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RESEARCH AND OBSERVATORY CATCHMENTS: THE LEGACY AND THE FUTURE

WILEY

Long-term shifts in feedbacks among glacier surface change, melt generation and runoff, McMurdo Dry Valleys, Antarctica

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Abstract

Glaciers of the McMurdo dry valleys (MDVs) Antarctica are the main source of streamflow in this polar desert. Because summer air temperatures hover near 0°C small changes in the energy balance strongly affect meltwater generation. Here we demonstrate that increased surface roughness, which alters the turbulent transfer of energy between the ice surface and atmosphere, yields a detectable increase in meltwater runoff. At low elevations on the glaciers, basin-like topography became significantly rougher over 13 years between repeat lidar surveys, yielding greater melt. In contrast, the smoother ice at higher elevation exhibited no detectable change in roughness. We pose a conceptual model of the cycle of glacier surface change as a result of climate forcing whereby glacier surfaces transition from being dominated by sublimation to becoming increasingly melt-dominated, which is reversible under prolonged cool periods. This research advances our understanding of warm season effects on polar glaciers.

1 | INTRODUCTION

The cryosphere is particularly sensitive to increasing global temperatures, which have warmed considerably across the Arctic and high elevations and to a lesser degree in the Antarctic, although studies have found significant warming in the sub-Antarctic and West Antarctica (Adamson et al., 1988; Steig et al., 2009; Vaughan et al., 2003). The impacts of warming include rapidly changing landscapes and increasing mass loss of glaciers in the Arctic, mid-latitude ice caps and mountain ranges, and on the Antarctic Peninsula (Arendt et al., 2009; Gardner et al., 2011; Immerzel et al., 2010; Van Den Broeke et al., 2009; Willis et al., 2012; Wouters et al., 2015). As warming enhances ice melt, there are concurrent changes to the thermal and optical properties of the remaining ice. Understanding the impacts of melt, particularly feedbacks on additional melt and implications for energy balance properties of the cryosphere is an emerging and important area of focus (Goosse et al., 2018). In locations where warming and ice mass loss is not as intensive, such as the McMurdo dry valleys (MDVs) of Antarctica, we have yet to identify and quantify the mechanisms and results of warming-induced glacier change and runoff generation.

The glaciers of the MDVs supply the overwhelming majority of water for the ecosystem (Fountain et al., 1999). Meltwater is generated on glacier surfaces (upper 25 cm of ice) and because the glaciers are cold, the water remains on the glacier surface until it cascades as a waterfall off the 20 m tall glacier terminal cliffs. Glaciers are the headwaters for ephemeral proglacial streams, which feed permanently ice-covered, closed basin lakes (Fountain et al., 1999; McKnight et al., 1999). Solar radiation is the dominant component of the energy balance and in an average summer, melt is about 50% of ice mass loss from glacier surfaces, while the other 50% is sublimation and evaporation (Hoffman et al., 2014; Lewis et al., 1998). The average climate during the austral summers puts the MDVs on a threshold where a slight increase in temperature or solar radiation can shift the mass loss partitioning to melt from sublimation and greatly increase ablation and runoff rates (Fountain et al., 2014; Fountain, et al., 2016). Hoffman et al. (2014) modelled MDV glacier ablation and found that melt was 13% and sublimation was 86% of total ablation in a very cold season while melt increased to 59% and sublimation decreased to 41% of total ablation in a very warm year.

In exceptionally warm summers, high volumes of meltwater may thermally erode a glacier surface as runoff drains off the glacier (Fountain & Walder, 1998; Marston, 1983). This results in change to the overall glacier surface morphology, particularly the supraglacial drainage network (Jarosch & Gudmundsson, 2012). From an energy balance perspective, the drainage network creates centimetre to meter scale surface roughness. This small-scale roughness can trap wind-blown sediment, which lowers albedo (Bergstrom et al., 2020; Johnston et al., 2005; MacClune et al., 2003) increases total surface area available for melt, and creates an ice surface with multiple aspects (rather than a uniform sloping surface with a single aspect). Increased roughness overall can increase the potential for increased direct, rather than diffuse, radiation (Cathles et al., 2011; Rippin et al., 2015), and increases turbulent heat transfer into the ice (Hoffman et al., 2016). Incised channels and basins containing melt ponds have been modelled to produce 2 to 10 times more melt than adjacent flat surfaces on MDV glaciers (Hoffman et al., 2016). From a runoff routing perspective, a better developed supraglacial drainage network may be able to move water off the glacier more efficiently than a less developed network. Meltwater may be transferred from porous ice to a supraglacial channel faster in a well-developed network with higher drainage density than one with fewer channels, reducing the potential for evaporation and refreezing. We have yet to determine the relative importance of these mechanisms and how they interact.

A warm event that occurred in the summer of 2001–2002 caused major glacier melt and runoff to proglacial streams (Fountain, et al., 2016; Gooseff et al., 2017). This event was followed by additional high melt seasons and continued transfer of water from ice to streams, soils, and lakes (Levy et al., 2018). While thinning and terminus advance/retreat have been examined for the MDV glaciers (Fountain, Basagic, & Niebuhr, 2016; Levy et al., 2018), we have yet to fully quantify how climate variability and extreme warm events have impacted glacier melt, runoff, and glacier surface change.

We aim to deepen our understanding of how recent climate variability and warm events may impact glacier melt and runoff and how changes in surface roughness and the supraglacial drainage network may be a feedback process. To this end we ask: *Is there a shift in the relationship between ablation and runoff generation from MDV glaciers in response to warming?* and *Is there measurable MDV glacier surface change and how might it relate to runoff generation?* We are able to address these questions through the benefit of repeat airborne lidar surveys conducted in December of 2001 and 2014 and long-term records of streamflow and glacier mass loss. The first lidar data set was collected at the end of a decade of relatively cool seasons with low melt and runoff, while the second lidar data set was collected 13 years later, after an extreme warm event and a number or relatively warm, high melt and runoff seasons. The long term mass balance and runoff data cover both warm and cool periods allowing us to make a comparison of glacier surface change (roughness and channel development) and runoff generation before and during a disturbance regime. We take a watershed-based approach, determining the on-glacier contributing areas for each proglacial stream, facilitating direct comparisons of glacier surface change to runoff. We can use this knowledge of relatively short-term glacier change to understand the direction and magnitude in which the physical landscape may be evolving and to better predict impacts on the ecosystem as a result of long term climate warming and short term warm disturbance events. Our results identify key characteristics that can be examined across polar glaciers to understand cryospheric change and improve predictions of future melt.

2 | STUDY SITE, DATA AND METHODS

2.1 | Study site and glacier hydrology overview

The MDVs are the largest ice-free area of Antarctica (Levy, 2013). The landscape contains many cold-based alpine glaciers (Figure 1). Ephemeral proglacial streams host abundant algal mats that hydrate and become active when glacial meltwater is supplied. The streams feed lakes that host the only year-round active microbial communities and rely on glacial meltwater to balance evaporation and sublimation of water and ice-cover from the lake surface (Fountain et al., 1999). The majority of valley is dry, rocky, frozen soils. There is no significant groundwater source to streams; only shallow hyporheic zones exist in the active layer (<1 m). Average summer air temperatures are near 0°C with continuous solar radiation (~700 W m^{-2} at solar noon; Doran, McKay, et al., 2002).

The glaciers are −18°C at depth (average annual air temperature), and occasionally reach the melting point at the near surface (Fountain et al., 1999). This means that the glaciers are frozen to their beds and no meltwater is generated from or travels beneath the glaciers. Supraglacial watersheds can be conceptualized similar to traditional watersheds because runoff is generated in the top ~25 cm of ice and remains on the glacier surface flowing down gravitational potential gradients; there is little to no water lost to cracks and crevasses in the glacier. The ice shallow subsurface is analogous to the vadose zone and supraglacial channels are analogous to terrestrial stream channels. In a given watershed, many supraglacial channels flow parallel to each other off the glacier but all drain to a single proglacial stream that runs along the base of the glacier, eventually turning and draining toward a lake in the valley bottom or the ocean (Figures 1 and 2).

