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Alpine Glaciers: An Introduction

1.1 Glacier Observation Programs

Glaciers have been studied as sensitive indicators of climate for more than a century (Forel, 1895; Zemp *et al.*, 2015). Glacier fluctuations in terminus position, mass balance, and area are recognized as one of the most reliable indicators of climate change (Haeberli, Cihlar, and Barry, 2000; Oerlemans, 2005). The recognition of glacier sensitivity to climate led to the development of a global reporting system for glacier terminus change and glacier mass balance during the International Geophysical Year (IGY). Today, this system is managed by the World Glacier Monitoring Service (WGMS). WGMS annually collects standardized observations on changes in mass, volume, area, and length of glaciers with time. This data on individual glacier fluctuations has been enhanced and supplemented in recent years by glacier inventories derived from satellite imagery. Glacier fluctuation and inventory data are today high-priority key variables in climate system monitoring (Sharp *et al.*, 2015; Pelto, 2015b) (Fig. 1.1).

Observations of alpine glaciers most commonly focus on changes in terminus behavior, to identify glacier response to climate changes (Forel, 1895). A number of nations have long-running annual terminus survey programs: Austria, Italy, Switzerland, Norway, and Iceland (WGMS, 2012). The data set of terminus change compiled by the WGMS has 42,000 measurements on 2000 glaciers (Zemp *et al.*, 2015).

Annual mass balance measurements are the most accurate indicator of short-term glacier response to climate change (Haeberli, Cihlar, and Barry, 2000; Zemp, Hoelzle, and Haeberli, 2009). Annual mass balance is the change in mass of a glacier during a year resulting from the difference between net accumulation and net ablation. The importance of monitoring glacier mass balance was recognized during the IGY in 1957. For the IGY, a number of benchmark glaciers around the world were chosen where mass balance would be monitored. This network continued by the WGMS has proven valuable with a total of annual glacier observations; from 1985 to 2014, the average number of glaciers reporting annual mass balance has been approximately 100. Thirty-seven of these are considered reference glaciers with at least a continuous 30-year record of mass balance.

In addition, the advent of frequent high-resolution satellite imagery has allowed for the completion of global mountain glacier inventories led by the Global Land Ice Measurements from Space (GLIMS) and the Randolph Glacier Inventory (RGI) (Arendt *et al.*, 2012; Pfeffer *et al.*, 2014). Detailed repeated inventories have developed a standard approach and also identified changes through time (Kääb *et al.*, 2002; Fischer *et al.*, 2014; Radić and Hock, 2014). The inventories focus on using standard methodologies to define glacier outlines and glacier attributes. Typical attributes include area, length, slope, aspect, terminal environment, elevation range, and shape classification. Satellite images can also be used to map transient snowlines (TSLs), the snowline separating the ablation and accumulation zone

