery and excluded forests, lakes and ice- and snow-covered areas~~, which were all masked out.~~. For winter DEMs, the previous late~~-~~ summer DEM was used as the master DEM to mitigate against any surface height changes in areas defined as stable terrain, due to processes such as rockfall or vegetation height change. During the spring surveys, there was little to no snow-free terrain, except rocky features with extreme slopes which are not used in the ~~co-registration~~coregistration (slope >40˚ excluded). We thus did not apply any vertical shift during ~~co-registration~~coregistration of winter DEMs.

We utilized satellite imagery from Landsat 7 and 8, Sentinel-2, and Planet Scope at 30, 10, and 3~~ to ~~–5 m resolution respectively (Bevington et al., 2018), to guide surface classification used to ~~co-register~~coregister DEMs and calculate geodetic mass change. We selected the latest snow-free imagery from September or late~~-~~ August, and used a band~~-~~ ratio and threshold method (Kääb, 2005) to classify areas of snow, firn, and ice. In some cases, we manually corrected surface maps where our automated methods failed to differentiate between firn and snow surfaces.

To calculate annual mass change (B_a), we (1) difference two DEMs to create a height change DEM (ΔDEM), (2) ~~determine~~ bias correct the height change by the mean height difference over stable terrain between ~~the~~ two DEMs ~~in stable terrain~~ after coregistration ($Bias_{\Delta h}$, Table 3~~) to bias-correct the observed height change by any systematic elevation difference between the DEMs after co-registration,~~), (3) derive a mask based on surface classification of ice, firn and snow from satellite imagery (Figure S1), then (4) apply the density of each respective surface type (Table 4), to the ΔDEM to calculate mass balance.

~~For late-summer DEMs, stable terrain generally covered 10-20 km², providing enough area for bias (Figure S2) and uncertainty assessment. To calculate uncertainty in ALS-derived height change, we also account for spatial correlation as assessed over stable terrain based on semi-variogram analysis (Figure S3) as described in Rolstad et al. (2009).~~ We chose not to use Digital

Terrain Models (DTMs), which represent gridded elevation based on last returns from the laser scanner, since our gridding algorithms employed in LAStools filled crevasses and did not preserve sharp ridges that aided in ~~co-registration~~coregistration of the DEMs.

Annual glacier mass balance is defined as the sum of accumulation and ablation throughout the balance year (Cuffey and Paterson, 2010), which can be expressed as the sum of winter and summer balance:

$$B_a = B_w + B_s \quad (1)$$

For geodetic and glaciological mass balance, we measure winter and annual balance, and ~~use~~ calculate summer balance as the difference between them ~~as summer balance (Cuffey and Paterson, 2010):~~:

$$B_s = B_a - B_w \quad (2)$$

To calculate geodetic winter balance (B_{w_geod}), we created a ΔDEM from a given spring DEM and the previous late~~-~~ summer DEM~~, and the given spring DEM of interest, and~~ then applied spring snow density (Table 4). We did not independently estimate B_{s_geod} because of the added uncertainty of partitioning elevation change due to melt or compaction of snow/firn surfaces.

2.2.2 Density Estimates

While ALS provides an accurate estimate of snow depth with vertical uncertainties of ±0.1–0.3 m ~~(Abermann et al., 2010; Bollmann et al., 2011; Joerg et al., 2012)~~(Abermann et al., 2010; Bollmann et al., 2011; Joerg et al., 2012), it provides no information regarding snow density. We use manual snow ~~course~~survey measurements available from the British Columbia River Forecast Center (BCRFC) ~~(Najafi et al., 2017; Schnorbus et al., 2014)~~(Weber and Litke, 2018) as independent data to estimate spring snow density, and compare this with our measured glaciological snow densities. These snow surveys are conducted as part of the BC snow survey program eight times per year, with most sites located between 1000 and 2000 m asl. We use these BCRFC data to evaluate whether reliable estimates of snow density can be obtained for regions where no snow observations over glaciers exist. The mean date of our spring field visits was May 1st (Table 2), so we chose May 1st snow ~~course~~survey data (n = 10,169) to derive a relation between SWE (kg m^{-2}) and snow depth (m) (Figure 2). The linear relation (regression fit) yields a slope of 470 ± 70 kg m^{-3} (r^2 = 0.97), which we use as the average May 1st snow density which we applied for our geodetic B_w calculations. For Haig Glacier, we chose only snow ~~course~~survey measurements from the Rocky Mountains for a linear relation yielding 440 ± 50 kg m^{-3} (n = 629). The estimated uncertainty in bulk snow density (±70 and ±50 kg m^{-3}) represents the standard deviation (σ) of the snow ~~course~~survey data. For our glaciological density-informed B_{w_geod}, we use the observed glacier-wide snow density (Table S1) and a linear regression of density versus day and used the slope (3.0 kg m^{-3} day^{-1}, r^2 = 0.43) and days between the survey and the observations to adjust for change in snow density (Figure 3). The lack of an altitudinal trend in snow density observed on many glaciers (Fausto et al., 2018; McGrath et al., 2015, 2018; Sold et al., 2016) and those of this study, coupled with the absence of high-elevation snow density measurements and the annual variability of snow density evolution, required the use of a single value for spring snow density.

Regional observations of late~~-~~ summer snow density are consistent (Table 5); ranging from 530~~–~~–630 kg m^{-3} for glaciers across the Pacific Northwest (Table 5). This is expected for temperate, mid-latitude glaciers, where snow densities range from the

Field Code Changed

"critical density" of about 550 kg m^{-3} (Benson, 1962; Herron and Langway, 1980) to around 600 kg m^{-3} depending upon regional climatology. Since we independently evaluate glaciological ~~vs.~~versus geodetic estimates of mass change, we compare application of our late~~-~~ summer glaciological snow density measurements to calculate net balance with estimates based on the average of typical observations from four regional sources (590 ± 60 kg m^{-3}; Table 5), to test the impact of uncertainties of up to 10% in this parameter. Firn density has not been reported for the study area, so we estimate 700 ± 100 kg m^{-3} for multi-year firn based on observations in the Alps (Ambach et al., 1966). This is also consistent with our firn core measurements for firn two or more years old (Table S2; average density of 703 ± 65 kg m^{-3}, n=4). Measurements of one-year-old firn averaged 619 ± 47 kg m^{-3} (n=8). Given the sustained mass loss of Pacific Northwest glaciers (Bolch et al., 2010; Menounos et al., 2018; Pelto, 2006), exposed firn is generally more than one year old, and we apply an uncertainty of two times the σ of our multi-year firn core observations (±15%), which captures the range of observed firn densities (664-776 kg m^{-3}). We use an ice density of 910 ± 10 kg m^{-3} (Clarke et al., 2013). After performing a pixel-based surface classification for each late~~-~~ summer, we used these classification masks to assign a density (Table 4) to each pixel (snow/firn/ice).

2.2.3 Firn ~~Densification~~Processes

Firn meltwater retention and densification are neglected in our study. Firn densification (Belart et al., 2017; Sold et al., 2013) can be modeled, but this approach assumes that net annual surface elevation change corresponds to the average annual accumulation layer transformed from end-of-year snow density to ice (Sold et al., 2013). ~~Our glaciers~~Glaciers in this study have a low average accumulation ablation area ratio (AAR, 0.38~~,~~%, Table 3), and ice area ratios range from 0.38% to 0.94% (mean: 0.47~~).~~%). In most years, a significant amount of multi-year firn is exposed on ~~our study~~these glaciers, similar to other glaciers experiencing strong mass loss (Fischer, 2011; Klug et al., 2018). Firn area and~~ column~~ thickness losses interrupt the normal cycle of firn densification. Using the firn model of Sold et al. (2013) ~~gives~~yields an estimated annual surface lowering over a given accumulation area due to densification of ~0.20 m~~, but with high~~ ; yet uncertainty in estimating surface lowering resulting from densification is high since we lack knowledge of the required input parameters. Because of this, and because firn densification is unlikely to produce firn densities outside the range of our estimate (700 ± 100 kg m^{-3}), we chose not to estimate firn densification in our study. Firn compaction therefore comprises one systematic uncertainty term in our analysis.

2.3 Glaciological Mass Balance

We collected glacier mass balance measurements using the glaciological method (Cogley et al., 2011) with a two~~-~~ season stratigraphic approach (Østrem and Brugman, 1991). Spring glaciological field campaigns typically occurred between mid-April and mid-May, and the summer/annual balance visits took place between mid-August and mid-September (Table 2). ~~Integration of the point measurements~~Measurements of B_a and B_w ~~allows~~allow the calculation of summer balance B_s (Eqn. 1). Glacier mass balance measurements included snow depth, snow density, ablation, and kinematic GPS surveys of the glacier surface (Figure 4).

Our methods apply to the four glaciers studied by UNBC: the Zillmer, Nordic, Conrad and Kokanee glaciers. For Haig Glacier, winter mass balance measurements followed the same field protocols, but summer mass balance is derived from a combination of point observations and a distributed model of glacier melt (Marshall, 2014; Samimi and Marshall, 2017). The glacier melt model has 30 m-resolution and uses a surface energy balance, driven by AWS data collected on the upper glacier and in the glacier forefield. Illecillewaet Glacier has been monitored by Parks Canada since 2009 (Hirose and Marshall, 2013). We calculated B_{a_glac} for Illecillewaet Glacier using the contour method~~, as~~ since there were insufficient point measurements to ~~apply~~estimate mass balance using the profile method.

Others have shown that snow depth is more variable than density (Elder et al., 1991; Pelto, 1996; Pulwicki et al., 2018), so we designed a sampling strategy that measures snow depth much more than density (an approximate sampling ratio of 25:1). We used G3 industrial aluminum probes to collect over 1,750 estimates of snow depth over the period of study. The probe can penetrate thick ice lenses and ~~allows~~allowed us to measure snow depths of up to 8 m. The boundary between snow and firn is typically made up of clearly defined ice lenses of variable thickness, which can be detected with a probe on mid-latitude glaciers (Østrem and Brugman, 1991; Pelto, 1996; Sold et al., 2013). This end-of-summer surface at ~~our~~the glaciers in this study has such strength that an industrial probe can penetrate no more than a couple centimeters, in contrast with internal ice lenses in seasonal snowpack, which can be penetrated due to weak underlying support. Initially, we collected four probe measurements per location, but after two spring seasons we determined that two measurements were sufficient per location. The average σ for probe measurements for four (two) measurements was 0.14 m (0.07 m) for spring and 0.10 m (0.08 m) for late~~-~~ summer. Two measurements per location allowed additional locations to be measured, since our observed low variability between proximal measurements is consistent with other studies (Beedle et al., 2014; Pelto et al., 2013).

We measured snow density with a 100 cm^3 box cutter (Hydro-Tech) in snow pits and from snow cores using a 7.25 cm-diameter Kovacs corer. Our rationale to use a snow corer was that average spring snow depth exceeded 4 m and we chose to have as many sites as possible to estimate snow density. The corer also allowed us to sample internal ice lenses, which are difficult to measure with a snow sampler (Proksch et al., 2016). We measured spring snow density at low, middle and high elevations for each glacier. If we observed an elevation trend in our density measurements, we applied a linear regression of density and elevation to our depth measurements prior to converting these data to water equivalent (mass). When there was no linear gradient, we averaged the snow density measurements to produce a glacier-wide snow density.

We conducted nine side-by-side pit/core comparisons that revealed density measured in our snow pits was comparable, with density from snow pits about 0.2 ± 5.7% heavier than measured by subsampling snow cores (Figure S4). The mean absolute difference between pit and core density was 4.8%, similar to observations made at Alto dell'Ortles (Gabrielli et al., 2010). Methodological differences (~~Supplement 1~~S1) are within the range expected between duplicate field-based measurements of snow density (1-6%) and with different cutters (3-12%) (Conger and McClung, 2009; Proksch et al., 2016).