The glacial surfaces are characterized by a range of morphology, from very smooth ice to highly dissected basin topography (Figure 1; Hoffman et al., 2014; Johnston et al., 2005). Flat, smooth ice has millimetre to centimetre scale roughness and is stippled by cryoconite holes. Cryoconite holes are cylindrical depressions where concentrated patches of wind-deposited sediment preferentially melt into the surrounding ice, and in this system are almost always capped with an ice lid flush with the glacier surface (Fountain et al., 2004). Ice with 1st order supraglacial channels is intermediate between smooth ice and basin topography. Here, roughness is on the order of 0.01–1 m where small supraglacial channels 0.05–0.5 m wide, sometimes ice covered, begin to dissect flat ice. The channels are frequently closely spaced and run largely parallel toward the glacier edge. Basin topography is characterized by partly refrozen melt pools 5 to 50 m in

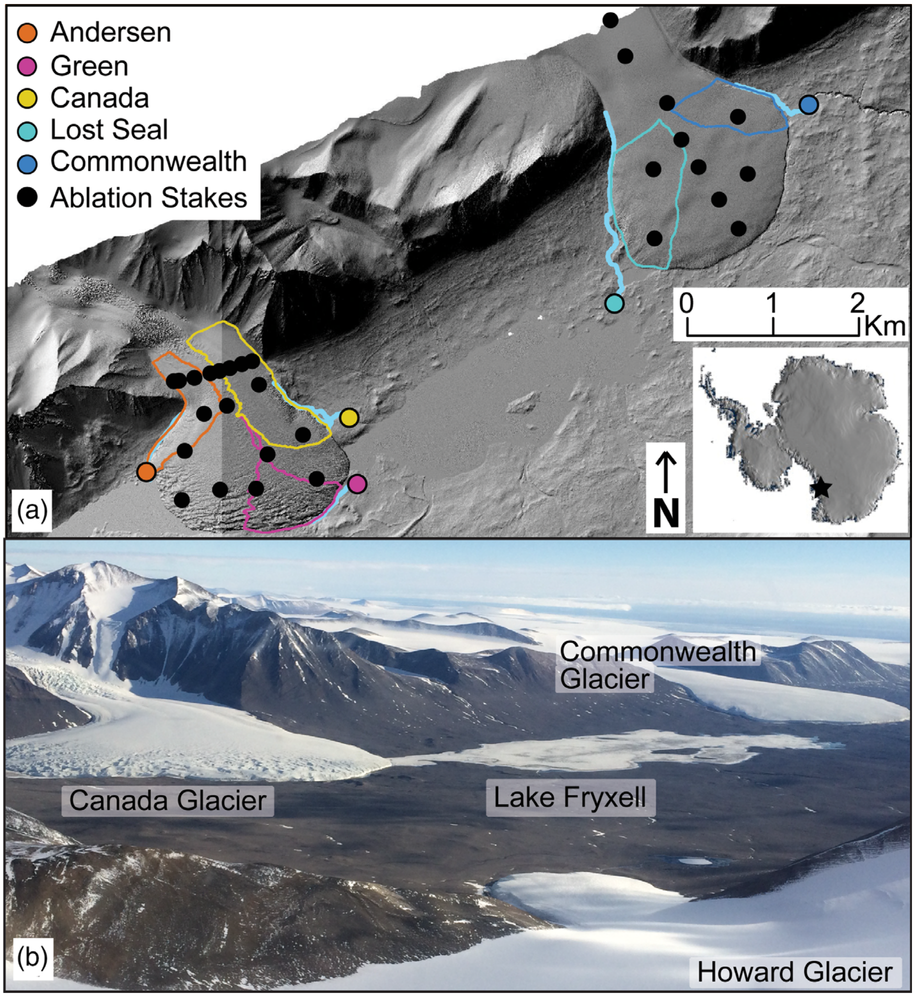

FIGURE 1 Canada and commonwealth glaciers and associated streams in the Taylor Valley (a). Black points in (a) are locations where glacier ablation measurements were made. The outlines on each glacier are stream contributing areas with stream gauge location indicated by a point of the same colour. Andersen (orange), green (pink), and Canada (yellow) streams drain Canada glacier. Lost seal (light blue) and commonwealth (dark blue) drain commonwealth glacier. (b) Photo of the lower half of Taylor Valley is taken looking north at Canada and commonwealth glaciers flowing south from the Asgard Mountains. The permanently ice-covered closed basin Lake Fryxell is between the two glaciers

length lined by a series of vertical steps and cliffs (0.5 to 15 m tall). Incised supraglacial streams connect the ponds as series of cascading pools (Bagshaw et al., 2010).

2.2 | Lidar data and geospatial analysis

To quantify glacier surface change, we use elevation data derived from repeat airborne lidar surveys of the MDVs. The first survey was collected in December 2001 with a resolution of 0.14–0.32 returns m^{-2}, and the second in December 2014, with at least 2 returns m^{-2} (Csatho et al., 2005; Fountain et al., 2017). The 2001 data were available as a 4 m^2-resolution digital elevation model (DEM) and the 2014 data as a 1m^2 DEM (Links to both data sets can be found through the Polar Geospatial Center: pgc.umn.edu/data/elevation). The 2014 survey mapped a much larger area than the 2001 survey, which is limited to valley bottoms and the ablation zones of glaciers. In order to make direct comparisons between data sets, we matched the 2014 DEM to the 2001 DEM so grid cells aligned, resampled the 2014 DEM to a 4 m^2 grid cell resolution, and clipped its extent to match the 2001 DEM.

We chose to focus on two glaciers in the Taylor Valley because they are included in both DEMs, have streams draining them with high-quality long-term discharge records and have long-term ablation measurements. Canada Glacier is drained by Andersen (−77.623375°S, 162.9064484°E), Green (−77.624099°S, 163.059750°E), and Canada (−77.613655°S, 163.054733°E) streams and Commonwealth Glacier is drained by Lost Seal (−77.595268°S, 163.244400°E) and Common wealth (−77.563332°S, 163.380828°E) streams (Figure 1). We used standard hydrological terrain analysis methods to delineate the glacial watersheds for each stream in this study using the open-source software package SAGA (System for Automated Geoscientific Analysis). First, we used a sink filling algorithm to eliminate local depressions. We then used the D-infinity flow routing algorithm (Band, 1986; Tarboton, 1997) to calculate upslope accumulated area for watershed delineation. The watershed was first delineated at the stream gauge for the entire watershed area, then clipped to the glacier because off-glacier area does not contribute any water and we are interested in on-glacier change. A cell was included within the watershed if a minimum of 25% of its area contributed to the watershed. Once included, then all meltwater generated in the cell is assumed to contribute to the watershed. The upper elevation of each watershed was bounded by the extent of the 2001 DEM (Figure 1). Catchment areas for runoff calculations were delineated from the original 1 m^2 2014 DEM because it covers the entirety of the ablation zones, the full runoff-generating

