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Reconstructing Climate from Glaciers

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Abstract

Glaciers offer the potential to reconstruct past climate over timescales from decades to millennia. They are found on nearly every continent, and at the Last Glacial Maximum, glaciers were larger in all regions on Earth. The physics of glacier-climate interaction are relatively well understood, and glacier models can be used to reconstruct past climate from geological evidence of past glacier extent. This can lead to significant insights regarding past, present, and future climate. For example, glacier modeling has demonstrated that the near-ubiquitous global pattern of glacier retreat during the last few centuries resulted from a global-scale climate warming of ~1°C, consistent with instrumental data and climate proxy records. Climate reconstructions from glaciers have also demonstrated that the tropics were colder at the Last Glacial Maximum than was originally inferred from sea surface temperature reconstructions. Future efforts to reconstruct climate from glaciers may provide new constraints on climate sensitivity to CO_2 forcing, polar amplification of climate change, and more.

1. INTRODUCTION

1.1. Glaciers: From Scientific Curiosity to Icons of Climate Change

Cosmogenic nuclide dating: provides the exposure age of a land surface using nuclides that accumulate through time from galactic cosmic ray bombardment

Surge-type glacier: a glacier that exhibits periods of episodic, abnormally fast flow, which is sometimes accompanied by terminus advance

Tidewater glacier: a glacier that terminates in the sea, with the terminus either floating or grounded below sea level

> *The sea, or rather the vast river of ice, wound among its dependent mountains, whose aerial summits hung over its recesses. Their icy and glittering peaks shone in the sunlight over the clouds. My heart, which was before sorrowful, now swelled with something like joy.*
>
> —Description of the Mer de Glace in Mary Shelley's *Frankenstein* (1831, p. 82)

Knowledge of glaciers (see the sidebar titled Glaciers) dates back to prehistory across a wide range of cultures. Glaciers have served as sources of knowledge and inspiration in art and literature as well as in oral histories (Cruikshank 2001). The earliest known drawing of a glacier (Vernagtferner in Austria) dates to AD 1601, and drawings and photographs of glaciers became common thereafter. During the Enlightenment, glaciers were accurately described, illustrated, and photographed, especially in the European Alps and Scandinavia (**Figures 1** and **2**). As clear visualizations of changing climate that can be appreciated by both scientists and the public, historical glacier positions are an important resource for reconstructing past climate change as well as contextualizing recent global warming.

Using records of past glacier fluctuations to improve the paleoclimate record has the potential to advance our understanding of the mechanisms of past, present, and future climate change. As a relatively simple physical system, which can be expressed as a conceptual or mathematical model, glacier response can be used either in a forward sense to estimate how a glacier will respond to given climatic variations or in a reverse sense to infer the climate that may have caused a glacier to advance to a particular position marked, for example, by a terminal moraine. The imprint of glacier fluctuations on the landscape is retained in the form of moraines and other glacial landforms and sediments. Cosmogenic nuclide dating has dramatically improved our ability to assign an age to these features (e.g., Balco 2011, Gosse & Phillips 2001). Using glacier models, there is potential to unravel the climatic stories hidden in well-dated moraine records. In this review, we explain the potential benefits and pitfalls of making this inverse journey from glacier length changes to reconstructions of past climate.

1.2. Historic Climate Change: Glaciers and Thermometers Tell a Similar Story

The historic record of climatic variations and glacier responses has been used to inform the physical understanding of the controls on glacier mass and length changes. Meteorological campaigns on glaciers have shown that the primary source for melt energy is solar (shortwave) radiation but that year-to-year fluctuations in mass balance are mainly due to changing air temperature (Oerlemans 2001) (see Section 2 and the sidebar titled Nonclimatic Glaciers for a few exceptions such as surge-type and tidewater glaciers). Before examining the physics responsible for this behavior, we note that the temperature sensitivity of glaciers is strongly supported by global observations

GLACIERS

A glacier is defined as perennial land ice that is maintained by accumulation of snow at high altitudes and balanced by ice ablation at low altitudes and/or discharge into the sea; it includes mountain, cirque, and valley glaciers as well as ice caps and ice fields. Ice sheets (continental-scale glaciers) are not the subject of this review for three reasons: (*a*) their response to climatic variations is generally slower than for smaller glaciers, (*b*) many of the ice sheet margins terminate in the ocean, and (*c*) the Antarctic ice sheets have always been remote from human settlement.