Aluminum and PVC ablation stakes were used on each glacier to measure ice and firn ablation. The stake heights were measured (±1 cm) and ~~re-drilled~~redrilled during each late~~-~~ summer visit. As a check on stake elevation, we measured depth to the previous snow surface for all stakes in firn, as stakes may self-drill in firn (Østrem and Brugman, 1991). Stakes were generally aligned along the centerline of a given glacier; however, we added a second transect of stakes to cover each branch to improve spatial coverage on each study site (Figure 4). Conrad Glacier also featured three latitudinal sets of ablation stakes. To calculate mass balance, we used the profile method (Escher-Vetter et al., 2009), applied over 100~~-~~ m hypsometric elevation bins. The area-altitude distribution of a given glacier was obtained using our annual late~~-~~ summer ALS DEMs. The boundary of each glacier was manually delineated using the ALS DEM hillshade of the previous late~~-~~ summer, and a ΔDEM (Abermann et al., 2010). We also calculated mass balance using linear regression. For Zillmer, Nordic and Conrad glaciers, we separately considered the measurements from two distinct branches or sides of each glacier and then separately applied the profile and linear methods to each branch.

To account for mass change between a given field visit and the associated ALS survey, we completed kinematic GPS surveys using a Topcon GB-1000 receiver as a rover and a second receiver as a base station. We corrected base~~-~~ station data using Natural Resources Canada Precise Point Positioning (https://webapp.geod.nrcan.gc.ca/geod/tools-outils/ppp.php~~),~~) before post-processed surveys using Topcon Tools. Height change observed between the ALS DEM surface and survey points were binned by elevation (Figure S5~~),~~) and assigned a density based upon surface classification as determined from satellite imagery. Since ALS surveys were essentially synchronous (typically flown over two to three~~-~~ days), we chose to apply the correction to the glaciological estimates of mass balance. We surveyed 2~~-~~–6 control points at each site to determine the survey reliability and found that horizontal and vertical uncertainties respectively averaged ±0.04 m and ±0.06 m.

2.4 Uncertainty Assessment

We analyzed ~~snow and ice-free~~stable terrain to derive statistical indicators of bias and data dispersion from ΔDEM ~~over stable terrain~~ using a late~~-~~ summer DEM as a reference~~, and~~. We report the mean, median and normalized median absolute deviation (NMAD) over stable terrain (Table 3), which generally covered 10~~-~~–20 km². To calculate uncertainty in ALS-derived height change, we also account for spatial correlation as assessed over stable terrain based on semivariogram analysis (Figure S3) as described in Rolstad et al. (2009). We bias~~-corrected~~ correct the height change over the glacier surfaces using the systematic elevation difference over stable terrain (~~h_{dDEM})~~$Bias_{\Delta h}$) in the ΔDEMs~~.~~ (Table S3). This bias correction ranged from -0.09~~ to~~ –0.05 m and averaged -0.01 m. NMAD reveals random errors that are typically below ±0.3 m, with a maximum of 0.6 m (Table 3). This maximum error occurred for Zillmer Glacier in late~~-~~ summer 2017 when the separation between site visit and ALS survey was large and new snow covered the glacier during the ALS survey (Table 2).

Random uncertainty stems from three sources that we assume to be independent: i) elevation change uncertainty ($\sigma h_{\Delta DEM}$), ii) glacier zone delineation uncertainty (σA), and iii) volume to mass density conversion uncertainty ($\sigma \rho$). We define elevation change uncertainty ($\sigma h_{\Delta DEM}$) following the methods detailed in (Menounos et al., 2018), and found an average decorrelation length of 0.75 km (Figure S3). Below, we have abbreviated our geodetic and glaciological uncertainty assessment (detailed version: ~~Supp. 2~~S2)

For delineation of ice/firn/snow zones from satellite imagery (Figure S1), we applied a buffering method (Granshaw and Fountain, 2006) to the perimeter of each zone that was not at the glacier boundary. Our satellite imagery resolution varied from

3 to 15 m, so we chose a buffer of four times the largest pixel size~~,~~ to derive an uncertainty in area per zone. This 60 m buffer accounts for uncertainty in zone delineation and changes in the positions of the zone boundaries occurring between ALS and satellite imagery acquisition dates. Total random uncertainty in volume change is:

$$\sigma\Delta V = \sqrt{(\sigma h_{\Delta DEM}(p + 5(1-p))A)^2 + \ (\sigma A \ \cdot \ h_{\Delta DEM})^2} \tag{3}$$

where A is the area of a given glacier and p is the percentage of surveyed area, which averaged 99.1% (Table 2). Random uncertainty on geodetic mass balance is:

$$\sigma\Delta M = \sum_i \sqrt{(\sigma\Delta V_i \ \cdot \rho_i)^2 \ + \ (\sigma\rho_i \cdot \Delta V)^2} \ \cdot \frac{A_i}{A_{tot}} \tag{4}$$

where ρ_i is individual density conversion values with associated uncertainties ($\pm\sigma\rho_i$) for spring snow, late~~-~~ summer snow, firn, and ice (Table 4). Prior to being summed to produce a final uncertainty, each zone (ice/firn/snow) is considered separately for B_a, with ΔV_i and A_i the volume and area change of each zone respectively.
Firn compaction or fresh snow on the surveyed surface introduce systematic uncertainty on geodetic balance. On Drangajökull ice cap, where B_w is more than 1 m w.e. greater than our average B_w, firn compaction and fresh snow densification increased geodetic B_w by 8%. Fresh snow off-glacier was negligible in all but a few cases. We thus assume a systematic uncertainty ($\sigma\Delta M_{sys}$) of 10% on $B_{a,w}$. Collectively, random and systematic uncertainty thus yield total uncertainty in mass balance:

$$\sigma B_{geod} = \sqrt{(\sigma\Delta M)^2 + \ (\sigma\Delta M_{sys})^2} \tag{5}$$

To determine uncertainty in glaciological mass balance, we derive a mean density (ρ) of mass change (Table 3) and uncertainty in height change for both observations and GPS survey corrections. Uncertainty in glaciological mass balance is calculated as:

$$\sigma B_{a,w} = \sqrt{\sigma\Delta h_{glac}{}^2 \cdot \rho^2 \ + \ \sigma\rho^2 \cdot B_{a,w}{}^2} \tag{6}$$

where $\sigma\rho$ is the uncertainty on density taken to be 10% of ρ, to account for uncertainty in density measurements and extrapolation of those measurements. The uncertainty in extrapolation of glaciological observations to glacier-wide mass balance is taken as the σ of the different calculations of mass balance for each season.
For both geodetic and glaciological mass balance, B_s was derived as the difference of annual and summer balance (Eqn. 1), and thus uncertainty on B_s yields:

$$\sigma B_s = \sqrt{\sigma B_a{}^2 + \ \sigma B_w{}^2} \tag{7}$$

3 Results

3.1 Glaciological Versus Geodetic Balance

Comparison of seasonal balance from glaciological and geodetic methods showed strong overall agreement (Figure 5), with glaciological winter balance (B_{w_glac}) averaging 1.95 ± 0.08 m w.e., 4% greater than our geodetic estimate. Average B_{s_glac} and B_{a_glac} were ~~4~~3% both more negative than B_{s_geod} and B_{a_geod} (Figure 6). ~~Average glacier-wise~~For individual glaciers, average difference between B_{a_glac} and B_{a_geod} was in excellent agreement (-0.03 m w.e. relative to B_{a_glac}), with an average absolute deviation of 0.10 ± 0.07 m w.e. a^{-1} between estimates (Figure 6). ~~Glacier-wise,~~ B_{w_glac} was 5% greater relative to B_{w_geod}, and B_{s_glac} was 4% more positive relative to B_{s_geod}~~.~~ when considering individual glaciers. For B_w and B_s, geodetic and glaciological balance were within 20% for over 85% of cases. Average ~~glacier-wise~~ mean annual balance from 2015~~-~~–2017 was -0.~~70~~73 ± 0.15 m w.e. and -0.76 ± 0.16 m w.e. for ~~both geodetic and~~ glaciological ~~mass balance separately~~and geodetic methods respectively (Table 3). Mean B_{s_glac} was -2.~~70~~67 ± 0.13 m w.e. All ~~glacier-wise~~individual estimates of seasonal and annual balance are within 2σ uncertainties, and only in three instances are they outside 1σ uncertainties (Figure 6).

We created a ΔDEM from the first and last late~~-~~ summer DEM for each site (Figure 7) and compared calculated mass change from this ΔDEM to the sum of the individual balance years that comprised that given period (Figure 8). We found that all cumulative seasonal B_a estimates from glaciological and geodetic balance were within uncertainty (2σ) of the last-first mass change approach (Figure 8). Glaciological balance was in net more positive (average +0.09 m w.e.) and had an average absolute difference of ±0.20 m w.e. from the last-first ΔDEM. Summed B_{a_geod} ~~agreed~~agree with our last-first estimates, with an average deviation of only 0.03 m w.e.

3.2. Glaciological density observations

~~Average glacier-wide~~Glacier averaged snow density from snow pits and cores ~~on a glacier-wise basis~~ for spring is 457 ± 48 kg m^{-3}, with a coefficient of variation (CV) of 0.14 (n=74). This estimate is 13 kg m^{-3} less than our snow ~~course~~ survey based geodetic ρ_{spring} but is within uncertainty (Table 4). For Haig Glacier, average spring density is 420 ± 45 kg m^{-3}, 20 kg m^{-3} lighter than our estimate obtained from nearby snow ~~course~~survey measurements, but again within uncertainty. Our average late~~-~~ summer glaciological density of 570 ± 20 kg m^{-3} (n=27) ranged from 536 to 617 kg m^{-3} (CV=0.04). Assigned geodetic ρ_{snow} is 18 kg m^{-3} greater than observations. Average probe depth for spring is 4.20 ± 0.06 m, with a CV of 0.33 (n=1,754). Average probe depth in late~~-~~ summer is 1.85 ± 0.10 m, with a CV of 0.78 (n=777). Observed glacier-wide average snow depths are ~~typically~~ between 3.4 and 6.9 m, and average 4.56 ± 0.21 m. While spring snow density showed greater variability than late~~-~~ summer snow density, snow depth is far more variable than snow density in both seasons.

Over the period ~~2015-~~2015–2017 ~~our glaciers had an~~ average AAR ~~of 0.~~was 38% (Table 3), with multi-year firn exposed over 13% of the glacier surface, thus leaving the remaining 49% of glacier area as bare ice. Located in the Rocky Mountains, Haig Glacier is the easternmost site in our study and is in a lower~~-~~ accumulation environment. It has lost nearly all its firn cover over the last 20 years, with firn area ~~reduced to 0.06~~at 6% in 2015. ~~Excluding this site, the remaining~~The study glaciers that lie in the Columbia Mountains had an AAR of ~~0.~~45% with ~~0.~~15% exposed multi-year firn cover and ~~0.~~40% bare glacier ice.

3.2.1 Geodetic density sensitivity

The effect of using a regional literature-based late~~-~~ summer snow density (Table 5) versus our ~~glacier-wise~~ glaciological density values (Table S1) depends on the amount of retained snow and glaciological density but produces a < 0.01 m w.e.

decrease on average, a negligible contribution. Varying firn density by ±15% also has an average effect of ±0.01 m w.e., with the largest impact (0.04 m w.e.) experienced at Conrad Glacier in 2015, when 17% of the glacier was exposed firn. However, misclassifying a given ~~pixel or~~ area of glacier surface has a significant impact, as ρ_{firn} is 17% greater than snow and ~~ρ_{ice} is~~ 26% ~~greater~~less than ~~ρ_{firn}~~ρ_{ice}. If we produce a single glacier-wide density (ρ) instead of distributing density based on surface classification, we change absolute magnitudes of B_{a_geod} by an average of ±0.10 m w.e. Though we did not use it for mass conversion, our ρ of B_{a_geod} ranged from 681 (Kokanee 2016) to 895 kg m^{-3} (Haig 2017) and averaged 748 ± 61 kg m^{-3}. Applying our snow ~~course~~survey density values for spring snow (Table 4) versus our glaciological snow density observations (Table S1) reduces average B_{w_geod} by 0.03 m w.e. (1.7%) and causes B_{w_geod} to be ~~5~~from 7% greater rather than 5% relative to B_{w_glac} ~~compared to 7% greater for the glaciological-density-based B_{w_geod} estimates. Glacier-wise~~. For individual glaciers, B_{w_geod} values between the two methods differ by 1 to 13%, ~~and~~but only 2% on average.