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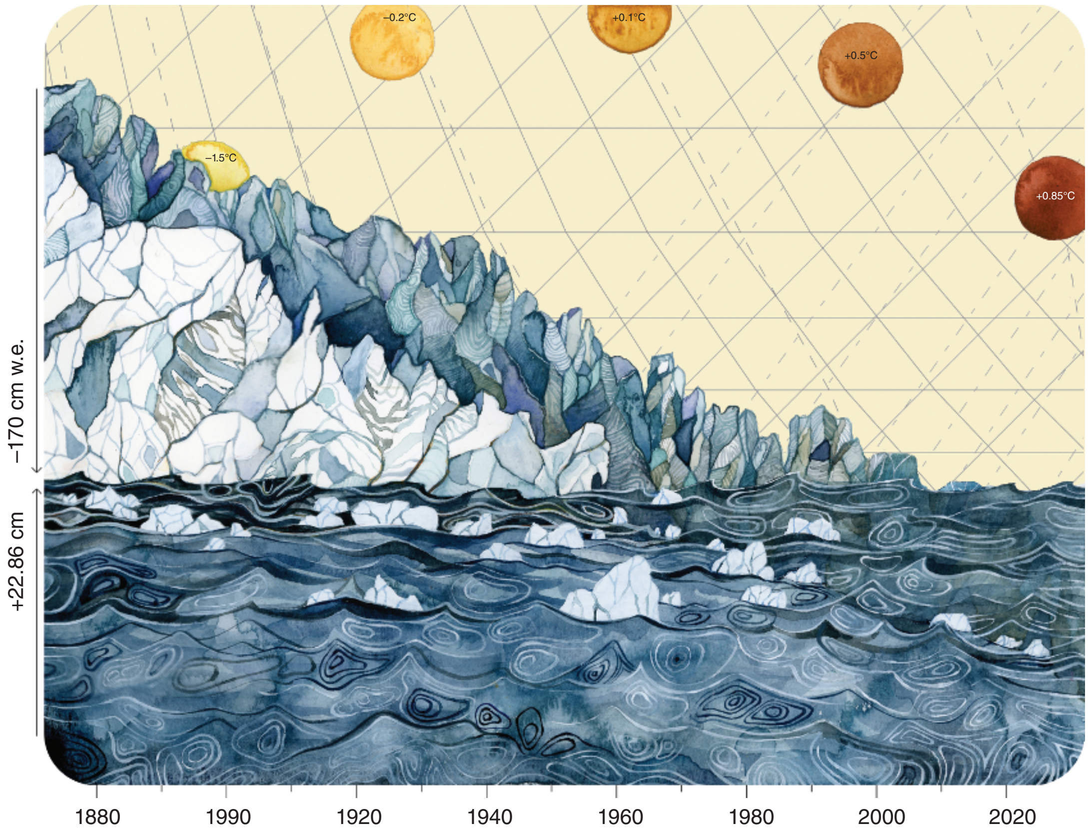

Figure 1.1 Decrease in Glacier Mass Balance based on data from Pelto (2015a) to illustrate the dramatic decline in North Cascades glaciers, WA. (Watercolor, Jill Pelto 2015).

during the summer; the end of summer TSL represents the equilibrium line altitude (ELA) on most alpine glaciers that lack superimposed ice formation (Østrem, 1975; Mernild *et al.,* 2013). These data provide baseline information for an assessment of glacier changes.

The geodetic inventories assess glacier area and in many cases glacier volume. ICESat and other instruments provide elevation change data to compliment areal extent change assessment (Neckel *et al.,* 2014). The remote sensing geodetic inventories and the field glaciological observations both indicate that rates of early twenty-first-century mass loss are historically unprecedented at global scale (Zemp *et al.*, 2015). The largest negative mass balances have occurred in one of the last two decades, depending on the region (WGMS, 2013). The decadal mean annual mass balance was −221 mm in the 1980s, −389 mm in the 1990s, and −726 mm for 2000s. The continued large negative annual balances reported indicate that glaciers are not approaching equilibrium (Pelto, 2010). The strong negative mass balance suggests that glaciers of many regions are committed to further volume loss even under current climatic conditions (Zemp *et al.*, 2015). Radić and Hock, (2014) indicate that future climate change will enhance the mass losses substantially.

The RGI version 3.2 was completed in 2014, compiling digital outlines of glaciers, excluding the ice sheets using satellite imageries from 1999 to 2010. The inventory identified 198,000 glaciers, with a total extent estimated at 726,800 ± 34,000 km^2 (Pfeffer *et al.,* 2014). An earlier RGI 2.0 has been used to estimate global alpine glacier volume at ~150,000 Gt (Radić and Hock, 2014), quantifying the important role as a water resource and potential contributor to sea-level rise.

This information on glacier mass balance and terminus change has been collected and made available from internationally coordinated efforts (WGMS, 2011, 2013). GLIMS and the RGI have made available their glacier inventory data as well (Arendt *et al.,* 2012; Pfeffer *et al.,* 2014). This is a wealth of information on the state of glaciers.

1.2 Importance of Mountain Glaciers

Mountain glaciers are important as water resources for agriculture, hydropower, aquatic life, and basic water supply (Schaner *et al.,* 2012; Bliss, Hock, and Radić, 2014). Alpine glaciers in many areas of the world are important for water resources – melting in the summer when precipitation is lowest and water demand from society is largest. The timing and magnitude of glacier melt are sensitive to climate change; hence, rational water resource management depends on understanding future changes in water resources from glaciated mountain ranges (Immerzeel, Beek van, and Bierkens, 2010).