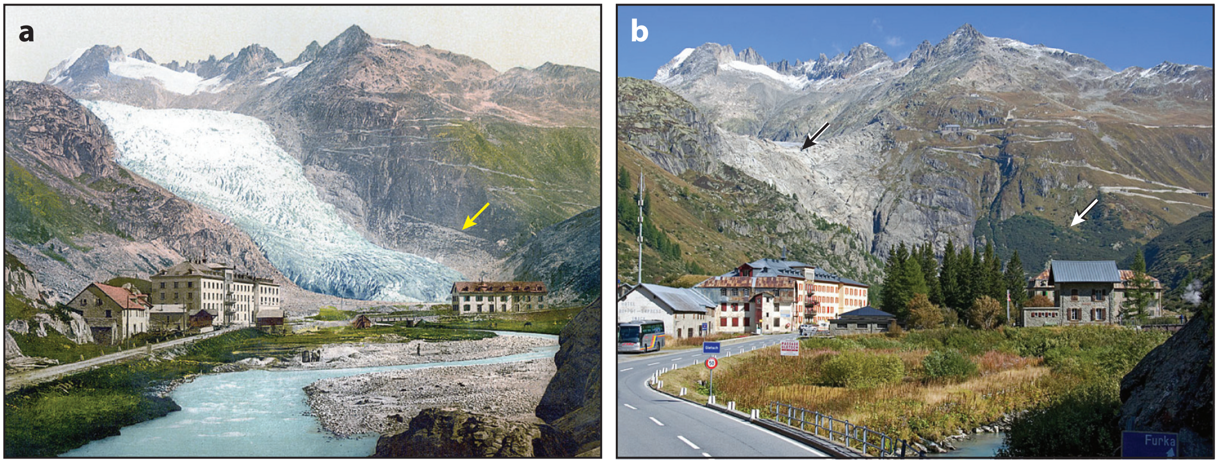

Figure 1

Rhonegletscher, Switzerland. (*a*) A hand-colored photograph from approximately AD 1900. The glacier extends to the valley floor but has already retreated from its most recent maxima in 1874 (*yellow arrow* indicates historic moraines from the seventeenth to nineteenth centuries). (*b*) Photograph taken by Juerg Alean in 2008. The glacier has retreated horizontally and vertically. Historic moraines are now vegetated (*white arrow*). Cosmogenic nuclide analysis of recently exposed, glacially scoured bedrock (*black arrow*) indicates that Rhonegletscher was smaller than it is today for >60% of the Holocene (Goehring et al. 2011).

of glaciers over the last few centuries. For example, Oerlemans (2005) and Leclercq & Oerlemans (2012) used simple glacier models to reconstruct past global temperatures over the last three to four centuries from recorded changes in hundreds of glaciers (for a subset of these glaciers, see **Figure 3**). These temperature reconstructions, which show a warming of 0.94 ± 0.31°C between AD 1830 and 2000 (Leclercq & Oerlemans 2012), are in close agreement with independent instrumental records of climate change over this period as well as multiproxy reconstructions of temperature from the preinstrumental period. More recently, Marzeion et al. (2014) used CMIP5 climate models to demonstrate that the negative glacier mass balance (since 1991) was

Glacier mass balance: change in the mass of a glacier over a fixed span of time, usually a year or season

NONCLIMATIC GLACIERS

Surge-type glaciers (e.g., Clarke et al. 1986, Kamb 1987) and tidewater glaciers (e.g., Benn et al. 2007, Meier & Post 1987) are known to undergo periodic, divergent, nonclimatic behavior. Rock avalanches onto glaciers may also alter glacier mass balance, potentially resulting in nonclimatic behavior (Shulmeister et al. 2009). However, even glaciers that are influenced by tidewater glacier cycles (Mercer 1961), periodic surges, and frequent rock avalanche input are influenced by climate to some degree. For example, in Iceland where surge-type glaciers are prevalent and many glaciers are located on active volcanoes, nearly all glaciers have exhibited dramatic retreat in response to recent climate warming (Sigurdsson & Jonsson 1995). Or in Glacier Bay, Alaska, where glacier advance and retreat into oceanic fjords is modulated by sediment supply and water depth, the dominant signal over the last century has been retreat (>100 km), with the magnitude of retreat reflecting massive ice loss due to calving of icebergs as climate warming has forced glaciers back from stable topographic positions into deep fjords (e.g., McNabb & Hock 2014). In this review, we focus on simple, climate-sensitive glaciers and point out ways to disentangle or remove signals associated with glacier dynamic influences by utilizing models that account for differences in glacier sensitivity and response time.