3.3 Glaciological and Geodetic Balance Discrepancies

~~Glacier-wise estimates~~Estimates of seasonal and annual balance for individual glaciers were outside 1σ uncertainties in a few cases. Conrad B_{w_glac} was 24% greater than B_{w_geod} in 2016. Snow accumulation may have occurred in the eight days between the Conrad Glacier ALS survey and field visit, as we observed over 1 m of fresh snow over four days during that interval while on Kokanee Glacier. Automatic snow weather stations near both glaciers at around 2050 m asl showed no accumulation, highlighting the steep balance gradient of the Columbia Mountains. Additionally, ALS acquisition failed over the terminus of the Conrad and Illecillewaet Glaciers in late~~-~~ summer 2015~~,~~ (Table 2), and our extrapolation based upon the typical gradient over the terminus may have underestimated melt (Figure 7). Kokanee Glacier B_{a_glac} in 2017 was 0.25 m w.e. more positive than B_{a_geod}, likely due to the burial of a few ablation stakes, and ~~sub-freezing~~subfreezing temperatures which limited our ability to take adequate snow measurements. Illecillewaet Glacier B_{w_glac} in 2017 was 46% higher than B_{w_geod}, but this difference may stem from limited B_{w_glac} observations that year (n=3). ~~Although the terminus was not acquired in the ALS survey, we only missed 3% of glacier area.~~

3.4 Interannual and spatial variability

Varied climatological conditions provided a range of balance outcomes for the period of study. The lowest B_{w_glac} of the four studied winters (1.81 ± 0.12 m w.e.) occurred in 2016, yet also the least mass loss with an average B_{a_glac} of -0.36 ± 0.17 m w.e. (Figure 5). The 2016~~-~~–2017 winter brought the greatest snowpack of our study period, 2.08 ± 0.18 m w.e., yet substantial mass loss was ~~still~~ observed (average B_{a_glac}: -0.84 ± 0.23 m w.e.). The balance year of 2014~~-~~–2015 saw high sustained mass loss (average B_{a_glac} of -1.30 ± 0.13 m w.e.), despite having an B_{w_glac} within 0.01 m w.e. of 2016.

The standard deviation between the seasonal and annual balances for each glacier reveals that B_w (σ = 0.14 m w.e., 7%) experiences lower interannual variability than B_s (σ = 0.38 m w.e., 14%) and B_a (σ = 0.35 m w.e., 56%). Kokanee Glacier experienced the highest B_w in all four years ~~2015-~~2015–2018 averaging 2.34 ± 0.30 m w.e. (Figure 6), while Haig ~~Glacier's~~

Glacier B_w was lowest, averaging 1.37 ± 0.11 m w.e., ~~and~~coupled with the highest mass loss~~, averaging~~ (average B_{a_glac} ~~=~~: -1.62 ± 0.34 m w.e~~.~~.).

We did not investigate the influence of crevasses for each glacier and each season, but for a test case for each glacier (n=6) we created DEMs with filled crevasses in the late~~-~~ summer, and then produced a ΔDEM. We found that crevasse-free ΔDEM B_w was on average <1% smaller than our standard B_w, with discrepancies up to -0.05 m w.e or -3%. The amount of crevassing is important, however, as some of ~~our~~the studied glaciers such as the Zillmer, Nordic and Conrad feature large crevasse fields.

4 Discussion

The consistency between our geodetic and glaciological seasonal balance estimates among six glaciers over multiple years implies that high-resolution geodetic surveys can be used to reliably measure both winter and summer mass balance. Our study builds upon previous work that established the feasibility of geodetic methods to accurately produce B_w (Belart et al., 2017; Sold et al., 2013), and B_a (Klug et al., 2018). While others show that geodetic surveys can be applied for a single winter (Belart et al., 2017; Sold et al., 2013) or for one glacier over a number of years (Klug et al., 2018), our study demonstrates remotely~~-~~measured seasonal balance is possible for widely varying rates of accumulation and ablation for multiple glaciers across entire mountain ranges.

4.1 Geodetic seasonal Balance

Our small estimate of ~~$\sigma h_{\Delta DEM}$~~$\sigma h_{\Delta DEM}$ (Table S3) and bias correction (Table 3) highlight that height change uncertainty is generally minor~~; these terms are still~~, but is important to quantify~~, however~~ (Joerg et al., 2012; Klug et al., 2018). As described below, density distribution and conversion factors comprise a large portion of total mass change uncertainty, with firn compaction, fresh snow at the time of ALS acquisition, and crevasses also contributing.

Field Code Changed

The spatial coverage of ALS is ~~far~~ superior to glaciological observations; however, isolating the ~~snow-depth portion~~mass change component of surface height change at a given location is difficult and requires detailed input data (Belart et al., 2017; Sold et al., 2013). While we can develop balance gradients from glaciological data, we have not attempted to do so using our ALS data. To date, studies have differenced their glaciological and geodetic data regarding surface height change and assigned the difference as a combination of vertical ice velocity and firn compaction (Beedle et al., 2014; Belart et al., 2017; Sold et al., 2013) or used full-Stokes ice flow model with a bedrock DEM, a surface DEM, and in situ GPS~~-observed horizontal~~ velocities as inputs (Belart et al., 2017). Then, after applying a simple firn model, vertical ice velocity is estimated. While this method appears robust, and differencing of our glaciological observations of height change from our ΔDEMs produces realistic emergence/submergence velocities, it is beyond the scope of this study.

4.1.1 Density distribution and conversion factors

Converting volume to mass change is a major challenge for geodetic studies (Huss, 2013; Moholdt et al., 2010). Over multiple years to decades, a constant value of density can produce tolerable uncertainty of mass change (Huss, 2013). For shorter timescales, and particularly for seasonal balance, a careful consideration of density is necessary (Klug et al., 2018). Klug et al (2018) used ALS intensity data and satellite imagery for a pixel-based classification of the glacier surface as firn and ice. Our ~~mapping~~study built on this work and mapped areas of ice, but also distinguished between snow and firn. To investigate the influence of density assumptions, we compare using independent estimates of density and our glaciological data to inform our geodetic estimates, to better constrain the uncertainty on, and compare against, glaciological seasonal balance. Varying the density assigned to each surface class to the maximum and minimum values within our conservative uncertainties has minor effect on seasonal balance but failing to distribute them appropriately has a large impact. If a single density value is used, the range of values of ρ of B_{a_geod} indicates that ~~a value close to~~ 750 ± 60 kg m^{-3} would be most appropriate for seasonal mass change over this period (Table 3). Given the spread of ρ between glaciers, however, a glacier-~~wise~~specific ρ would be best.

Like Klug et al. (2018), our applied firn density was selected based on a core from a temperate glacier in the Alps (Ambach et al., 1966), and our in~~-~~ situ density measurements for firn ≥2 years old matched this value (Table 4). Our glaciological density values for one-year-old firn and late~~-~~ summer snow density are respectively 13.1 and 22.4% (Table 4) less than the assumed value of 700 kg m^{-3} for both snow and firn taken by Klug et al. (2018). Had we also ~~taken this value for our~~combined snow~~-covered areas~~ and firn density, we would have ~~introduced a negative bias to~~biased B_{a_geod} by varying magnitudes depending upon the surface cover. As glacier mass loss rates continue to accelerate (Menounos et al., 2018), it is reasonable to expect more and older exposed firn during the ablation season, which for geodetic studies, may imply a higher density conversion factor for firn.

Applying glaciological late~~-~~ summer snow density versus our independent regional average density (Table 5) had little effect on B_{a_geod}. Future geodetic studies should use the best available local data, however, as different regions and mountain ranges have different late~~-~~ summer densities (Table 5).

Using our glaciological winter density values to produce B_{w_geod} estimates resulted in a slightly greater discrepancy relative to B_{w_glac} than applying our snow~~-course-based densities (Table 3). The two B_{w_geod} estimates produced similar results in net, however, with glacier-wise only a 2% difference between B_{w_geod} estimates. Our regression-slope approach (Figure 3) to adjust glaciological observations of spring snow density (Table S1), is suitable over the period mid-April through late-May, but we caution against its use for other periods of the year when densification is far slower and less predictable. For Haig Glacier, a linear relation also appears to exist between mid-April through late-May (Marshall, 2012, p.18, Fig. 2.3). Coincident with our spring field visits and ALS surveys, springtime warming is influencing the snowpack, and homogenizing~~ survey-based densities (Table 3). The two B_{w_geod} estimates produced similar results in net, and only a 2% average difference between B_{w_geod} estimates for individual glaciers. In the Columbia and Rocky Mountains, the first significant warming event of the spring typically occurs~~ snow density (Adams, 1976; Elder et al., 1991). In western Canada, onset of snow melt is occurring earlier on average relative to 1970-2006 (Déry et al., 2009). However, great variability still exists, with the first significant warming event of the spring occurring in the Columbia and Rocky Mountains~~ between early April and early May (Marshall, 2012). Springtime warming tends to homogenize and increase snow density (Adams, 1976; Elder et al., 1991). ~~The tendency for a more homogenous snow density, and lack of~~Our linear regression approach (Figure 3) to adjust glaciological observations of spring snow density (Table S1), appears suitable over the period mid-April through late-May, but we caution against its use for other periods of the year when densification is far slower and less predictable. For Haig Glacier, a linear relation also exists between mid-April through late May (Marshall, 2012, p.18, Fig. 2.3). The tendency for a more homogenous snow density, and lack of a consistent altitudinal trend both lend credence to using a single snow density (Fausto et al., 2018; McGrath et al., 2018).

4.1.2 Firn ~~compaction~~and internal processes

While firn compaction is only incorporated in our uncertainty analysis, others estimate its effect to derive B_{w_geod} (Belart et al., 2017; Sold et al., 2013), but not ~~for~~ B_{a_geod} (Klug et al., 2018). For B_{w_geod}, firn compaction was estimated based upon the annual balance in the accumulation zone over a decade (Sold et al., 2013) or over a single balance year (Belart et al., 2017). ~~The~~Currently accumulation areas on ~~our~~alpine glaciers are in constant flux, ~~however, and~~ are ~~nearly always~~typically discontinuous. Exposed firn is ~~the norm,~~common (Figure S1), implying that the firn zone on our study sites is shrinking in area and thickness, interrupting the cycle of firnification, and invalidating firnification models which assume that one annual layer is transformed from snow to ice annually. Nevertheless, a carefully considered treatment of firn could improve seasonal geodetic balance estimates, but as demonstrated by Belart et al. (2017), firn and fresh snow densification have little effect on B_{w_geod} if the magnitude of winter accumulation is large. For regions with low winter balance, or a colder climate, compaction would have a larger relative influence on B_w~~. We also chose not to apply a firn correction since it requires glaciological measurements that we purposely withheld in order to evaluate the feasibility of measuring seasonal balance without surface observations from the glaciers~~.