Mountain glaciers have also contributed to sea-level rise (Radić *et al.,* 2013; Marzeion, Jarosch, and Hofer, 2012). The annual contribution has been approximately 1 mm a^{-1} during the twentieth century since (Marzeion, Jarosch, and Hofer, 2012). Mountain glaciers can also increase local natural hazards such as glacial lake outburst floods (Bajracharya and Mool, 2009).

1.3 Glacier Terminus Response to Climate Change

In this book, we examine glacier responses during the 1985–2015 period, with the primary climate change being the global temperature rise since 1976 (GISTEMP Team, 2015). Changes in mass balance control a glacier's long-term behavior. Terminus and glacier area changes are then impacted with a lag time for both an initial and more complete response.

For any glacier, there is a lag time (*Ts*) between a significant climate change and the initial observed terminus response (Paterson, 1994); this is also referred to as the reaction time of the glacier. It should be noted that *Ts* cannot be considered a physical property of a glacier and is expected to depend on the mass balance history and physical characteristics of the glacier.

In addition, for each glacier there is a response time to approach a new steady state for a given climate-driven mass balance change (*Tm*), referred to as length of memory by Johannesson, Raymond, and Waddington (1989). They defined *Tm* as the timescale for exponential asymptotic approach to a final steady state (approximately 63% of a full response), resulting from a sudden change in climate to a new constant climate. The magnitudes of *Ts* and *Tm* are crucial to interpreting past and current glacier fluctuations and predicting future changes (Paterson, 1994; Johannesson, Raymond, and Waddington, 1989).

For glaciers in the North Cascades, Washington, Pelto and Hedlund (2001) found a *Ts* of 10–20 years and a *Tm* of 20–100 years.

1.3.1 Equilibrium Response

Typically, glacier terminus retreat results in the loss of the lowest elevation region of the glacier. Since higher elevations are cooler than lower elevations, the disappearance of the lowest portion of the glacier reduces overall ablation, thereby increasing mass balance and potentially reestablishing equilibrium (Pelto, 2010). Typically, a glacier's thinning is greatest at the terminus, and at some distance above the terminus; usually in the accumulation zone, the glacier is no longer thinning appreciably even during retreat (Schwitter and Raymond, 1993). This behavior of greatest thinning at the terminus and limited thinning in the accumulation zone suggests a glacier that will retreat to a new stable position (Schwitter and Raymond, 1993).

A period of sustained positive mass balance will lead to an increase in glacier thickness, an increase in velocity, and eventually an advance. The advance expands the area of the glacier at the lowest elevations where mass balance is more negative. When the expansion at low elevation is sufficient to offset the increased mass balance, the retreat will end as equilibrium is reached.

1.3.2 Disequilibrium Response

In recent years, an increasing number of glaciers have been identified to be experiencing a disequilibrium response to climate (Pelto, 2010; Carturan *et al.,* 2013). There is no point to which such a glacier can retreat to reach equilibrium and the glacier would then disappear. For alpine glaciers, typically 50–70% of the glacier must retain snow cover even at the conclusion of the melt season to be in equilibrium, this is referred to as the accumulation area ratio (AAR). Without a substantial consistent accumulation area, a glacier cannot survive. If a nonsurging alpine glacier is experiencing extensive thinning and marginal retreat in the accumulation zone of the glacier, it lacks a persistent accumulation (Pelto, 2010). The result is a more unstable form of retreat with substantial thinning throughout the length and breadth of the glacier. A glacier in this condition is unlikely to be able to survive in anything similar to its present extent given the current climate. This is evident in satellite or aerial photographs of glaciers. The emergence of bedrock outcrops or the recession of the upper margins of a glacier is the key symptoms to observe (Kääb *et al.,* 2002; Pelto, 2010; Carturan *et al.,* 2013). Glaciers disappearing and fragmenting have become a common reporting category of glacier inventories, both indicating disequilibrium with climate (Tennant *et al.*, 2012; Kulkarni and Karyakarte, 2014; Carturan *et al.,* 2013).