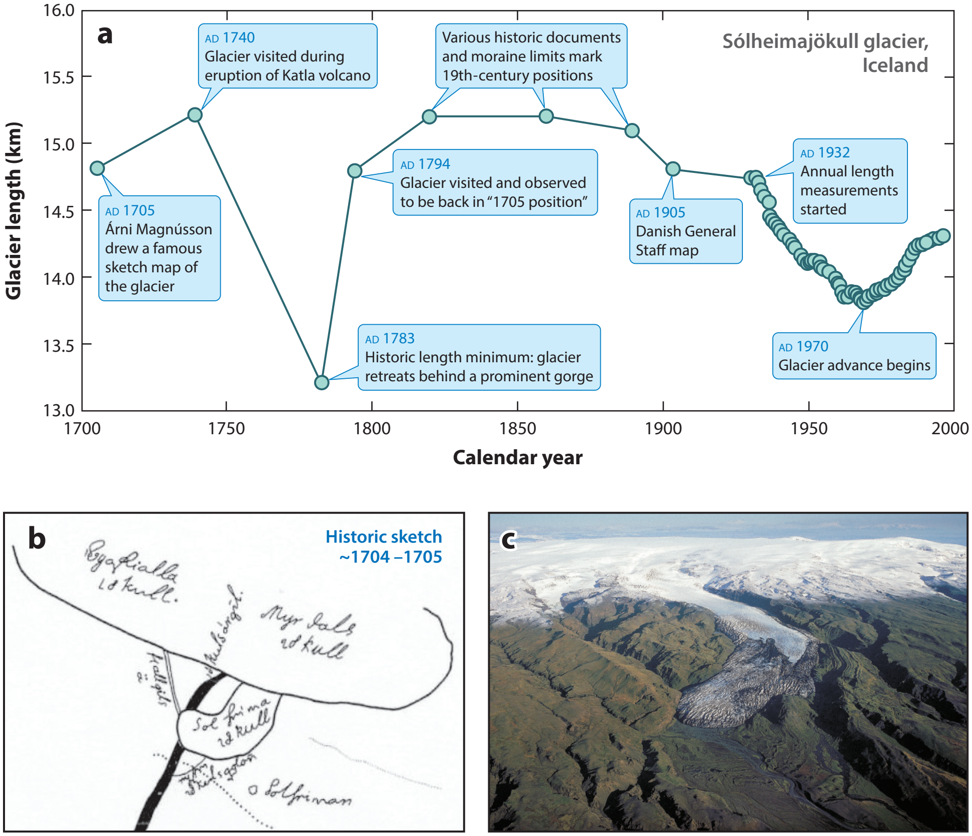

Figure 2

Example of how a glacier length record (in this case, from Sólheimajökull, Iceland) can be developed from a mixture of historical inferences, maps, sketches, landscape features, and direct measurements (Mackintosh et al. 2002). Panel *a* shows reconstructed glacier length from approximately 1705 to 2000. Panel *b* shows a historic sketch of the glacier by Árni Magnússon dating to 1704/1705. Panel *c* shows an oblique aerial photograph of Sólheimajökull in 1985. (Panel *c* courtesy of O. Sigurdsson, Natural Energy Authority of Iceland.) For the more recent record of fluctuations at Sólheimajökull, please see **Figure 3**.

predominantly due to anthropogenic influences on climate. A formal climate change attribution study of glacier length changes is still needed, although Roe et al. (2017) have recently made progress in this field (see Section 7).