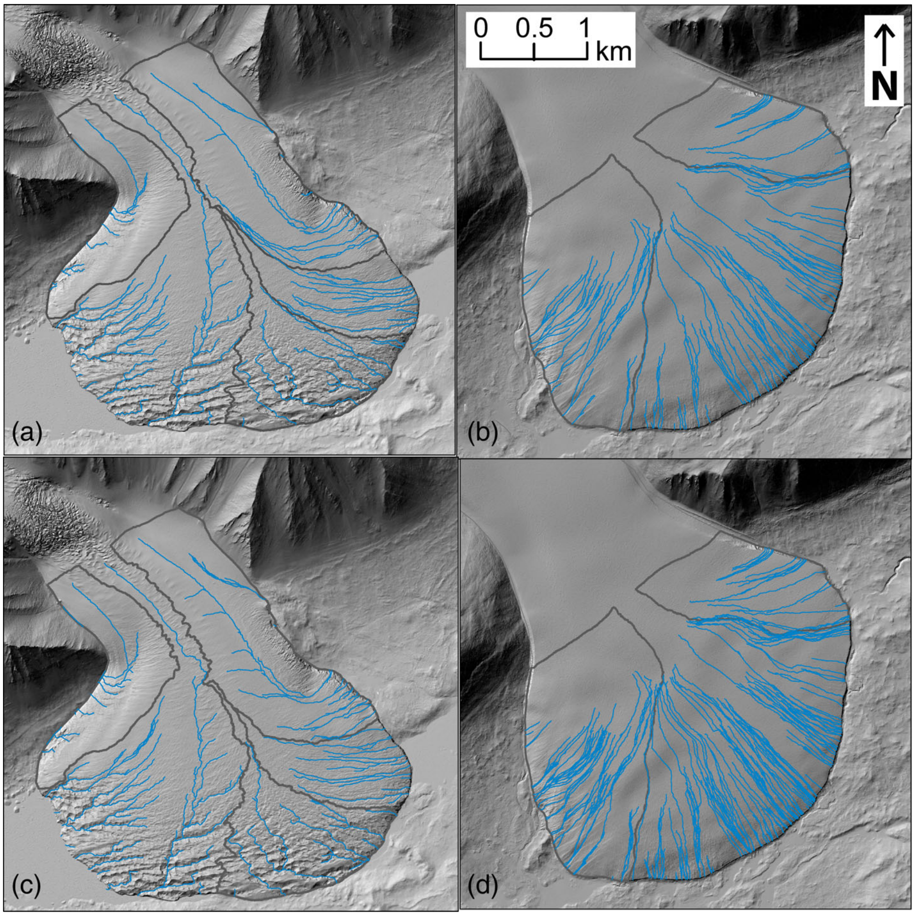

FIGURE 2 Maps of 2001 supraglacial channel networks on Canada (a) and Commonwealth (b) glaciers, and 2014 supraglacial channel networks on Canada (c) and Commonwealth (d) Glaciers. Channels are shown in blue with study stream contributing areas outlined in grey. All four maps have the same scale and orientation

area. Methods for watershed delineation using the 1 m^2 DEM are the same as described above for the coarser DEMs.

Supraglacial flow paths (stream channels) were identified for both the 2001 and 2014 DEMs using the upslope accumulated area data set described above and a channel initiation threshold (linear flow threshold) of 20 000 m^2. This threshold was chosen by comparing calculated channel networks to GPS points of four channel heads on Canada Glacier as well as by experience from extensive fieldwork on both glaciers. We calculated drainage density by summing all grid cells designated as stream cells and dividing that by the total watershed area, units: km/km^2.

To quantify surface roughness of both DEMs we calculated the terrain ruggedness index (TRI), the root-mean-square deviation of elevation surrounding a given grid cell (Riley et al., 1999) using Whitebox, an open-source GIS program. We binned TRI by 20 m elevation bands. We tested for significant differences in the median TRI using the Wilcoxsons Rank-Sum test with a significance level of 0.05. Because the uncertainty in elevation differences between the two DEMs is 0.25 m (Levy et al., 2018), differences <0.25 m were ignored.

Snow can confound the terrain analysis by accumulating in depressions, resulting in an apparent smoothing of the glacier surface (Eveland et al., 2012). The 2001 survey was largely snow-free but was present during the 2014 survey. We assessed the extent of snow on the glaciers by visually inspecting true-colour Landsat 8 imagery (LC82251292014353LGN01, 19 December 2014, 30 m resolution). Snow was visible on valley walls immediately adjacent to the upper elevations of the glacier ablation zones. It also appeared that some snow was present on the upper elevations of the glacier ablation zones, particularly on the east side of Commonwealth Glacier (in Commonwealth watershed, a location of persistent snow drifting). For the most part locations with snow were at elevations greater than the extent of the 2001 DEM and therefore excluded from this analysis. Snow and glacier ice have very similar reflective properties and it is therefore challenging to objectively mask snow covered locations (Levy et al., 2018). We instead use caution in interpreting results from the highest elevations of the ablation zone where snow may have confounded our roughness estimates.

2.3 | Time-series data

Complete glacier ablation and streamflow (*Q*) records begin in the 1993–1994 austral summer (Fountain et al., 2006; Wlostowski et al., 2016). Cumulative *Q* was calculated for each stream from the start of flow to the day that the end-of-season glacier surface mass balance (*M*) measurements were made (i.e., excluding additional flow for that season). We report all *Q* and *M* values normalized by catchment area: a depth of water per season (Figure 3).

Glacier mass loss estimates were calculated as the average of ice mass loss of all stakes in the glacier ablation zone (Figure 3). We used an ice density of 900 kg m^{-3} ignoring snow, which sublimates, not contributing to melt (Hoffman et al., 2014). We chose to use a mean of all stakes in the glacier ablation zone for annual mass loss because there are few stakes in any given watershed (Figure 1). Adding more stakes to our average provides a more representative estimate of mass loss across various slopes, aspects and elevations in a watershed and minimizes bias due to measurement error at any one stake.

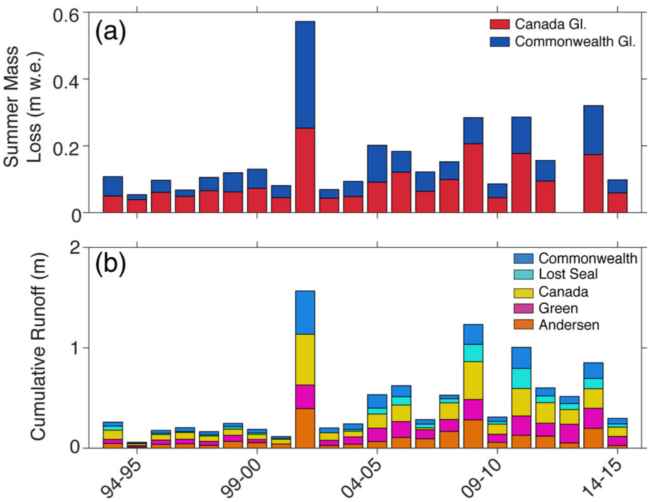

FIGURE 3 Records from 1993 to 2015 of (a) summer mass loss, area normalized to meters of water equivalent (m w.e.) and (b) runoff (normalized to contributing area) cumulative over the entire flow season. Andersen, Green, and Canada streams drain Canada glacier and lost seal and commonwealth streams drain Commonwealth Glacier. Summer mass loss was not measured for the 2012–2013 season and runoff data are missing from Green Creek in 00–01 and from lost seal stream from 2001 to 2003

To account for measurement uncertainty when comparing *Q:M* ratios, each reported streamflow value has an associated quality rating and uncertainty: good (10%), fair (25%), poor (30%). Additionally values were reported as missing with known flow (30%, assumed to be close to no flow at these times), or missing but known to be not flowing (0%). This streamflow error estimation is subjective based on the person developing the rating curve and performing quality control for a given season, however, we believe it is better to use available data rather than ignore potential error in the record. We calculated the proportion of the stream flow record each season with a quality rating and determined a weighted error for the entire season. *M* uncertainty is estimated at ±10% for bare ice and ± 20%–40% with snow (Fountain et al., 2006). We used a consistent 10% error for ablation measurements as we are ignoring snow in our stake measurements. We calculated a combined error for the ratios using the square root of the sum of squared errors.

Ablation stakes are located in more stable, flat glacier surfaces. This is a particularly important issue for Canada Glacier because the ablation measurements are biased towards smoother terrain. In order to compare ablation measured at stakes to ablation of the lower elevation basin areas, a repeat terrestrial lidar survey was conducted over 2 years in a basin on Canada Glacier. This basin is ~4500 m^2 and located at 77°37′41″S, 162°56′27″E, on the west side of the glacier, draining toward Lake Hoare. To compare the volume losses to the local stake measurements, specific volume (m) was calculated by dividing volume change by map-view area.