Meltwater retention is not incorporated into our annual balance estimates. At Haig Glacier, firn meltwater retention has not been measured, but meltwater retention in the supraglacial snowpack is a negligible contributor to mass balance, though it does create an effective “energy sink”, that should be accounted for in mass balance modeling (Samimi and Marshall, 2017). For glaciers in Svalbard, coupled energy balance and snow/hydrology models have been used to quantify the effects of meltwater freezing and retention on glacier mass balance (Van Pelt et al., 2012; Van Pelt and Kohler, 2015). Rates of meltwater retention are decreasing for Svalbard glaciers (Van Pelt and Kohler, 2015), and on the Devon Ice Cap (Bezeau et al., 2013), due to decreasing firn area and in particular, warmer temperature. Like at our glaciers, melt-freeze cycles form thick ‘summer surface’ layers on these Svalbard glaciers and Devon Ice Cap, which could act as a barrier for vertical water transport and is likely to promote near-surface lateral water flow, limiting deep firn water storage (Gascon et al., 2013; Van Pelt and Kohler, 2015).

Geodetic balance implicitly includes internal and basal mass change, which are not captured by the glaciological method. Studies of these processes are rare and are based upon estimates rather than verified measurements. Estimates of annual mass loss from geothermal heat, potential energy released by runoff or ice motion, and basal friction are typically around 0.01 to 0.10 m w.e. (Huss et al., 2009; Oerlemans, 2013; Sold et al., 2016). Crevasses and internal processes likely combine to be 0 to 4% of the magnitude of average annual ablation (e.g. Klug et al., 2018; Sold et al., 2016), and thus are likely not major contributors to seasonal balance in the Columbia Mountains. Modeled meltwater accumulation in firn would tend to increase mass balance, possibly offsetting typical basal/internal mass loss, but would not be captured by geodetic or glaciological measurements. Most mass balance models only assume vertical percolation of meltwater yet given thick impermeable ice layers observed in our cores and snow pits, and in other studies (Gascon et al., 2013; Van Pelt and Kohler, 2015), this assumption would lead to an overestimation of refreezing. Without regional data to constrain firn processes it is difficult to incorporate them into mass balance calculations. Regionally, a better understanding of firn processes could improve annual balance and runoff estimates, and likely has a greater influence on the large icefields in western North America, which have received little attention. Although firn processes are not resolved, our approach markedly improves the quality of annual results compared to calculations based on a fixed glacier-wide conversion density.

4.1.3 Fresh ~~Snow~~snow

Presence of fresh snow at the time of acquisition is a challenge for any geodetic survey estimating mass change ~~(Belart et al., 2017; Joerg et al., 2012; Klug et al., 2018)~~(Belart et al., 2017; Joerg et al., 2012; Klug et al., 2018). Fresh snow can change the height of the target surface by tens of centimeters. Our bias~~-~~ correction of ΔDEM height change (Figure S2, Table 3) corrected for small quantities of fresh snow, assuming that snow was present over stable terrain. In late~~-~~ summer, we could detect fresh snow visually, as a hillshade of the glacier surface at 1~~-~~ m resolution captures intricate details which are easily disguised by snow depths of 0.2 m or more. Off-glacier, the depth and distribution of fresh snow is variable due to redistribution and the thermal properties of bedrock and other surfaces~~, and redistribution~~. In spring, we are unable to detect fresh snow as the only snow-free pixels in our scenes are typically rock faces with extreme slopes and tree~~-~~ tops. Our $\sigma\Delta M_{sys}$ attempts to approximate the systematic uncertainty introduced by fresh snow and firn compaction.

4.1.4 Crevasses ~~and internal mass change~~

Crevasses can affect both B_{w_geod}~~,~~ and B_{w_glac} since crevasses bridged by winter snowpack will overestimate B_{w_geod} snow volume, and crevasses filled by snow would not be captured by B_{w_glac}. We produced ‘crevasse-free’ glacier surfaces by resampling late~~-~~ summer DEMs to 10 m using the maximum elevations within the smoothing window to avoid in-crevasse height measurements. Using the 10 m ‘crevasse-free’ DEMs versus the original 1~~-~~ m DEMs had little influence on B_{w_geod}, with only the Zillmer and Nordic glaciers showing a difference >1%. We did not extend these test cases to cover B_{a_geod} estimates because the ~~number and~~ area of exposed crevasses varied little year to year. On Hintereisferner, crevasse effects biased B_{a_geod} by only 0.03% (-0.047 m w.e.) over a decade (Klug et al., 2018). Despite the small influence of crevassing on B_{a_geod} observed in this study, additional studies should quantify the magnitude of this bias~~. Our field operations have been impacted by the melting out of crevasses: as strongly negative years are becoming the norm, and glacier flux is likely decreasing, crevasses are exposed for longer periods of time, and slower to close. This means that the total void area of~~

~~crevasses is increasing due to ablation, which we have observed on Conrad, Zillmer, Nordic, and Haig glaciers, which could possibly increase their influence on B_w.~~ in greater detail.

~~Geodetic balance implicitly includes internal and basal mass change, which are not captured by the glaciological method. Studies of these processes are rare and are based upon estimates rather than verified measurements. Estimates of annual mass loss from geothermal heat, potential energy released by runoff or ice motion, and basal friction are typically around 0.01 to 0.10 m w.e. (Huss et al., 2009; Oerlemans, 2013; Sold et al., 2016). Crevasses and internal processes likely combine to be 0 to 4% of the magnitude of average annual ablation per literature estimates (e.g. Klug et al., 2018; Sold et al., 2016), and thus do not appear to be important contributors to seasonal balance in the Columbia Mountains. Further, the capacity for the firn reservoir of a glacier to retain meltwater may be approximated by a simple firn model, and would tend to increase mass balance, possibly offsetting typical basal/internal mass loss, but would not be captured by geodetic or glaciological measurements.~~

4.2 Glaciological seasonal balance

Observational biases include the representativeness of sampling sites and number of measurements (Cogley, 1999; Fountain and Vecchia, 1999), and the extrapolation of those measurements to produce a glacier-wide balance estimate (Sold et al., 2016; Thibert and Vincent, 2009). The difficulty of comparability between methods and sites (Cogley, 1999; Fountain and Vecchia, 1999) is an ongoing challenge due to logistical and financial obstacles to in~~-~~ situ mass balance studies. Areas of a glacier may be inaccessible, and preferred paths chosen for measurement may be biased to areas which better retain snowpack for safety purposes (Østrem and Brugman, 1991). ~~Our glaciological measurement densities ranged from 0.5 to 18.5 points km^{-2} (Table 2), whereas our ALS data had around one million points km^{-2}.~~

4.2.1 Snow depth

We observed best agreement between geodetic and glaciological measurements of winter balance during years of dense field surveys. Safety or logistical constraints prevented us from completing all transects of snow depth measurements in some years, ~~and again,~~with greater ~~discrepancy was found~~discrepancies between estimates in cases ~~where our coverage was~~with incomplete coverage. In both spring and late~~-~~ summer, we encountered internal ice layers at some or all sites. Ice lenses were most common in the accumulation zone, but they were also found in the ablation zone in spring. These internal layers form via rain-on-snow events (McCabe et al., 2007) or, as the ~~year~~melt season progresses, via internal storage of meltwater (Pfeffer and Humphrey, 1996). Ice layers 2~~-~~–6 cm thick were present nearly every year in the accumulation zone of the Conrad Glacier, and often at other sites. We were able to penetrate these layers and successfully measure spring snow depth using our industrial avalanche probe. A conventional avalanche probe is unsuitable for glaciological observations in ~~our region~~the Columbia Mountains.

The greater B_{w_geod} of 2016 on Conrad Glacier ~~may in part~~is likely due to both snow accumulation between the glaciological visit and ALS survey, and to the late~~-~~ summer 2015 ALS survey missing the lowest reaches of the glacier, preventing calculation of surface height change for that portion of the glacier. We estimated the snow depth for the lower reaches of the glacier based upon the ratio of snow depth observed there for other years relative to the rest of glacier, and snow depths along the cut-off margin. The B_w discrepancy for Zillmer Glacier in 2016 is likely due to glaciological sampling bias, as the east transect (Figure 4), which has a lesser snowpack, was not sampled, and the 30~~-~~ day difference between field and ALS survey date (Table 2) may not be ~~fully~~adequately resolved by the GPS survey correction.

4.2.2 Mass change between measurements

Previous studies account for mass change that occurs between measurements by using a distributed temperature index model (Sold et al., 2013) or degree-day model (Belart et al., 2017), but these models do not account for snow gain. We utilized in~~-~~situ GPS surveys of the glacier height which were then compared with ALS DEMs. We binned and averaged our height change estimates by 100 m elevation bands (Figure S5), and then applied a density to each band based on satellite observations of a given surface class. Limitations in our approach include: 1) fresh snowfall ~~in spring or late-summer, between the GPS and ALS surveys; and 2) significant densification of the snowpack in spring. A weather station situated proximal to the glacier that could be used to drive a surface mass balance model (Fitzpatrick et al., 2017) would be required to capture local changes in glacier mass due to precipitation events and compaction.~~between the GPS and ALS surveys; and 2) significant densification of the snowpack in spring. Terrain presents a further challenge to kinematic GPS survey observations. The GPS antenna is securely mounted in the backpack of a field member, but the measured height of the antenna above the glacier surface may vary due to the uneven glacier terrain, particularly during travel on steep slopes (Beedle et al., 2014).

Our median dates of late~~-~~ summer glaciological visits and geodetic surveys are September 6th, and September 18th respectively (Table 2). Snowfall can occur at any time of the year in the Columbia and Rocky Mountains (Schnorbus et al., 2014), and in late August, throughout September, and even into early October, either melt or accumulation can prevail (Marshall, 2014). Lowering of the surface via ablation post ALS survey dates (Table 2) is not accounted for and would cause an underestimated winter snowpack. While our methods are comparable year-to-year, and between sites, our B_w and B_s values are not the total amount of snow and runoff during a year. We do not include snow which falls between May and August and melts off and cannot measure ablation occurring after our ALS survey or glaciological visit, whichever occurs later. Thus, our B_w and B_s values represent a ~~low threshold~~conservative estimate of runoff contributions from snow and ice melt.

~~The close agreement between geodetic and glaciological seasonal balance estimates might be fortuitous in that most of our glaciers have moderate mass turnover; regions with low mass turnover would require treatment of more ancillary factors such as firn compaction.~~

5. Conclusions

Estimates of seasonal mass balance presented here show strong agreement between glaciological and geodetic methods ~~on a glacier-by-glacier basis~~for individual glaciers, and are within 1σ uncertainties for average winter, summer, and annual balance. These independent estimates of seasonal mass change accord over three years from glaciers separated by hundreds of kilometers. Our findings suggest that high-resolution geodetic methods, such as from ALS (Klug et al., 2018; Sold et al., 2013), aerial photogrammetry (Nolan et al., 2015), and stereo satellite imagery (Belart et al., 2017; Berthier et al., 2014) can be used to produce accurate seasonal and annual balance estimates over large areas. The quality of geodetic annual balance estimates depends more on distributing density via surface classification (Klug et al., 2018), than on the density values themselves. The spatial coverage, density of observations, and measurement precision of high-resolution geodetic terrain analysis compensates for ~~errors~~uncertainty associated with fresh snow and firn compaction, internal and basal mass change, and crevasses (Belart et al., 2017; Klug et al., 2018). The minimal impact of these factors on mass balance stems from the large mass changes observed at our sites, as reported elsewhere (Belart et al., 2017; Klug et al., 2018). For glaciers with low mass turnover, errors introduced by firn compaction, crevasses, and fresh snow may be ~~considerable~~considerably larger than observed in our study, however. Our estimate of spring snow density for geodetic measurements ~~based upon~~from provincial snow ~~course~~survey observations (Figure 2) is within the uncertainty of our measured glaciological spring snow density (Table 4). Our approach holds promise for being able to use regional density estimates when in situ measurements are unavailable, yet discrepancies of up to 13% between geodetic and glaciological winter balance estimates indicate the uncertainty introduced when using density values which are not site-specific. Estimates of end-of-season snow density introduce a possible bias, but given the regional consistency of late~~-~~ summer snow density, and the overall lack of a density-altitude gradient in spring, using a single snow density is a robust method for converting snow depth to water equivalence (Fausto et al., 2018; McGrath et al., 2018). We observed greater variability in B_s relative to B_w, highlighting the need for models of glacier mass balance that can be able to reliably reproduce widely varying rates of mass loss corresponding to the multitude of energy fluxes that influence alpine glaciers (Fitzpatrick et al., 2017), to reliably estimate seasonal mass balance.