1.3.3 Accumulation Zone Changes

Paul *et al.* (2004) utilized satellite imagery to identify nonuniform changes in glacier geometry, emerging rock outcrops, disintegration, and tributary separation to determine collapse versus a dynamic (equilibrium) response to climate change. It has become practical to examine the terminus and areal extent change of all glaciers in the region (Bajracharya and Mool, 2009; Paul *et al.,* 2004). This does quantify the extent of the retreat, but not the nature of the equilibrium or disequilibrium response. Identifying disequilibrium requires identification of significant thinning in the accumulation zone, which will occur if a persistent accumulation zone is lacking. Accumulation zone thinning is evidenced by the emergence of rock outcrops in the accumulation zone, changes in the accumulation zone margin, and reduced crevassing (Paul *et al.,* 2004; Pelto, 2010).

1.3.4 Terminus Response Factors

Glacier terminus response to climate depends on several dynamic features primarily: the existence of debris cover, having a calving terminus, and being a surging glacier. These dynamic features do not render a glacier insensitive to climate change in the long run, but can mitigate or accentuate the response in the short run.

Debris cover is common in three of the regions examined in this volume: Patagonia, New Zealand, and Himalayas. Once the debris cover in the ablation zone exceeds a thickness of 1 cm, it insulates the glacier below from the atmosphere, reducing ablation rates (Brock *et al.,* 2010). This leads to a slower thinning and less negative mass balances (Scherler, Bookhagen, and Strecker, 2011). In the long run, debris-covered glaciers exhibit the same response to climate – just delayed and subdued.

Calving glaciers lose mass by the calving of icebergs into a lake or ocean. A glacier that terminates in water leads to enhanced melting of the glacier front, which typically leads to greater calving. Calving increases with the ratio of ice thickness to water depth. The greater this ratio, the less the buoyancy of the calving front and the less the calving rate. A greater ratio would also lead to less ablation from contact with the water. The key parameter identified by Benn, Warren, and Mottram (2007) is the variation of longitudinal strain rate, which determines the crevasse depth near the calving front. Calving typically enhances retreat as it is another mechanism to lose volume. If there is a change in water depth at the calving front, and calving is reduced or increased, the glacier sensitivity to climate is reduced (Benn, Warren, and Mottram, 2007).

Surging glaciers experience periodic periods of slow and rapid flows. The periods of rapid flow are much shorter in duration than the slow-flow periods. During the periods of slow flow, a glacier thickens, which promotes higher velocity and also leads to hydrologic changes at the glacier bed that is crucial to generating a period of rapid flow. During the slow phase, a glacier's response to climate change is often enhanced, while during the rapid-flow periods a surging glacier is less sensitive to climate conditions.

1.4 Glacier Runoff

Glaciers act as natural reservoirs storing water in a frozen state instead of behind a dam. Alpine Watersheds are comprised of rainfall-dominated (pluvial), snowmelt-dominated (nival), and glacier melt-dominated segments. Glacially fed streams peak during late summer, July and August in the Northern Hemisphere, and January and February in the Southern Hemisphere, during peak glacier melt (Fig. 1.2) (Fountain and Tangborn, 1985; Dery *et al.,* 2009). The loss of glaciers from a watershed would result in reduced streamflow primarily during late summer minimum flow periods (Pelto, 2008; Nolin *et al.*, 2010; Stahl and Moore, 2006).

1.5 Climate Change and Impact of Runoff

Glacier runoff is the product of glacier area and glacier melt rate. If the percent decline in glacier area is greater than the percent increase in melt rate, runoff declines. If melt rate percentage increase is greater than the percentage of area loss, glacier runoff will increase. The point at which glacier runoff begins to decline due to area loss is considered the peak glacier runoff. In some areas, peak runoff has already occurred. Bliss, Hock, and Radić (2014) observed that of 18 glaciated alpine areas, most regions exhibit a fairly steady decline in runoff, demonstrating that they have passed their peak runoff. Moore *et al.* (2009) found significant declines in stream discharge from glaciers in British Columbia during late summer. Pelto (2011) found the Skykomish River glacier runoff had peaked by 2006. In the Swiss Alps, peak runoff overall has not occurred (Huss, 2011). Isaak *et al.* (2011) have found significant and ubiquitous warming of streams in the Pacific Northwest during the summer in unregulated rivers from 1980 to 2009 of 0.12 °C/decade. The combination of reduced flow and temperature rise will be an ongoing threat to aquatic life (Grah and Beaulieu, 2013).