We used shortwave radiation, air temperature and wind speed data from the Lake Hoare meteorological station for the same period as *M* and *Q* records (Figure 4, Doran, McKay, et al., 2002; Doran and Fountain, 2019). Wind speed is calculated as an annual mean (1 July–30 June) and air temperature and solar radiation are calculated as a flow season mean (start to cessation of streamflow measured at Andersen Creek). We tested for significant differences of mean flow

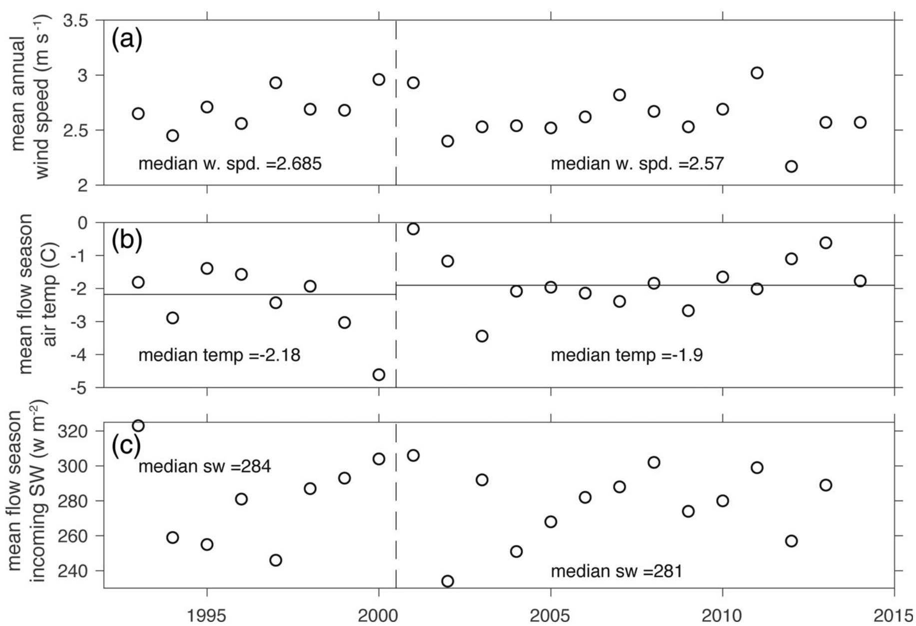

FIGURE 4 Climate data from 1993 to 2015 measured at the Lake Hoare meteorological station. Annual mean wind speed is measured from 1-Jun to 31-May (a). Air temperature and incoming shortwave radiation are averaged over the summer season from first to last measured streamflow (b–c). Dashed line indicates 2001, the time of LiDAR data collection and 'flood year'. Median seasonal air temperature was significantly different (*p*-value < 0.05) in seasons between 2001 and 2014 relative to the 2000–2001 season and before. No significant differences were found for wind speed and shortwave radiation

season air temperature and incoming shortwave radiation and annual mean wind speed in the seasons before and after the first lidar flight using the Wilcoxson rank sum test.

3 | RESULTS

3.1 | Hydrometeorological observations

To characterize the climatic conditions in the years leading up to the first lidar flight and between the two flights, we use mean annual wind speed as a proxy for sublimation and potential aeolian sediment deposition, and mean temperature and solar radiation during the summer season (Figure 4). No significant differences in wind speed and incoming short-wave radiation were detected, however flow-season mean air temperature increased significantly (Figure 4). The median flow season air temperature from 2001 to 2014 was 0.28°C higher than the median flow season air temperature from 1993 to 2000. In addition, *SD* of air temperature and solar radiation decreased from 1.0 to 0.8°C and 26.3 to 25.0 W m^{-2}, respectively.

Over the course of our long-term annual mass loss and streamflow records we observe an increase in *M* and *Q* for both glaciers, shifting around the 2001 flood year. Canada Glacier typically has higher *M* than Commonwealth Glacier. In the years that the TLS survey was conducted in the basin on lower Canada Glacier, average *M* measured from ablation stakes was −0.18 and − 0.09 m w.e. for the 2010–2011 and 2011–2012 seasons respectively. Basin ablation measured by TLS was −0.49 and − 0.22 m w.e., indicating that melt in rough areas was underestimated by ablation stake surveys by 172% and 144% in the 2010–2011 and 2011–2012 seasons.

To understand the relationship between glacier melt and runoff, we plot *Q:M* ratios over time. We found that after 2001, ratios are higher or more variable (Figure 5). Higher ratios suggest that more runoff is measured per unit mass loss on the glacier. Ratios measured at Green Creek were consistently below 1 (i.e., more ablation than runoff) prior to 2001 but shifted to consistently above 1 after 2001 (Figure 5b). Ratios less than 1 are expected considering mass loss due to sublimation and melt are subsumed in *M*. Ratios above 1 suggest there is melt not captured by ablation stake measurements, including melt from rough areas and terminal cliffs (Hoffman et al., 2016; Lewis et al., 1999). Ratios increase for Andersen and Canada watersheds, though the trend is less obvious than for Green Creek (Figure 5a,c). Ratios also increase for Commonwealth Glacier watersheds but more notably, variability increases substantially (Figure 5d,e).

Normalizing the *M* and *Q* data using the coefficient of variation (CV) shows concurrent increases of *M* and *Q* in Andersen and Canada streams but very little change in Green Creek *Q* (Table 1, Figure 6a). On Commonwealth Glacier, the large increases we observe in *M:Q* ratios are driven by both mass loss and runoff in Commonwealth Stream watershed but only mass loss in Lost Seal watershed (Figures 5, 6a, Table 1). The CV of *M* and *Q* increased for both glaciers and all streams, but the magnitude of the increase varied widely (Table 1, Figure 6a). We also calculated the median, skewness and CV of 15 min streamflow data each season and found that while the median increased after 2001, the skewness and CV showed no noticeable increase. This suggests that the intra-annual variability of *Q* does not change. Instead, flow simply increases, which is reflected in total annual Q increases (Figure 3b).

3.2 | Surface change

On both glaciers, the study watersheds cover a large part of the total glacier ablation zones, however there are other areas draining to ungauged streams and directly to Lakes Fryxell and Hoare from Canada Glacier (Figure 1). The ablation zone of Commonwealth Glacier is not entirely included in the 2001 DEM and therefore calculated watersheds may be smaller than their true extent. Watershed area changed from 2001 to 2014 (Table 1, Figure 6c), the largest being increases in Lost Seal and Commonwealth watersheds. In addition, drainage density increased in all watersheds except Green Creek, which had the highest drainage density in 2001 and 2014 (Table 1). Again, the largest drainage density increases were in Commonwealth and Lost Seal watersheds (Table 1, Figures 6b, 2b,d).

In each elevation band with few exceptions, the ice surface changed significantly from 2001 to 2014 (Figures 7,8). Upper elevations in the ablation zone of Canada Glacier show little change in roughness and above the 0.25 m threshold. In contrast, upper elevations on Commonwealth Glacier (120–280 m.a.s.l) show significant smoothing. Low elevations of both Canada and Commonwealth Glaciers, 40–180 and 60–120 m.a.s.l, respectively, were significantly rougher in 2014 than in 2001. These regions generally have the largest median TRI, indicating that rough terrain has become significantly more rough. Because the glaciers have a fan-shaped terminus, low elevations constitute a large proportion of their total ablation zone. Whereas only the lowest 3 of 13 elevation bands of Commonwealth Glacier had significant roughening in 2014, those bands comprise 53% of the total ablation area examined.

On Canada Glacier, both Andersen and Green watersheds show an overall roughness increase in (Figure 7b,c). Green Creek watershed exhibited significant increases across the most elevation bands (Figure 7c). In Andersen Creek watershed, 52% of the area significantly roughened at low elevation while 26% of the watershed significantly smoothed at upper elevations (Figure 7b). There were no significant TRI changes in lower elevations of Canada Stream watershed and no clear pattern of smoothing or roughening at upper elevations (Figure 7d).

Both watersheds on Commonwealth Glacier have a higher proportion of area that smoothed over time than roughened (Figure 7f,g). Lost Seal watershed generally smoothed by 2014, while Commonwealth Stream watershed significantly smoothed above 160 m.a.s.l. (46% of total watershed area) but significantly roughened below 160 m.a.s.l. (42% of the total area). (Figure 7g). However, smoothing within the Commonwealth Stream watershed may be due to snow present during the 2014 survey. Roughness changes outside the