The hydrologic cycle of western North America is dominated by snowfall in the mountains, but observations of alpine snowpack above 2000 m asl are sparse. As the climate continues to change, there is a growing need for a more detailed understanding of the seasonal balance of glaciers and snowpack. Geodetic methods are needed to supplement in~~-~~ situ observations across many mountain regions in order to address the contribution of ~~glacier~~glaciers to ~~sea level rise and~~ changes in freshwater runoff availability, and to sea level rise. To date, the majority of high-resolution geodetic balance studies of seasonal or annual balance have been conducted in the European Alps, where extensive, multi-decadal glaciological data are available (Klug et al., 2018; Sold et al., 2013, 2016). Our study suggests that geodetic methods can be used to assess seasonal balance of glaciers, even in mountain ranges lacking long-term records of mass balance, ~~as long as~~if density is carefully considered (Belart et al., 2017; Klug et al., 2018). Recent advances in satellite technology (Berthier et al., 2014; Marti et al., 2016) suggest that such efforts can be made with increasing spatial and temporal coverage, greatly adding to our understanding of the seasonal contribution of snow and glaciers to mountain hydrology~~, on which so much depends, but so little is known~~.

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Table 1: Glacier specific details. Firn ratio ~~is~~refers to the area of a glacier covered by multi-year firn~~:~~, which is the combination of accumulation area and exposed firn ~~and accumulation area~~ from 2015 imagery.

Glacier	Area (km^2)	Max Elev. (m)	Min Elev. (m)	Range (m)	Mean Elev. (m)	Length (km)	Firn Ratio	Aspect
Zillmer	5.43	2860	1860	1000	2380	5.59	0.59	NW
Nordic	3.39	2990	2065	925	2515	3.30	0.62	N
Illecillewaet	7.72	2908	2147	761	2532	4.29	0.48	WNW
Haig	2.62	2870	2461	409	2660	2.45	0.06	SE
Conrad	11.45	3235	1825	1410	2595	12.18	0.58	N
Kokanee	1.79	2805	2220	585	2585	2.20	0.48	N

Table 2: ~~Acquisition date~~Date and number of observation locations (n) for glaciological visits and geodetic ~~survey~~acquisition dates~~.~~ and point density. Field dates are median date of glacier visit.

Year	Glacier	~~Late-summer~~Autumn Glac.	n	~~Late-summer Geod.~~Autumn ALS	Cover (%)	Points m^{-2}	Winter Glac.	n	Winter ~~Geod.~~ALS	Cover (%)
2015	Zillmer	8/23/2015	23	10/3/2015	100	2.75	5/30/2015	20	4/19/2015	100
2016	Zillmer	8/15/2016	23	9/14/2016	100	2.44	4/14/2016	46	4/18/2016	100
2017	Zillmer	8/22/2017	26	11/3/2017	100	1.49	4/13/2017	31	5/20/2017	100
2018	Zillmer	~~na~~—	~~na~~—	~~na~~—	~~na~~—	—	5/19/2018	42	4/29/2018	100
2014	Nordic	8/29/2014	8	9/11/2014	100	8.71	4/27/2014	16	~~na~~—	~~na~~—
2015	Nordic	8/31/2015	~~10~~11	9/11/2015	99	1.99	5/1/2015	20	4/19/2015	100
2016	Nordic	8/21/2016	21	9/12/2016	99	3.27	5/2/2016	28	4/17/2016	100
2017	Nordic	9/14/2017	18	9/27/2017	100	2.35	5/1/2017	21	5/21/2017	100
2018	Nordic	~~na~~—	~~na~~—	~~na~~—	~~na~~—	—	5/1/2018	21	4/26/2018	100
2015	Illecillewaet	9/24/2015	9	9/11/2015	97	~~na~~1.02	~~na~~—	—	4/19/2015	100
2016	Illecillewaet	9/13/2016	7	9/12/2016	100	~~na~~1.37	~~na~~—	—	4/17/2016	100
2017	Illecillewaet	9/27/2017	7	9/17/2017	100	2.59	5/19/2017	3	5/21/2017	100
2018	Illecillewaet	~~na~~—	~~na~~—	~~na~~—	~~na~~—	~~na~~—	~~na~~—	—	4/26/2018	100
2015	Haig	9/12/~~15~~2015	2	9/12/2015	100	0.93	5/12/2015	33	4/20/2015	100
2016	Haig	9/13/~~16~~2016	1	9/13/2016	100	1.85	5/18/2016	33	4/17/2016	100
2017	Haig	9/16/~~17~~2017	1	9/16/2017	97	4.82	5/12/2017	33	5/21/2017	100
2018	Haig	~~na~~—	~~na~~—	~~na~~—	~~na~~—	~~na~~—	~~na~~—	—	4/27/2018	100
2014	Conrad	9/4/2014	7	9/11/2014	100	~~na~~10.38	~~na~~—	~~na~~—	~~na~~—	—
2015	Conrad	9/5/2015	9	9/12/2015	92	1.35	4/23/2015	~~8~~38	4/20/2015	100
2016	Conrad	8/28/2016	31	9/12/2016	100	2.45	4/26/2016	44	4/17/2016	100
2017	Conrad	9/10/2017	42	9/17/2017	94	3.70	5/15/2017	59	5/21/2017	100
2018	Conrad	~~na~~—	~~na~~—	~~na~~—	~~na~~—	—	4/24/2018	56	4/26/2018	100
2015	Kokanee	8/27/2015	11	9/12/2015	100	1.04	4/20/2015	20	4/19/2015	100
2016	Kokanee	9/5/2016	23	9/13/2016	100	2.07	4/19/2016	33	4/17/2016	100
2017	Kokanee	9/19/2017	15	9/16/2017	83	2.63	4/17/2017	23	5/21/2017	100
2018	Kokanee	~~na~~—	~~na~~—	~~na~~—	~~na~~—	—	4/18/2018	21	4/26/2018	100

Table 3: Seasonal balance and uncertainty estimates for geodetic (geod) and glaciological mass balance (glac~~.)~~)) in m w.e. Kinematic GPS survey-derived corrections applied to glaciological data (surv.corr). Statistical analysis of the DEMs ~~in snow and ice-free~~over stable terrain include NMAD, median height difference, and bias correction applied over the glacier ($Bias_{\Delta h}$~~) and mean~~). Mean density of Ba_{geod}~~.~~ is $\bar{\rho}$. Average values ~~exclude~~include only cases where ~~only~~both geodetic ~~or~~and glaciologic data were collected. $Bw_{geod.gl}$ is calculated using glaciological densities (Table S1), and $Bw_{geod.ss}$ is calculated using snow ~~course~~survey data (Figure 2). Listed Bs_{geod} is derived using $Bw_{geod.ss}$ ~~, and Ba_{geod} used regional~~. Regional late-summer snow density (Table 5~~).~~) was used to calculate Ba_{geod}.

Year	Glacier	$Bw_{geod.gl} \pm \sigma_{geod.bw}$	$Bw_{geod.sc} \pm \sigma_{geod.bw}$	$Bs_{geod} \pm \sigma_{geod.bs}$	$Ba_{geod} \pm \sigma_{geod.ba}$	$Bw_{glac} \pm \sigma_{glac.bw}$	$Bs_{glac} \pm \sigma_{glac.bs}$	$Ba_{glac} \pm \sigma_{glac.ba}$	$Bw_{surv.corr}$	$Ba_{surv.corr}$	AAR	ELA (m)	NMAD Ba (m)	NMAD Bw (m)	Median $Ba_{\Delta h}$ (m)	$Bias_{\Delta h}$ (m)	$\bar{\rho}$ (kg m^{-3})
2018	Zillmer	1.70 ± 0.19	1.75 ± 0.20			1.65 ± 0.17			-0.15					1.4			
2018	Nordic	1.87 ± 0.26	2.07 ± 0.27			2.18 ± 0.14			-0.04					1.76			
2018	Illecillewaet	1.61 ± 0.17	1.65 ± 0.18			—			na					2.26			
2018	Haig	1.25 ± 0.15	1.31 ± 0.19			1.42 ± 0.15			na					1.83			
2018	Conrad	1.62 ± 0.21	1.84 ± 0.23			1.83 ± 0.12			0.00					2.34			
2018	Kokanee	2.07 ± 0.25	2.31 ± 0.26			2.25 ± 0.13			0.01					1.76			
2017	Zillmer	2.12 ± 0.24	2.03 ± 0.25	-2.70 ± 0.27	-0.67 ± 0.10	1.93 ± 0.26	-2.44 ± 0.35	-0.51 ± 0.23	0.15	-0.31	0.48	2440	0.6	1.83	-0.1	-0.05	729 ± 45
2017	Nordic	2.14 ± 0.29	2.18 ± 0.30	-2.77 ± 0.31	-0.59 ± 0.09	2.03 ± 0.22	-2.78 ± 0.32	-0.75 ± 0.23	-0.04	-0.10	0.39	2540	0.28	1.8	0.01	-0.09	732 ± 43
2017	Illecillewaet	1.47 ± 0.19	1.54 ± 0.20	-2.55 ± 0.27	-1.01 ± 0.18	2.00 ± 0.16	-2.84 ± 0.32	-0.84 ± 0.28	—	—	0.36	2615	0.32	2.19	0.01	0	718 ± 49
2017	Haig	1.58 ± 0.20	1.65 ± 0.23	-3.56 ± 0.31	-1.91 ± 0.21	1.50 ± 0.17	-3.43 ± 0.29	-1.93 ± 0.24	—	—	0.04	na	0.31	1.62	0.01	0.04	885 ± 10
2017	Conrad	2.10 ± 0.22	1.91 ± 0.23	-2.97 ± 0.26	-1.06 ± 0.13	2.17 ± 0.17	-3.12 ± 0.29	-0.95 ± 0.24	-0.16	-0.16	0.48	2600	0.31	2.68	0	-0.01	730 ± 45
2017	Kokanee	3.15 ± 0.32	2.86 ± 0.33	-3.14 ± 0.34	-0.28 ± 0.08	2.84 ± 0.25	-2.87 ± 0.34	-0.03 ± 0.23	0.00	0.01	0.62	2560	0.34	1.99	-0.08	-0.01	711 ± 55
2016	Zillmer	1.68 ± 0.19	1.72 ± 0.20	-2.27 ± 0.22	-0.55 ± 0.07	1.99 ± 0.23	-2.61 ± 0.33	-0.62 ± 0.24	0.02	-0.38	0.49	2410	0.21	1.76	0.01	-0.02	726 ± 46
2016	Nordic	1.79 ± 0.22	1.70 ± 0.23	-1.85 ± 0.24	-0.15 ± 0.08	1.79 ± 0.14	-1.90 ± 0.21	-0.11 ± 0.16	-0.08	0.01	0.43	2555	0.16	1.63	0	-0.04	727 ± 40
2016	Illecillewaet	1.41 ± 0.17	1.46 ± 0.18	-1.73 ± 0.18	-0.27 ± 0.05	—	—	-0.19 ± 0.28	—	—	0.60	2550	0.45	1.9	-0.01	0.05	718 ± 54
2016	Haig	1.15 ± 0.15	1.21 ± 0.17	-2.27 ± 0.20	-1.06 ± 0.11	1.34 ± 0.17	-2.49 ± 0.29	-1.15 ± 0.24	—	—	0.03	na	0.38	1.24	-0.01	-0.04	893 ± 10
2016	Conrad	1.40 ± 0.18	1.47 ± 0.19	-1.74 ± 0.20	-0.27 ± 0.06	1.88 ± 0.12	-2.08 ± 0.20	-0.20 ± 0.16	0.11	-0.13	0.55	2530	0.14	2.1	0	-0.02	734 ± 50
2016	Kokanee	1.98 ± 0.22	2.05 ± 0.23	-1.93 ± 0.23	+0.12 ± 0.05	2.07 ± 0.13	-1.94 ± 0.26	+0.13 ± 0.22	-0.05	0.12	0.72	2545	0.15	1.67	0	0	681 ± 64
2015	Zillmer	—	—	—	—	2.06 ± 0.30	-2.82 ± 0.40	-0.76 ± 0.27	0.00	-0.32	0.30	2500	—	—	—	—	—
2015	Nordic	1.74 ± 0.22	1.81 ± 0.23	-2.81 ± 0.28	-1.0 ± 0.16	1.83 ± 0.19	-3.02 ± 0.31	-1.19 ± 0.24	-0.16	0.06	0.32	2610	0.26	1.76	0	0.02	744 ± 42
2015	Illecillewaet	—	—	—	—	—	—	-1.17 ± 0.47	—	—	0.30	2600	—	—	—	—	—
2015	Haig	—	—	—	—	1.23 ± 0.25	-3.02 ± 0.25	-1.79 ± 0.25	—	—	0.00	na	—	—	—	—	—
2015	Conrad	1.65 ± 0.17	1.64 ± 0.18	-3.06 ± 0.24	-1.42 ± 0.16	1.80 ± 0.13	-3.20 ± 0.35	-1.40 ± 0.32	-0.02	-0.31	0.44	2685	0.21	2.2	-0.01	-0.03	736 ± 43
2015	Kokanee	—	—	—	—	2.18 ± 0.29	-3.38 ± 0.40	-1.20 ± 0.28	0.00	—	0.20	2680	—	—	—	—	—
All	Average	1.84 ± 0.11	1.88 ± 0.09	-2.59 ± 0.16	-0.72 ± 0.16	1.95 ± 0.08	-2.71 ± 0.13	-0.70 ± 0.15	-0.04	-0.14	0.38	2553	0.29	1.89	-0.01	-0.01	748 ± 62