The impact from the Nooksack River, WA, serves as an example (Pelto, 2015a). Discharge and water temperature are measured in each river. The South Fork has 0% glacier cover, Middle Fork 2.1%

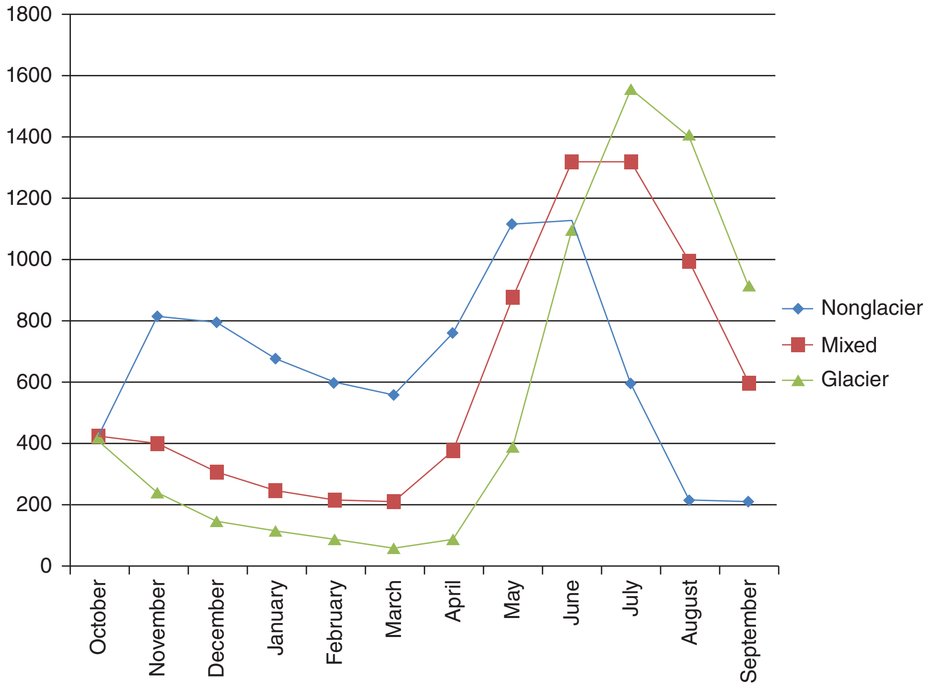

Figure 1.2 Comparison of hydrographs for nonglacier- and glacier-fed watersheds.

glacier cover, and North Fork 6.1% glacier cover, allowing differentiation of glacier impact. In addition, streamflow is measured directly below one glacier in the North Fork, Sholes Glacier. To distinguish the different discharge and thermal responses, the comparison was made during the most stressful period – late summer warm weather events. For the 2009–2013 period, 12 warm weather events were identified. The mean increase in air temperature during the warm weather events from prior to their beginning was 7 °C. Warm weather events consistently generate a significant increase in stream water temperature only in the nonglaciated South Fork Basin; the mean increase was 3.2 °C (Table 1.1). Increased glacier discharge largely offset the impact of increased air temperature on stream water temperature during the warm weather events, leading to a mean change of 1.1 °C in the North Fork and 1.0 °C in the Middle Fork.

Table 1.1 Mean response of Nooksack River watershed to the 14 warm weather events from 2009 to 2013.

Basin	Air temperature change (°C)	Stream temperature change (°C)	Stream discharge change (%)
South Fork Nooksack	+8	+3.4	−15
North Fork Nooksack	+8	+1.1	+23
Middle Fork Nooksack	+8	+1.0	+16

The change in air temperature is assessed as the rise in the mean daily temperature at the Middle Fork Nooksack SNOTEL site.

For discharge during the same warm weather events, a 15% increase is set as the key threshold for a significant response to each warm weather event. This threshold was chosen as only significant rain or melt events generate this large a change in daily flow. For the North Fork, 11 of 12 warm weather events exceeded this limit, in the Middle Fork 8 of 12 events had a significant response, and for the South Fork none of the 12 events led to a 15% flow increase. The average discharge change for the warm weather events are +26% in the North Fork, +19 % in the Middle Fork, and −16% in the South Fork (Table 4.3). It is apparent that warm weather events increase glacier melt enhancing flow in the North Fork, and in a basin without glacier runoff, South Fork, the hydrologic system consistently experiences reduced discharge.

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