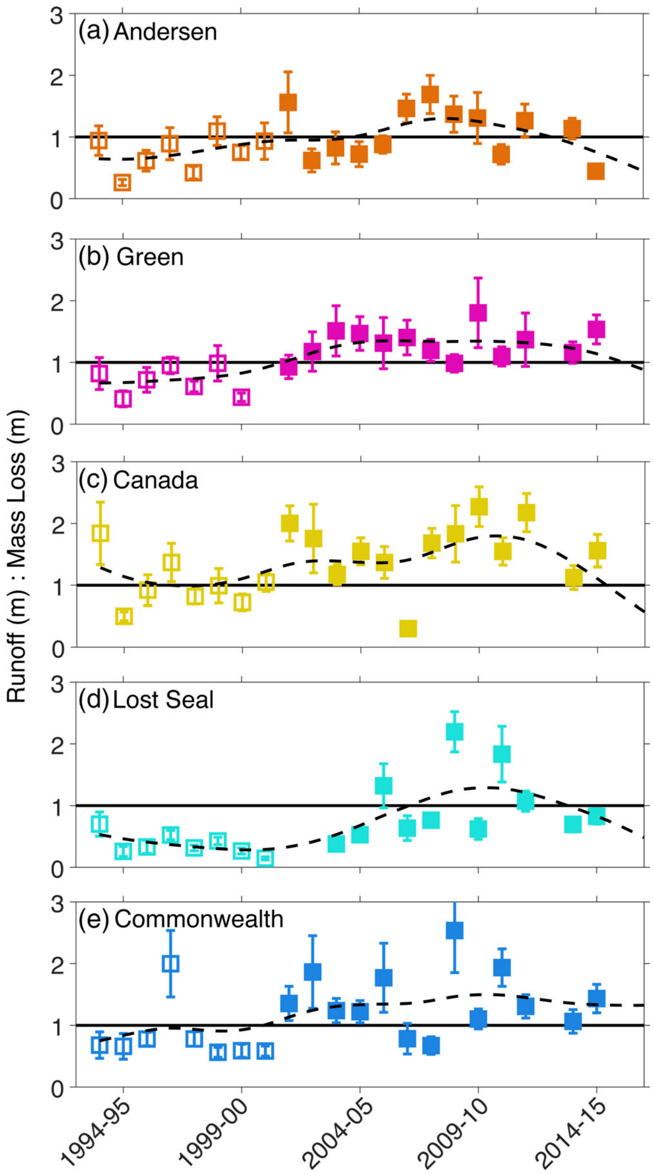

FIGURE 5 Ratios of runoff to glacier mass loss by season from 1993 to 2015 for all study streams draining Canada (a–c) and Commonwealth Glaciers (d,e). Open squares are data before the first lidar flight in 2001 and filled squares are data between the first and second flight from 2001 to 2014, with error bars representing a combined error across runoff and mass loss. The dashed line in each plot is a spline fit of the timeseries of ratios

watersheds were also analyzed and indicate significant roughening at low elevations of both glaciers (Figure 8).

4 | DISCUSSION

4.1 | Recent warm melt seasons

We can determine recent climate patterns from high-quality records collected in the MDVs and nearby weather stations (Fountain, Basagic, & Niebuhr, 2016; Obryk et al., 2020). Over the period of record in the Taylor Valley, there was a decadal cooling trend from 1993 to 2000 (Doran, Priscu, et al., 2002) followed by an extreme warm event (Fountain, Saba, et al., 2016) and over a decade of warmer years from 2002 to 2015 (Figure 4b, Gooseff et al., 2017). Our analysis finds these differences are apparent in air temperature but not solar radiation (Figure 4b,c). The switching from a slightly cooler to slightly warmer decade is reflected in glacier melt and streamflow (Figure 3). Low temperatures result in very low melt seasons in the 1990s (Figure 3a), positive annual mass balances measured on Commonwealth and Howard glaciers (Figure 1b, Fountain, Basagic, et al., 2016), and low runoff in proglacial streams (Figure 3b). After 2001, glacier melt and runoff is much higher almost consistently across every season (Figure 3). We have found evidence of increased *M*, *Q*, and glacier surface roughness over a 13-year period, but is there a connection between MDV glacier surface change and runoff generation? We pose and explore two hypotheses that are broadly applicable to polar glaciers where the majority of meltwater remains on the glacier surface.

4.2 | Interactions between melt, runoff, and glacier surface change

4.2.1 | The energy balance hypothesis

In this polar environment, rougher ice surface shifts the energy balance towards more melt in surface depressions. This creates a positive feedback such that the bottom of the depressions ablate faster than the ridges, producing rougher surfaces and more melt.

Previous studies have shown that a positive feedback exists in rough surface topography whereby it alters the micrometeorology to favour melt over sublimation (Hoffman et al., 2014, 2016; Johnston et al., 2005; Lewis et al., 1998). As proportionally more melt is produced, there is more energy available for thermal erosion of the ice surface (Marston, 1983), deepening and enlarging melt ponds and stream channels. Given that a rougher surface produces more melt than a flatter surface for a given amount of incoming energy, we would expect that surface roughening produces both more glacier melt and more streamflow. Thus, an increase in *M* and *Q* would support this hypothesis. Alternatively, additional melt and runoff could be produced due to a more positive energy balance if the glacier simply receives more incoming energy independent of glacier surface changes. In that case we could observe increased *M* and *Q* with or without glacier surface change but increased air temperature and/or solar radiation.

4.2.2 | The runoff efficiency hypothesis

A more extensive supraglacial stream drainage network will increase glacier runoff per unit energy input compared to a less extensive network by conveying water faster off the ice surface, reducing residence times, evaporative loses and refreezing.

TABLE 1 Comparing coefficient of variation of mass loss (Canada and commonwealth glaciers) and runoff (Andersen, Cr., Green Cr., Canada St., Lost Seal St., and Commonwealth St.) and the difference in the coefficient of variation for seasons before versus after 2001

	Coefficient of variation			Drainage density			Contributing area		
	93–94 to 00–01	01–02 to 14–15	Difference	2001	2014	Difference	2001	2014	% change
Canada Gl.	0.20	0.58	0.38						
Andersen Cr.	0.39	0.81	0.42	1.35	1.55	0.20	1.22	1.20	−1.67
Green Cr.	0.32	0.40	0.08	2.64	2.64	0.00	1.43	1.35	−5.93
Canada St.	0.34	0.70	0.37	2.41	2.60	0.19	2.04	2.06	0.97
Commonwealth Gl.	0.40	0.84	0.44						
Lost Seal St.	0.72	0.74	0.02	1.03	1.50	0.47	2.27	2.42	6.02
Commonwealth St.	0.31	0.86	0.55	0.68	1.35	0.67	1.02	1.19	14.29

Note: Drainage density and contributing area are calculated using the 2001 and 2014 DEMs.

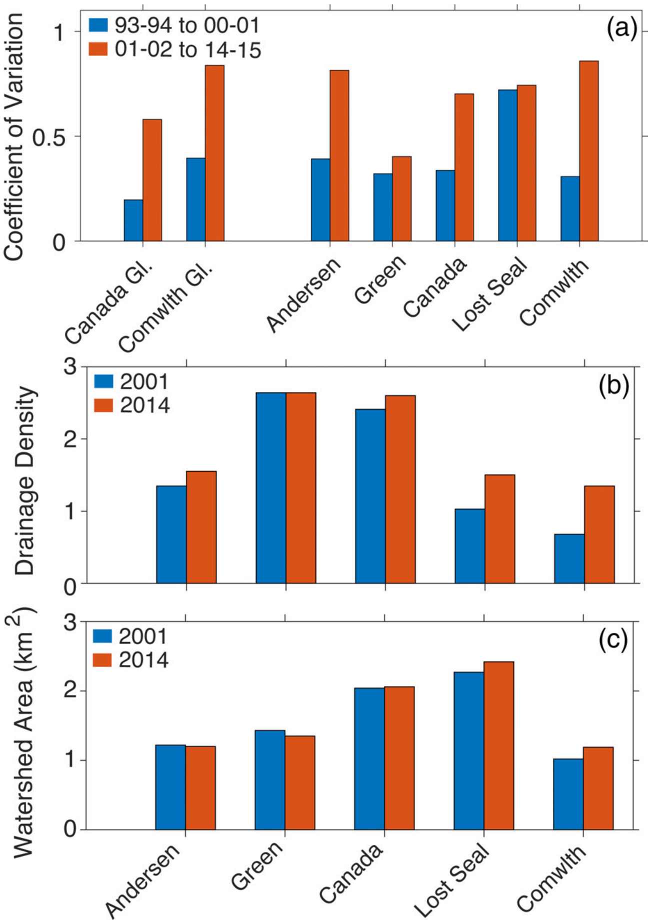

FIGURE 6 (a) Comparison of the coefficient of variation of mass loss (cm w.e.) for Canada and Commonwealth Glaciers and for runoff measured in the three streams draining Canada glacier (Andersen, Green, and Canada) and two streams draining Commonwealth Glacier (Lost Seal and Commonwealth) for years prior to the first lidar data collection (1993–1994 to 2000–2001) and between the first and second lidar data collection (2001–2002 to 2014–15). (b) Comparison of drainage density (km per km^2) of stream channels within each watershed and (c) watershed area calculated from the lidar DEMs collected in 2001 and 2014

In terrestrial systems, as drainage density increases, runoff and event peaks increase as storage and travel time decrease (D'Odorico & Rigon, 2003). Travel times decrease because water spends less time travelling through the porous medium (soil, fractured bedrock) or via overland flow and reaches a stream more quickly, where it travels at higher velocity as open channel flow with smaller frictional losses. In supraglacial stream networks, faster travel times from 'hillslope' to stream mean a lower likelihood of refreezing in the porous ice and more efficient transfer of water from glaciers to proglacial streams. Under this hypothesis we would observe higher streamflow per unit mass loss on the glaciers, an increase in the *Q*:*M* ratio.