Although the response of glaciers to historic climate change has been relatively homogenous on a global scale, regional differences in the timing and magnitude of glacier changes provide rich insight into both regional-scale climate variability and the internal characteristics of glaciers (see Section 7). The late-nineteenth-century retreat of glaciers in the European Alps (**Figure 3**) may have been due to an early anthropogenic influence, as the darkening of glacier surfaces by soot emitted during the industrial revolution may have led to greater absorption of shortwave radiation and increased glacier melt (Painter et al. 2013). The late-twentieth-century advance of glaciers, notably in western Scandinavia and New Zealand, reflects interannual to multidecadal variability in atmospheric and oceanic circulation (e.g., Dowdeswell et al. 1997, Fitzharris et al.

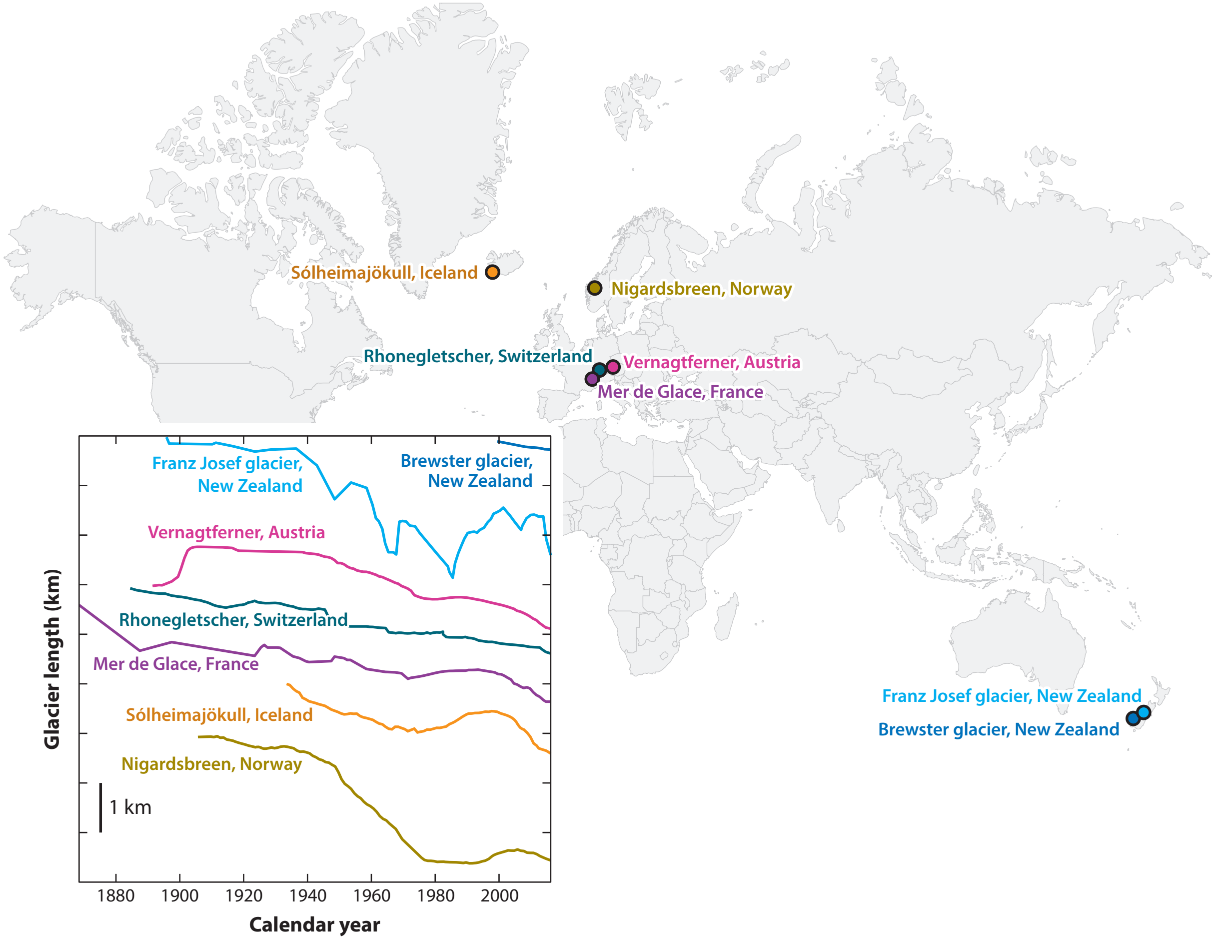

Figure 3

Length changes since the 1860s for various glaciers mentioned in the text. Glaciers with short response times (e.g., Franz Josef Glacier) have experienced repeated advances within an overall period of retreat. Slower-responding glaciers (e.g., Nigardsbreen) have experienced more muted changes. Regional differences are also evident; for example, retreat of glaciers in the European Alps began earlier than in New Zealand, and a marked readvance in the 1990s was seen in New Zealand, Norway, and Iceland. Data are from the World Glacier Monitoring Service archive (**http://wgms.ch**).