Year	Glacier	$Bw_{geod.gl} \pm \sigma_{geod.bw}$	$Bw_{geod.sc} \pm \sigma_{geod.bw}$	$Bs_{geod} \pm \sigma_{geod.bs}$	$Ba_{geod} \pm \sigma_{geod.ba}$	$Bw_{glac} \pm \sigma_{glac.bw}$	$Bs_{glac} \pm \sigma_{glac.bs}$	$Ba_{glac} \pm \sigma_{glac.ba}$	$Bw_{surv.corr}$	$Ba_{surv.corr}$	AAR	ELA (m)	NMAD Ba (m)	NMAD Bw (m)	Median $Ba_{\Delta h}$ (m)	$Bias_{\Delta h}$ (m)	$\bar{\rho}$ (kg m^{-3})
2018	Zillmer	1.70 ± 0.19	1.75 ± 0.20			1.65 ± 0.17			-0.15					1.4			
2018	Nordic	1.87 ± 0.26	2.07 ± 0.27			2.18 ± 0.14			-0.04					1.76			
2018	Illecillewaet	1.61 ± 0.17	1.65 ± 0.18			–			na					2.26			
2018	Haig	1.25 ± 0.15	1.31 ± 0.19			1.42 ± 0.15			na					1.83			
2018	Conrad	1.62 ± 0.21	1.84 ± 0.23			1.83 ± 0.12			0.00					2.34			
2018	Kokanee	2.07 ± 0.25	2.31 ± 0.26			2.25 ± 0.13			0.01					1.76			
2017	Zillmer	2.12 ± 0.24	2.03 ± 0.25	-2.70 ± 0.27	-0.67 ± 0.10	1.93 ± 0.26	-2.44 ± 0.35	-0.51 ± 0.23	0.15	-0.31	0.48	2440	0.6	1.83	-0.1	-0.05	729 ± 45
2017	Nordic	2.14 ± 0.29	2.18 ± 0.30	-2.77 ± 0.31	-0.59 ± 0.09	2.03 ± 0.22	-2.78 ± 0.32	-0.75 ± 0.23	-0.04	-0.10	0.39	2540	0.28	1.8	0.01	-0.09	732 ± 43
2017	Illecillewaet	1.47 ± 0.19	1.54 ± 0.20	-2.55 ± 0.27	-1.01 ± 0.18	2.00 ± 0.16	-2.84 ± 0.32	-0.84 ± 0.28	–	–	0.36	2615	0.32	2.19	0.01	0	718 ± 49
2017	Haig	1.58 ± 0.20	1.65 ± 0.23	-3.56 ± 0.31	-1.91 ± 0.21	1.50 ± 0.17	-3.43 ± 0.29	-1.93 ± 0.24	–	–	0.04	na	0.31	1.62	0.01	0.04	885 ± 10
2017	Conrad	2.10 ± 0.22	1.91 ± 0.23	-2.97 ± 0.26	-1.06 ± 0.13	2.17 ± 0.17	-3.12 ± 0.29	-0.95 ± 0.24	-0.16	-0.16	0.48	2600	0.31	2.68	0	-0.01	730 ± 45
2017	Kokanee	3.15 ± 0.32	2.86 ± 0.33	-3.14 ± 0.34	-0.28 ± 0.08	2.84 ± 0.25	-2.87 ± 0.34	-0.03 ± 0.23	0.00	0.01	0.62	2560	0.34	1.99	-0.08	-0.01	711 ± 55
2016	Zillmer	1.68 ± 0.19	1.72 ± 0.20	-2.27 ± 0.22	-0.55 ± 0.07	1.99 ± 0.23	-2.61 ± 0.33	-0.62 ± 0.24	0.02	-0.38	0.49	2410	0.21	1.76	0.01	-0.02	726 ± 46
2016	Nordic	1.79 ± 0.22	1.70 ± 0.23	-1.85 ± 0.24	-0.15 ± 0.08	1.79 ± 0.14	-1.90 ± 0.21	-0.11 ± 0.16	-0.08	0.01	0.43	2555	0.16	1.63	0	-0.04	727 ± 40
2016	Illecillewaet	1.41 ± 0.17	1.46 ± 0.18	-1.73 ± 0.18	-0.27 ± 0.05	–	–	-0.19 ± 0.28	–	–	0.60	2550	0.45	1.9	-0.01	0.05	718 ± 54
2016	Haig	1.15 ± 0.15	1.21 ± 0.17	-2.27 ± 0.20	-1.06 ± 0.11	1.34 ± 0.17	-2.49 ± 0.29	-1.15 ± 0.24	–	–	0.03	na	0.38	1.24	-0.01	-0.04	893 ± 10
2016	Conrad	1.40 ± 0.18	1.47 ± 0.19	-1.74 ± 0.20	-0.27 ± 0.06	1.88 ± 0.12	-2.08 ± 0.20	-0.20 ± 0.16	0.11	-0.13	0.55	2530	0.14	2.1	0	-0.02	734 ± 50
2016	Kokanee	1.98 ± 0.22	2.05 ± 0.23	-1.93 ± 0.23	+0.12 ± 0.05	2.07 ± 0.13	-1.94 ± 0.26	+0.13 ± 0.22	-0.05	0.12	0.72	2545	0.15	1.67	0	0	681 ± 64
2015	Zillmer	–	–	–	–	2.06 ± 0.30	-2.82 ± 0.40	-0.76 ± 0.27	0.00	-0.32	0.30	2500	–	–	–	–	–
2015	Nordic	1.74 ± 0.22	1.81 ± 0.23	-2.81 ± 0.28	-1.0 ± 0.16	1.83 ± 0.19	-3.02 ± 0.31	-1.19 ± 0.24	-0.16	0.06	0.32	2610	0.26	1.76	0	0.02	744 ± 42
2015	Illecillewaet	–	–	–	–	–	–	-1.17 ± 0.47	–	–	0.30	2600	–	–	–	–	–
2015	Haig	–	–	–	–	1.23 ± 0.25	-3.02 ± 0.25	-1.79 ± 0.25	–	–	0.00	na	–	–	–	–	–
2015	Conrad	1.65 ± 0.17	1.64 ± 0.18	-3.06 ± 0.24	-1.42 ± 0.16	1.80 ± 0.13	-3.20 ± 0.35	-1.40 ± 0.32	-0.02	-0.31	0.44	2685	0.21	2.2	-0.01	-0.03	736 ± 43
2015	Kokanee	–	–	–	–	2.18 ± 0.29	-3.38 ± 0.40	-1.20 ± 0.28	0.00	–	0.20	2680	–	–	–	–	–
All	Average	1.87 ± 0.11	1.88 ± 0.09	-2.59 ± 0.16	-0.76 ± 0.16	1.95 ± 0.08	-2.67 ± 0.13	-0.73 ± 0.15	-0.04	-0.14	0.38	2553	0.29	1.89	-0.01	-0.01	748 ± 62

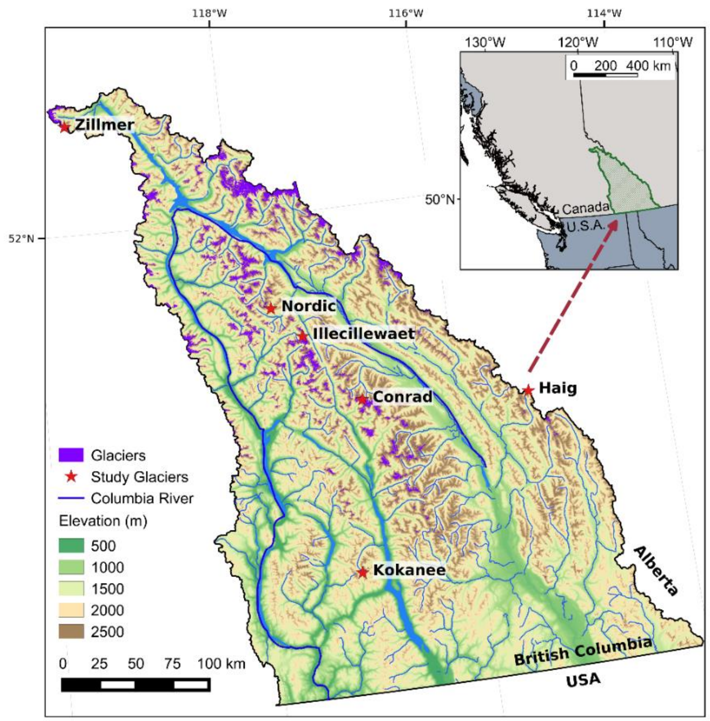
Table 4: Density values used for geodetic and glaciological balance. Glaciological values are average values.

Density	Geodetic (kg m^{-3})	Glaciological (kg m^{-3})	n
ρ_{spring}	470 ± 70*	457 ± 50*	74
ρ_{snow}	590 ± 90	570 ± 20	27
ρ_{firn}	700 ± 100	703 ± 65	4
ρ_{ice}	910 ± 10	—	—

***Geodetic spring snow density (ρ_{spring}) is 440 ± 50 kg m^{-3} for Haig Glacier and glaciological is 420 ± 45 kg m^{-3} (n = 46).**

Table 5: Late~~-~~ summer snow density observations from regional studies. We use ~~575~~570 kg m^{-3} as our density of late~~-~~ summer snow for geodetic mass balance, but also separately calculate mass balance using the average for regional studies excluding those from ~~our study~~ glaciers in this study (590 kg m^{-3}).