4.2.3 | Evaluating hypotheses—Inextricably linked?

Our analysis of glacier surface change indicates that across both Canada and Commonwealth Glaciers, there is an increase in roughness and drainage density (Figures 2, 6,7,8). Ratios of *Q*:*M* appear to increase somewhat in all watersheds (Table 1), lending some support for the runoff efficiency hypothesis. However, the most striking difference in mass loss and runoff prior to and after the flood year is the increase in interannual variability (Figures 5, 6a, Table 1). In a cooler season, there will rarely be sufficient energy to produce melt, meaning low *Q* and more *M* occurring as sublimation regardless of glacier surface roughness. A season with increased incoming energy can shift to conditions favourable for melt. Rougher topography both lowers the threshold for melt (lower albedo and changing micrometeorology) and creates a positive feedback resulting in very high *M* and *Q* with more *M* as melt than sublimation. After a change in surface roughness, there may be cooler, low *Q* seasons where even increased surface roughness cannot help overcome melt thresholds. However, slightly warmer seasons will have much higher *Q* as a result of the positive feedback, resulting in high inter-annual variability. Therefore, we need to consider not only the magnitudes and correspondence of *M* and *Q* but also the CV of *M* and *Q*. We generally observe a concurrent increase in CV of mass loss and runoff, which suggests that the energy balance hypothesis is dominant (Figures 5 and 6a, Table 1). When austral

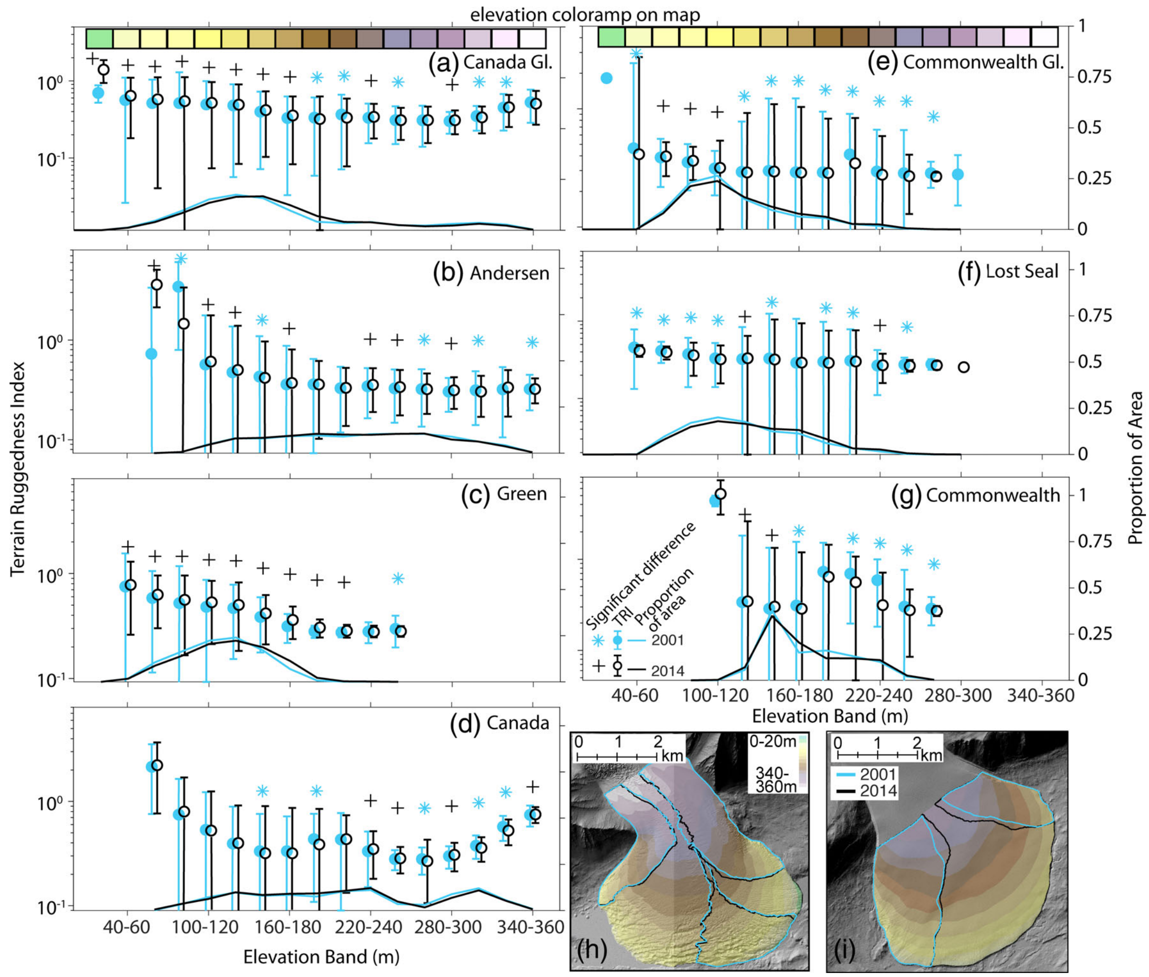

FIGURE 7 Comparisons of the median (point) and one standard deviation (lines) of terrain ruggedness index (TRI) calculated for 20 m elevation bands of the 2001 and 2014 DEMs on the entire Canada glacier (a), its three watersheds (b-d), the entire Commonwealth Glacier (e) and its two watersheds (e,f). Symbols over medians indicate significant differences between 2001 and 2014 TRI ($p < 0.05$). A blue asterisk denotes the 2001 DEM had a significantly higher TRI and a black plus indicates that the 2014 DEM had a significantly higher TRI for that elevation band. Lines represent the relative proportion of area in that elevation band in the 2001 (blue) and 2014 (black) DEMs, right y-axis. Maps depict 20 m elevation bands and stream contributing area boundaries calculated using the 2001 and 2014 DEMs on the Canada (h) and commonwealth (i) glacier

summers have sufficient energy to produce melt, a disproportionately high amount of glacier mass loss and streamflow is generated, resulting in high mass loss and runoff CVs.

Hoffman et al., (2016) modelled the energy balance on these glaciers, which underpredicted melt after 2005. They suggest this may due to an alteration of the glacier physical characteristics which could not be incorporated into their model due to lack of data. We found that TRI is highest at low elevations in both 2001 and 2014. This is a reflection of the negative correlation between elevation and glacier melt due to adiabatic lapse rates (Fountain et al., 2006; Fountain, Basagic, & Niebuhr, 2016). We observe significant roughening broadly across the lower half of Canada Glacier making up 73% of the total area, and across a 60 m elevation band of Commonwealth Glacier, making up 53% of the total area (Figure 7). These increases appear to be driven by the areas outside the watersheds we consider here (as well as within Green and Commonwealth stream watersheds, Figures 7, 8a,b). These two pieces of evidence, surface roughening and increased Q and M CV, point toward the possibility that a change to the physical properties of the glaciers are bringing them closer to the threshold of melt in summer rather than climate being the primary driver of the energy balance shift.

While we find more evidence supporting the energy balance hypothesis, we still see slight increases in mass loss to runoff ratios. The watersheds where $Q{:}M$ increased the most are places that already had areas of high surface roughness, namely Green Creek (Figure 5). We have evidence from the TLS survey that these basins produce over twice as much melt than what is measured by ablation stakes, which is representative of typical morphology of lower Canada Glacier, and agrees with modelling efforts (Hoffman et al., 2016). We then must consider mass loss estimates more of a relative measure for the lower elevations rather than an absolute measure. This suggests that increasing $Q{:}M$ may be the result of glacier melt not captured by ablation stakes rather than a true increase in $Q{:}M$.

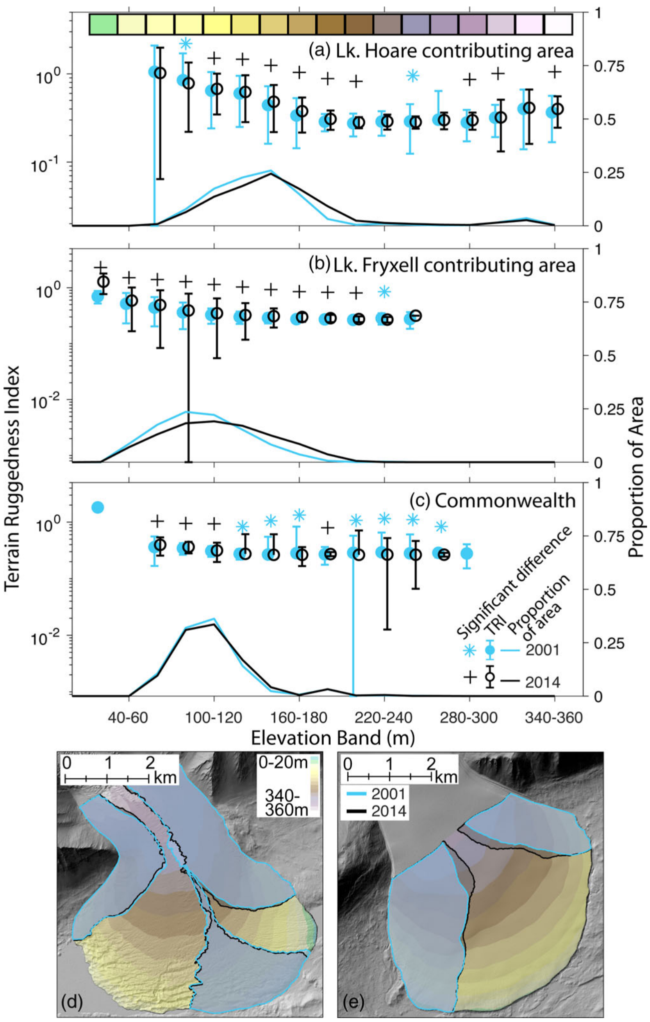

FIGURE 8 Comparisons of the median (point) and one standard deviation (lines) of terrain ruggedness index (TRI) calculated for 20 m elevation bands of the 2001 and 2014 DEMs on the glacier surface outside the study contributing areas. Canada Glacier is divided into the area that contributes water to Lake Hoare to the west between Andersen and Green creeks (a), and the area that contributes water to Lake Fryxell, between Green Creek and Canada stream (b). There is one large area between Lost Seal And Commonwealth streams on Commonwealth Glacier (c). Symbols over medians indicate significant differences between 2001 and 2014 TRI ($p < 0.05$). A blue asterisk denotes the 2001 DEM had a significantly higher TRI and a grey plus indicates that the 2014 DEM had a significantly higher TRI for that elevation band. Bars are the relative proportion of area in that elevation band in the 2001 (blue) and 2014 (grey) DEMs. Maps depict 20 m elevation bands and stream contributing area boundaries for reference on Canada (d) and commonwealth (e) glacier

We suggest that rather than increased drainage density being the cause of increased drainage efficiency and higher runoff, it is instead an indicator of the glacier adjusting to past warm events. In terrestrial systems, an increase in drainage density is a reflection of increased catchment wetness (Collins & Bras, 2010; Rinaldo et al., 1995), which translated to glaciers, is an increase in meltwater and the ability to thermally erode and head-cut channels (Jarosch & Gudmundsson, 2012). Our calculated drainage density values are surprisingly comparable with what have been observed on the Greenland Ice Sheet (0.9 to 4.7 km km^{-2}, Yang et al., 2016). The entire channel network is not always active on MDV glaciers but is evidence that the drainage networks recently and regularly convey high volumes of water (as they did in the flood year).

In the Green Creek watershed, TRI (glacier surface roughness) increased significantly at almost all elevation bands, but drainage density did not. Additionally, the CV of runoff was much lower than every other watershed except Lost Seal. Green Creek already had a well-established supraglacial stream network with channel features more analogous to glaciers with higher discharge (Irvine-Fynn et al., 2011; Marston, 1983; Rippin et al., 2015). Karlstrom and Yang (2016) suggest that on the Greenland Ice Sheet, channels thermally erode toward a topographic steady state. Our finding that Green Creek drainage density was initially high and remained constant over 10+ years suggests that parts of MDV glaciers receiving the most meltwater may also be approaching a topographic steady state. While the extent of the channel network in Green Creek remained relatively static, the increased meltwater may have further incised streams and basins, increasing TRI. An additional contributing factor for why we may not see large changes in *Q* CV is storage in melt ponds at low elevations of the glaciers (Bagshaw et al., 2010), which are present in the Green Creek watershed.

Ultimately, while we pose two hypotheses in order to systematically interrogate the mechanisms causing increased runoff due to glacier surface change, we believe that the energy balance and supraglacial drainage network development are inextricably linked and likely feedback on each other. High melt rates thermally erode the glacier surface increasing surface roughness and creating a more developed, higher-density drainage network. The watersheds we examine here are good examples of watershed structure and function along a continuum of increasing melt. More parallel channels, lower roughness and higher *Q:M* variability are indicative of a watershed that has less frequent energy balance conditions favourable for melt (i.e., Commonwealth watershed). A more temporally stable dendritic network, with high and increasing roughness and low *Q:M* variability is indicative of a location that frequently has an energy balance with favourable melt conditions (i.e., Green Creek). We combine these hypotheses and how they fit into a continuum of surface change and runoff generation in a conceptual model below (Figure 9).

4.3 | Conceptual model of glacier surface change, meltwater generation and runoff

Surface roughness is a major driver of meltwater generation magnitude on polar glaciers. In the trough of a rough surface, particularly in basin areas at lower elevation, wind speeds are slowed, drastically reducing the loss of energy back to the atmosphere and favouring melt over sublimation (Figure 9a; Lewis et al., 1998; Johnston et al., 2005; Hoffman et al., 2008, 2016) Furthermore, the angle of basin walls are such that solar radiation reflected off one wall may be directed toward another, thus increasing the total amount of radiation penetrating ice in the basins (Cathles et al., 2011). Finally, basin areas