Location	Mean ρ_{snow} (kg m^{-3})	Range ρ_{snow} (kg m^{-3})	References
South Cascade Gl., WA, USA	580	530 – 600	(Bidlake et al., 2010; Krimmel, 1996)
Juneau Icefield, AK, USA	560	540 – 580	(Miller and Pelto, 1999; Pelto and Miller, 1990)
Castle Creek Gl., BC, CA	600	—	(Beedle et al., 2014)
North Cascades, WA, USA	600	590 – 630	(Pelto and Riedel, 2001)
Haig Glacier, AB, CA	545	530 – 570	(Marshall, 2012)
Columbia Basin, BC, CA	570	535 – 615	This study



~~**Figure 1: Map of the Canadian Columbia River Basin and locations of study sites.**~~

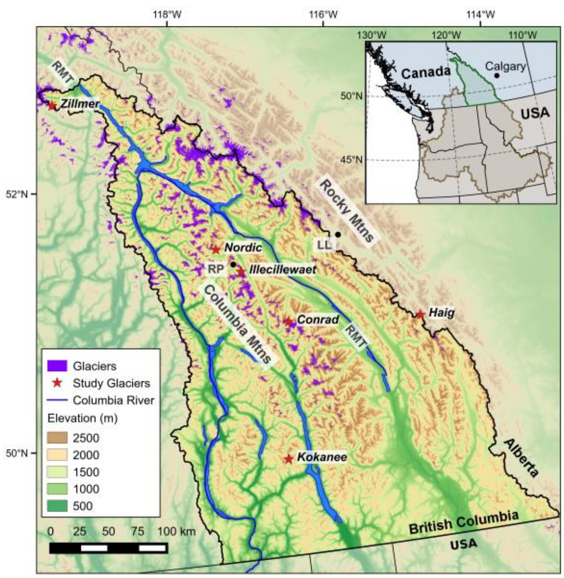


Figure 1: Map of the Canadian Columbia River Basin (black outline, brighter topography) and locations of study sites. Inset shows regional context of the Canadian portion of the Columbia River Basin which contributes to the river when it crosses the international border (green). The remainder of the basin is also depicted (brown). The Columbia and Rocky Mountains are separated by the Rocky Mountain Trench (RMT). Weather stations (black dots) at Rodgers Pass (RP) and Lake Louise (LL) are referred to in the introduction. Map coordinates are in NAD83/BCAlbers.

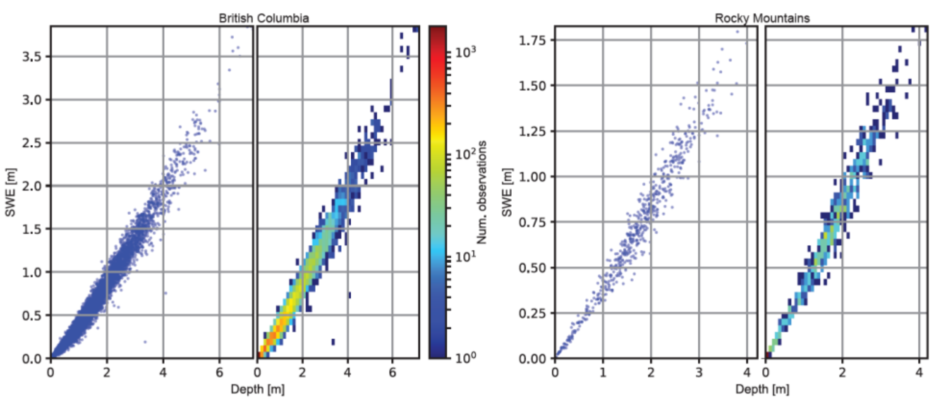
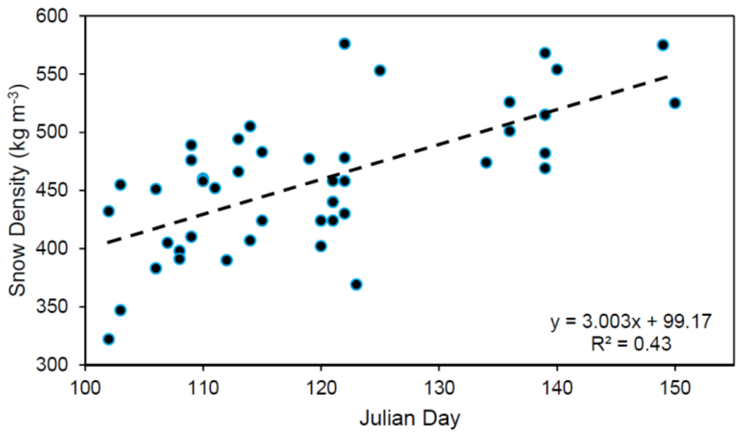


Figure 2: Snow depth versus snow water equivalent from May 1 provincial snow ~~course~~survey data. The mean date of our spring field seasons was May 1, and so we chose May 1 BC snow ~~course~~survey data (left) to derive a SWE/snow depth regression from

which we determined the average May 1 snow density ~~is (~~(470 ± 70 kg m^{-3} (r^2 = 0.97, n = 10,169~~).~~)). For Haig Glacier, we derived a regression from only snow stations within the Rocky Mountains ~~from~~south of Pine Pass ~~south~~ to derive winter density (440 ± 50 kg m^{-3} (r^2 = 0.97, n = 629~~)~~)).



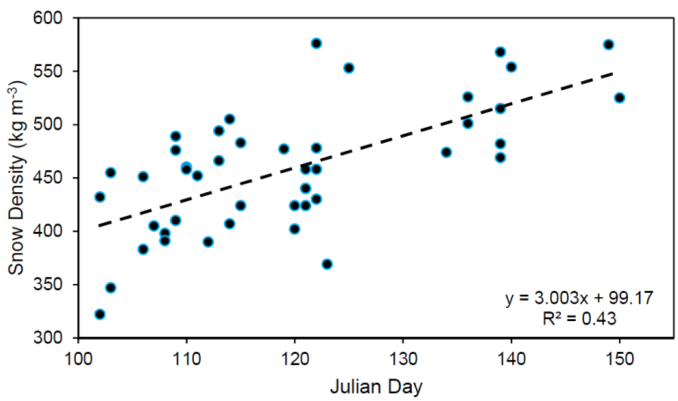
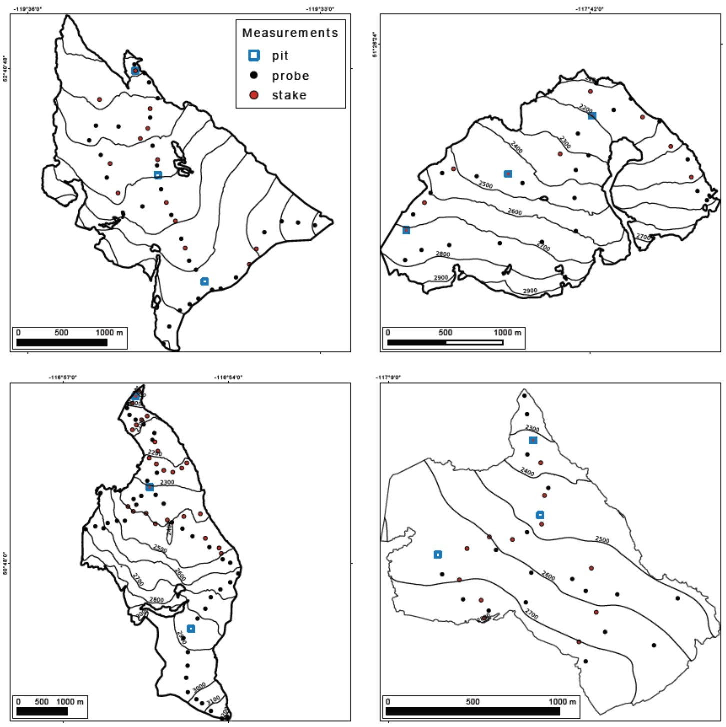


Figure 3: Snow density versus Julian day for all discrete snow pit and snow core locations (n = 46). For our glaciological density-informed estimates, we use the observed glacier-wide snow density and a linear regression of density versus day and used the slope (3.0 kg m^{-3} day^{-1} (r^2 = 0.43)) and days between the survey and the observations to adjust for change in snow density (Table S1).


Measurements
pit
probe
stake
0 500 1000 m

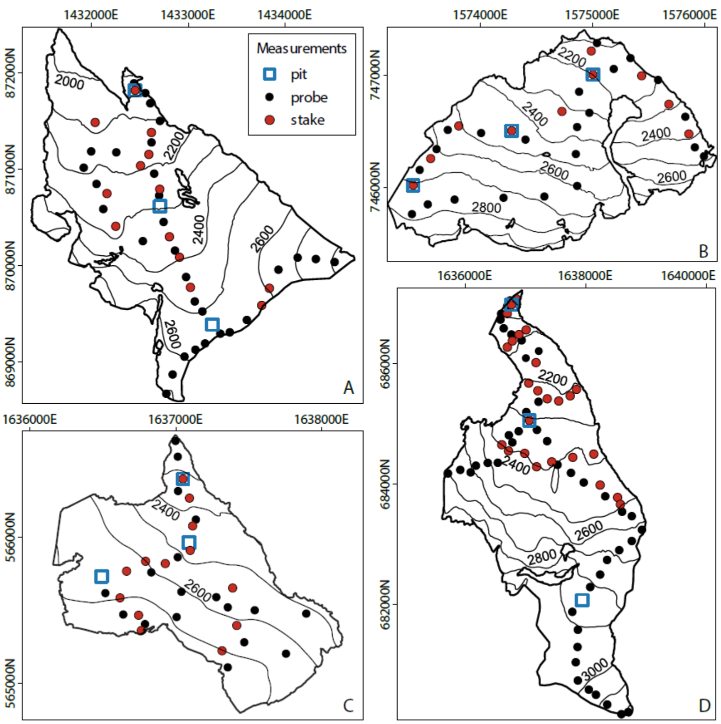


Figure 4: Measurement networks for the ~~(clockwise from top left~~A) Zillmer, B) Nordic, C) Conrad, and D) Kokanee~~, and Conrad~~ glaciers. Snow depth measurement locations, ablation stakes, and snow pit/core locations are pictured. Refer to Marshall et al. (2014) for the Haig Glacier, and Hirose and Marshall (2013) for the Illecillewaet Glacier, Map coordinates are in WGS84/UTM11N.

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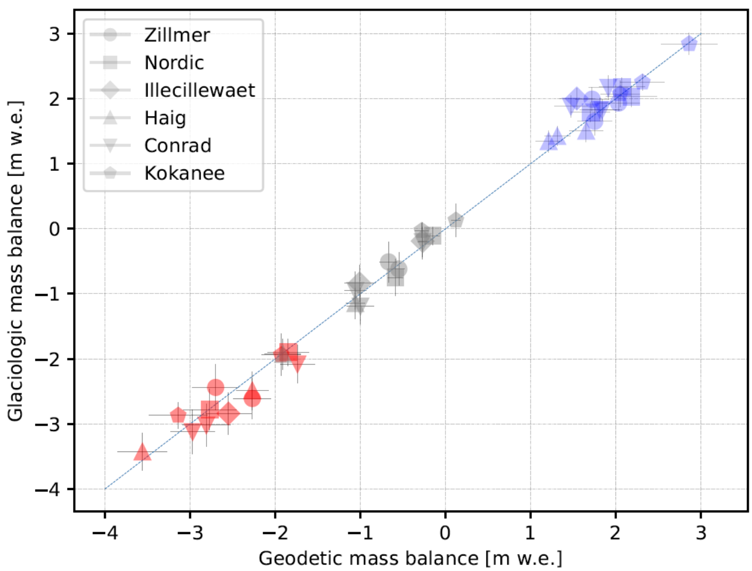


Figure 5: Geodetic versus glaciological mass balance estimates for 2015 through 2018 for all six study glaciers with a one-to-one line. Winter balance (blue) covers the accumulation season from mid-September to late April, summer balance (red) spanning the remaining months, and annual balance (grey). Errors depicted are 1σ uncertainties. Average B_{w_glac} was 4% greater than B_{w_geod}, and B_{s_glac} and B_{a_glac} were 4% greater than our geodetic estimates on average.