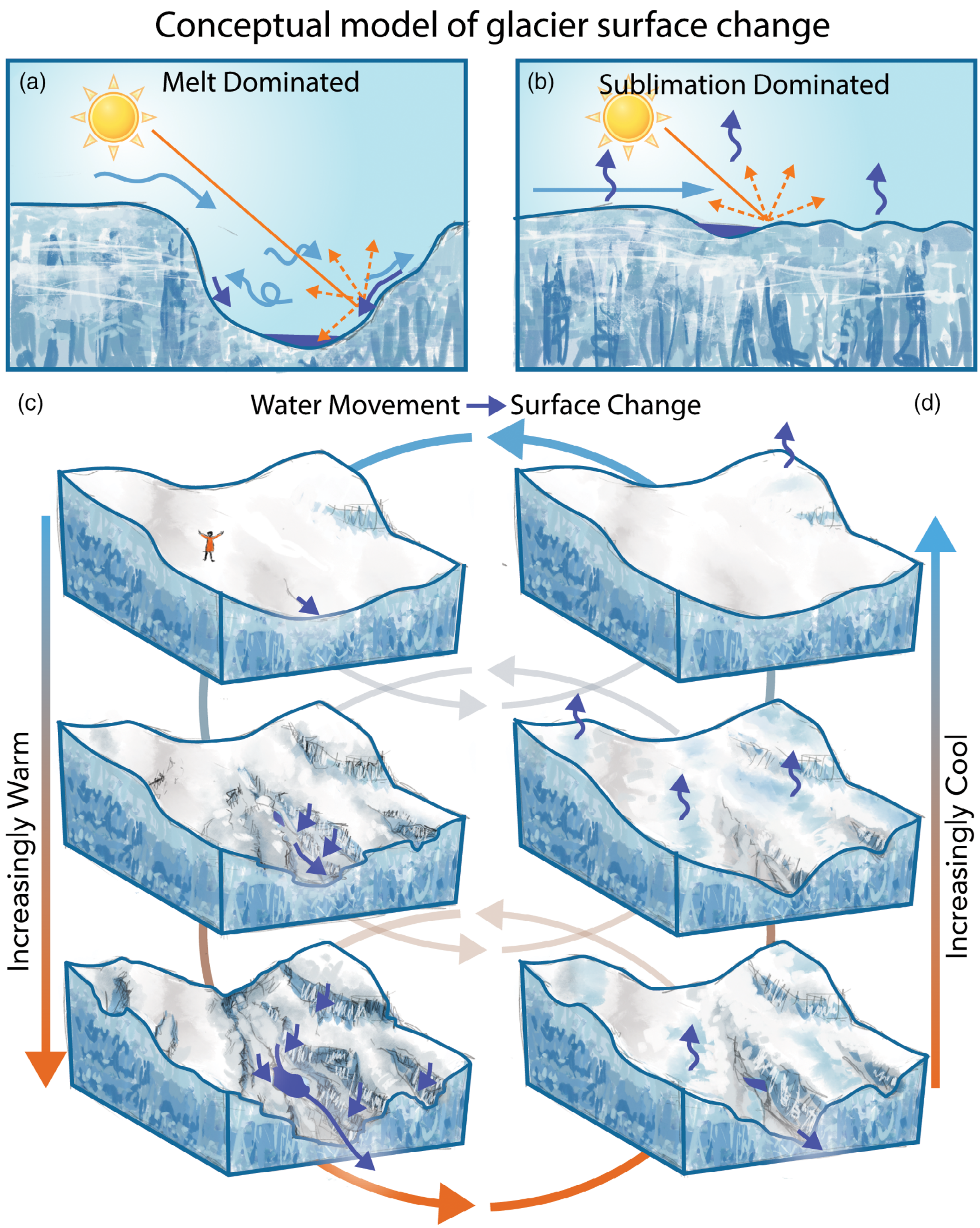

FIGURE 9 Conceptual model of glacier surface change under different climatic conditions. Surface roughness exacerbates melt by reflecting energy into other parts of the ice surface and slows wind in the depressions (a). While on flat surfaces, the reflection of shortwave radiation and higher wind speeds favour sublimation over melt (b). Under increasingly warm conditions, we expect that more energy will produce increasing meltwater generation (c). This additional meltwater will thermally erode the glacier surface, increasing surface roughness and supraglacial channel density. Under increasingly cool conditions, sublimation will dominate and supraglacial streams will freeze over (d). Eventually sublimation will strip the top layers of ice, particularly on ridges, and the ice surface will smooth. We suggest this process is cyclical but can be 'short-circuited' at any point, and occurs at a range of timescales

generally have lower albedo due to the excess sediment trapped in basins (Bergstrom et al., 2020; Johnston et al., 2005; MacClune et al., 2003) and the darker ice created by supraglacial channels and ponds (Bagshaw et al., 2010; Rippin et al., 2015). In contrast on smooth surfaces, wind speeds are not slowed and more energy is removed via turbulent heat fluxes resulting in higher sublimation rates (Hoffman et al., 2008; Figure 9b). Furthermore, when meltwater is generated on flat ice surfaces, lower slope and lack of an existing drainage network means that water cannot not drain as quickly. This increases the likelihood of refreeze or evaporation rather than drainage off the glacier to proglacial streams (e.g., Arnold et al., 1998). While melt is generated in sediment-laden cryoconite holes in smooth areas, cryoconite holes frequently remain isolated, contributing little to total runoff volumes (Fountain et al., 2004, 2008).

This differential melt magnitude as a function of surface roughness and energy balance creates a cycle of surface morphological change. During warm, sunny periods, enough meltwater can be generated on flat ice surfaces to promote the formation of supraglacial channels (Figure 9c; Lampkin & Vanderberg, 2014). As melt continues, these channels will thermally erode the glacier surface and down cut (Karlstrom & Yang, 2016; Marston, 1983), beginning to enhance surface roughness (Figure 9c) analogous to the development of parallel rills on a hillslope (Figure 2, Horton, 1945). As channels grow and drain increasingly large areas, flows will increase (St Germain & Moorman, 2019). This can result in the development of a more dendritic network, meandering and the formation of supraglacial ponds in depressions (Figure 9c; Horton, 1945; Fountain & Walder, 1998; Bagshaw et al., 2010; Karlstrom & Yang, 2016; Pitcher & Smith, 2019; St Germain & Moorman, 2019). Once surface roughness is established via channelization of flow, solar radiation can enhance the backward melting of channel walls, adding additional melt to streams (Cathles et al., 2011; St Germain & Moorman, 2019).

In cooler periods, there is less energy available for melt and ablation that does occur is dominated by sublimation (Hoffman et al., 2008). This means that flow will slow or cease in supraglacial channels (Figure 9d). If colder periods are sustained for sufficiently long periods of time and sublimation continues, the glacier surface may smooth, reducing surface roughness (Figure 9d).

We present this conceptual model of glacier surface evolution as cyclical. While it is introduced as a full cycle (e.g., Figure 9c,d) we want to emphasize that during a summer season, conditions can frequently switch from cooler to warmer or vice versa and for example the glacier surface can jump from conditions depicted in the middle of Figure 9c to the middle of Figure 9d. Additionally, the timescale over which this occurs can be variable and strongly depends on the energy balance. To date, timescales of surface morphological evolution are not well constrained. In this study we are limited to two snapshots in time 13 years apart. Melt can erode the surface more quickly than sublimation can smooth it, and therefore surface change should progress more quickly along the continuum depicted in Figure 9c than depicted in Figure 9d. In winter, mass loss is almost entirely via sublimation which removes the low-density ice, resetting to a fresh high-density ice surface at the beginning of the next summer season (Hoffman et al., 2016). The balance between summer melting and winter sublimation is an additional factor affecting the glacier surface morphology, and should be interrogated with further field data collection and modelling efforts.

5 | CONCLUSION AND IMPLICATIONS

Landscape change due to melt is widespread across polar regions and the McMurdo Dry Valleys of Antarctica are no exception (Chinn, 1980; Levy et al., 2018). Fountain et al. (2014) suggested that the MDVs are on the threshold of change and identified the glacier ablation zones as part of the most ‘at risk’ areas. Levy et al. (2018) suggest that major melt and land surface alteration occurs episodically during 1–2 week warm events in the austral summers. This is also likely the case for MDV glaciers. The large melt event that occurred in the 2001–2002 austral summer was a catalyst for change across the landscape,

We expect that if warm events continue, distribution of water in the landscape will continue to shift from stored frozen in glaciers to stored as liquid in lakes, with higher water flux in streams serving as the connection between the two. Glacier runoff is critically important for the MDV ecosystem. The supply of meltwater supports the most productive locations in the valleys. However very high flows from excessive melt events can scour algal mats and sediment in proglacial streams and deliver high volumes of freshwater to lakes (Gooseff et al., 2017). Additionally, Canada Glacier has area that contributes directly to Lakes Hoare and Fryxell which have roughened significantly meaning both lakes are directly receiving increased amounts of fresh glacier meltwater with no proglacial stream processing, contributing to recent drastic lake level increase and impacting chemoclines, lake chemistry and ecological dynamics (Gooseff et al., 2017).

These findings would not have been possible without consistent long-term measurements of glacier mass loss and streamflow and repeat lidar flights, indicating the value of long term study areas with intensive data collection. This research takes a unique approach by considering the watersheds on the glaciers themselves contributing to each stream. We found evidence supporting the hypothesis that increased meltwater generation is partly due to positive feedbacks between the energy balance and glacier surface roughness. While increased drainage density may also be contributing by increasing drainage efficiency, the evidence is not as clear and confounded by underestimating melt on glacier surfaces with high roughness. We suggest that surface change is a relatively underexplored parameter affecting the energy balance of glaciers and ice sheets. We believe that more careful consideration and accounting of surface roughness and energy balance feedbacks can improve model performance and predictive capability of melt across the cryosphere.

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DATA AVAILABILITY STATEMENT

Long term data sets are available on MCM LTER database. Streamflow data are available at: http://mcm.lternet.edu/streams-data-sets. Glacier ablation data are available at: http://mcm.lternet.edu/Glacier-data-sets. Lake Hoare meteorological data are available at: http://mcm.lternet.edu/meteorology-data-sets. Lidar DEMs can be found at the USGS US Antarctic Resource Center: usarc.usgs.gov and from the National Center for Airborne Laser Mapping (NCALM) at opentopography.org.

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