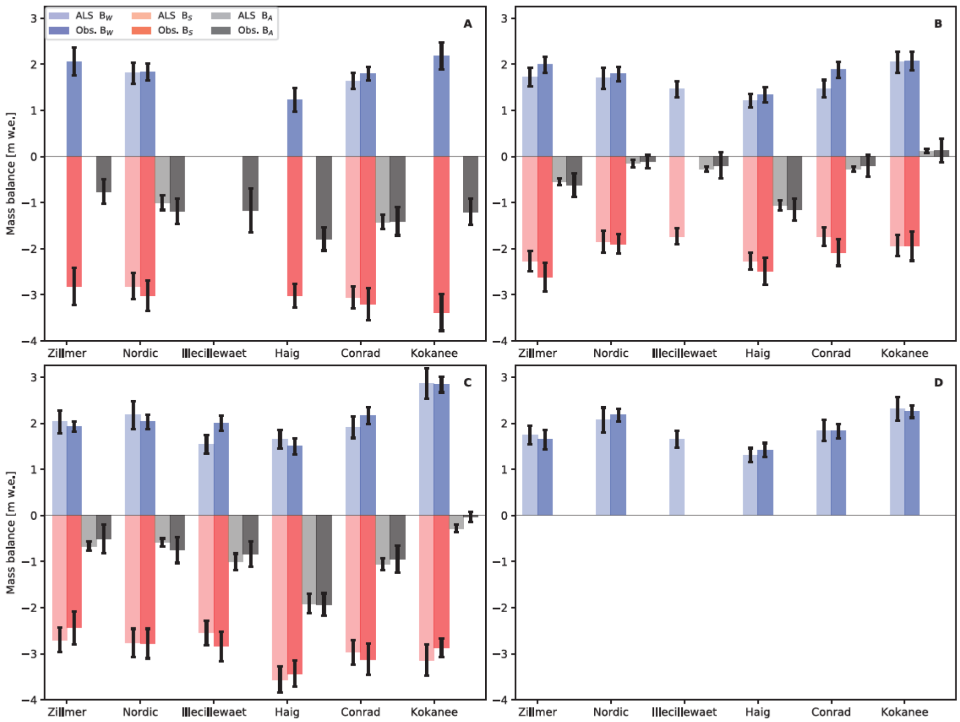
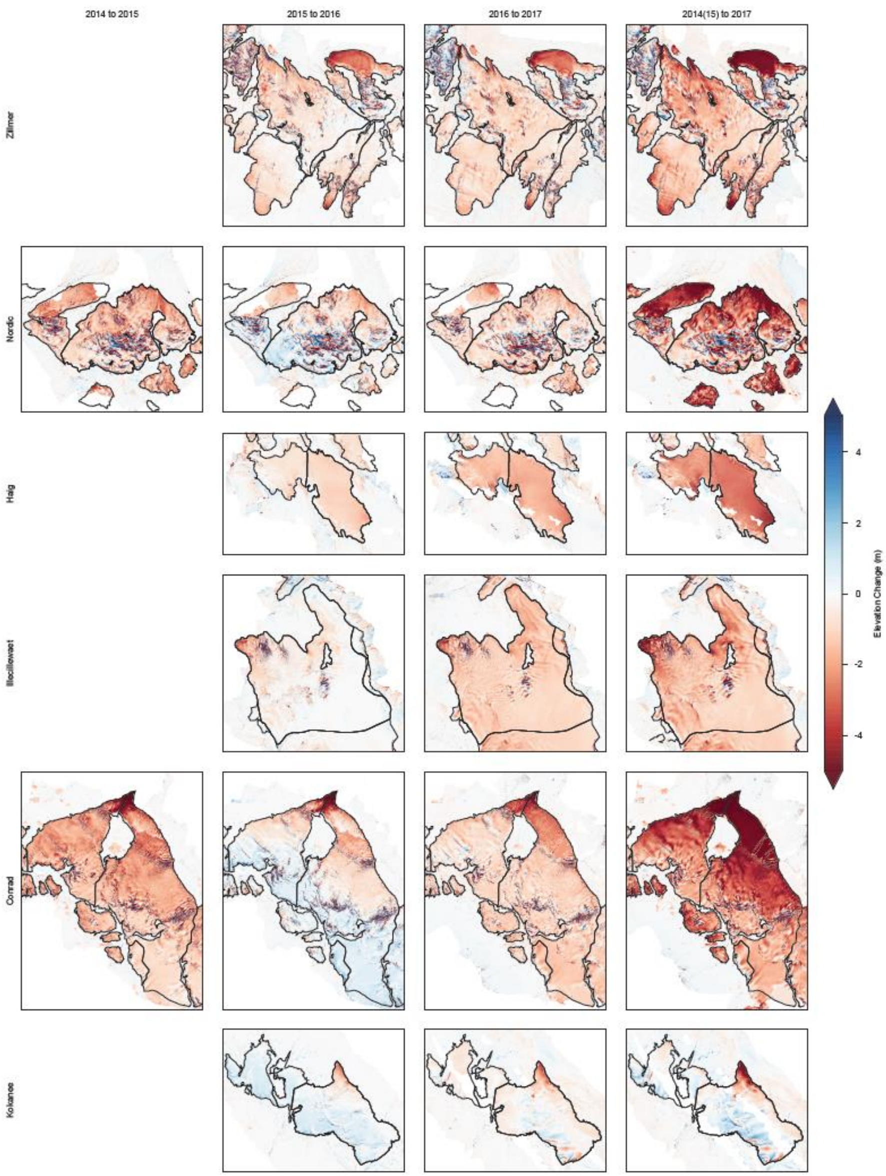


Figure 6: Seasonal and annual mass balance for all study glaciers from both geodetic and glaciological measurements for each balance year from 2014 to 2018 with 1σ uncertainties. A) 2014 to 2015 balance year, B) 2015 to 2016 balance year, C) 2016 to 2017 balance year, D) 2017 to 2018 winter balance.



2014 to 2015
2015 to 2016
2016 to 2017
2014(15) to 2017
Zillmer
Nordic
Haig
Illecillewaet
Conrad
Kokanee
4
2
0
-2
-4
Elevation Change (m)

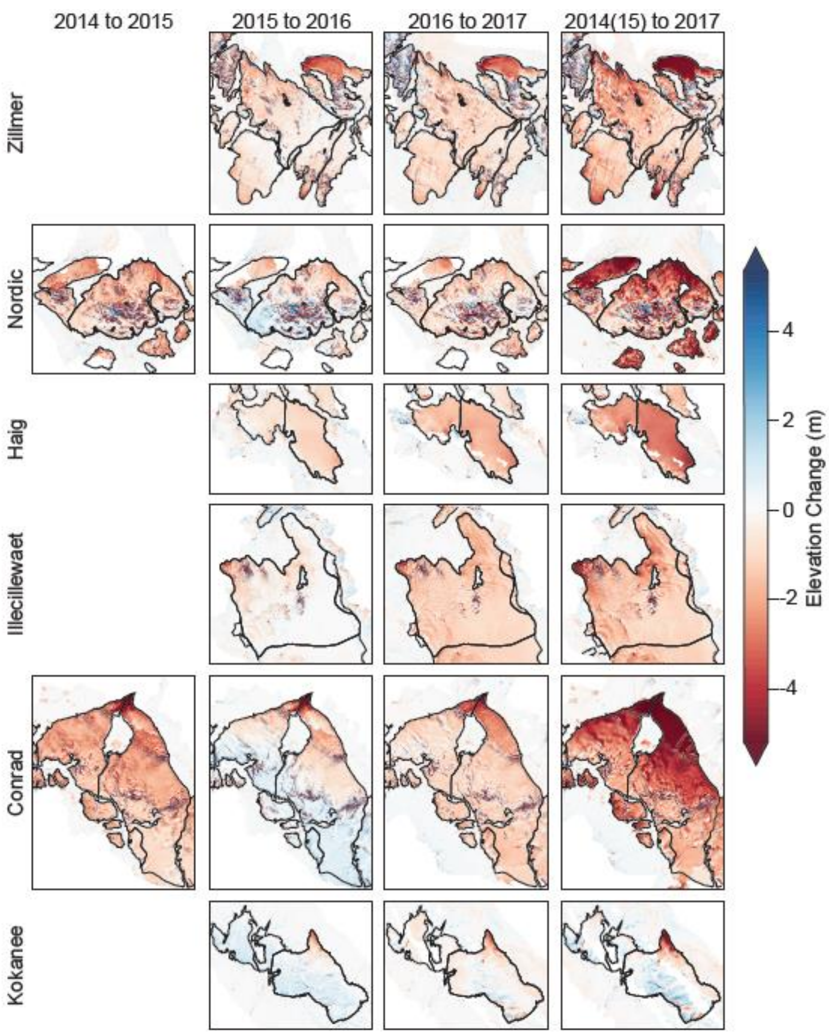


Figure 7: Surface height change for the Zillmer, Nordic, Haig, Illecillewaet, Conrad, and Kokanee glaciers from the first late~~-~~ summer DEM (2014 or 2015) until late~~-~~ summer 2017. Study glaciers are outlined with thick black line and other glaciers with a thin black line. Off-ice areas deemed stable terrain were used for error analysis and ~~co-registration~~coregistration.

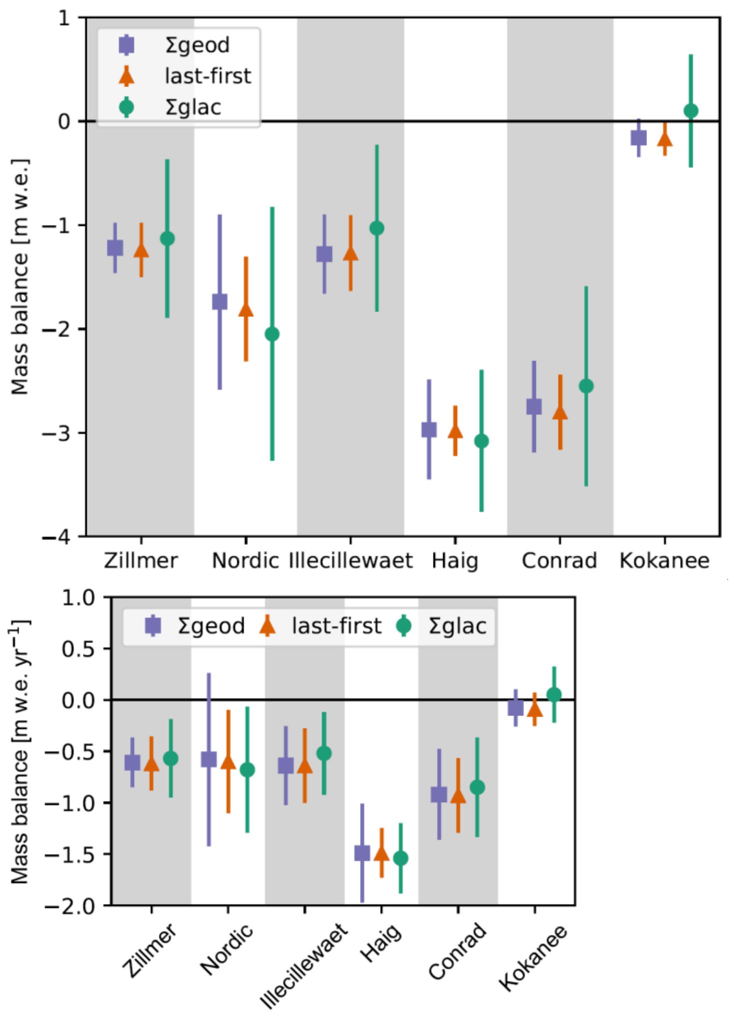


Figure 8: Summed annual mass balance from glaciological data (∑glac), geodetic data (∑geod), and last-first ΔDEM. Last-first ΔDEMs were created by differencing the first available DEM (2014 or 2015 late-summer) from the last available DEM (2017) for each site (Table 2). Errors denote 2-sigma uncertainties.