2007, Huss et al. 2010, Mackintosh et al. 2017). Glacier records could potentially be further utilized to understand anthropogenic and natural influences on the climate system.

2. GLACIER MASS BALANCE AND CLIMATE

The simulation capacity of the temperature-based melt-index method is too good to be called crude and inferior.
—Ohmura (2001, p. 753)

Glaciers can be conceptualized as a balance system: For a glacier to be in a steady state, inputs from accumulation must equal outputs from ablation over a given glacier area. Accumulation is greater than ablation in the higher-elevation portion of glaciers, and ablation is greater than

Accumulation: process that adds mass to a glacier from direct snowfall, avalanche and wind-deposited snow, refrozen meltwater, and rime ice

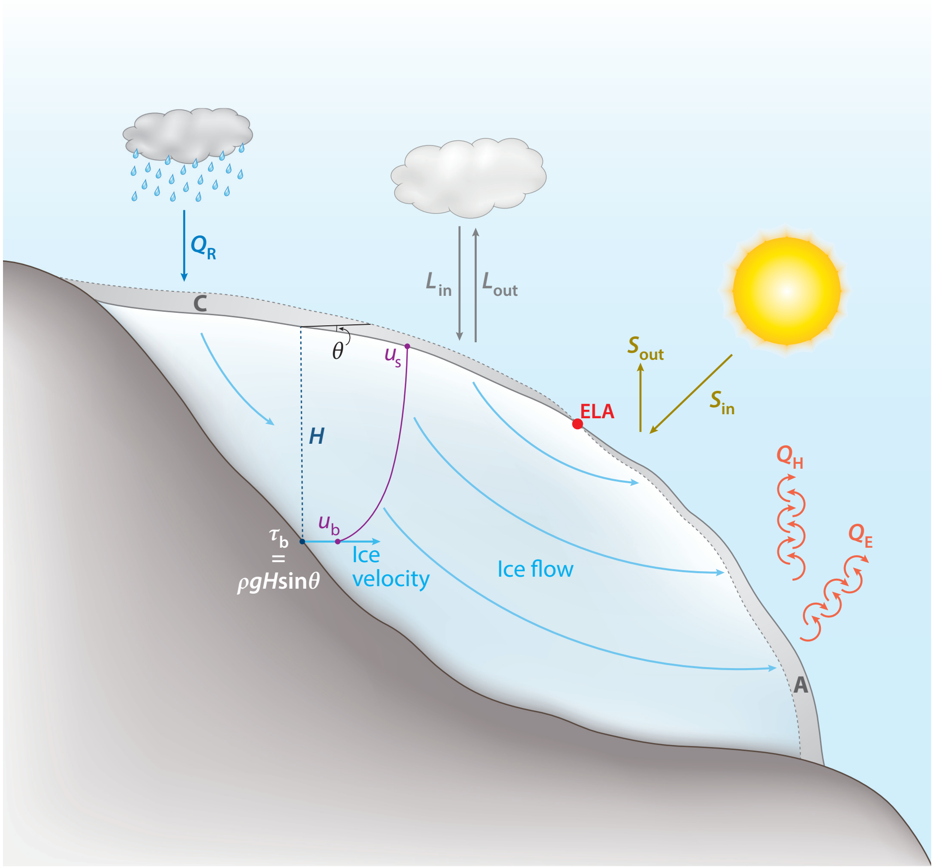

Figure 4

Schematic cross section of a glacier showing the equilibrium line altitude (ELA, *red dot*) as well as the major surface energy balance fluxes and ice flow terms. In the simplest models, ice flow is primarily controlled by local surface slope θ and ice thickness H. These values are used to calculate the driving stress τ_d, which controls the internal deformation of ice and is assumed to be locally balanced by the basal shear stress τ_b, upon which sliding u_b is often assumed to depend, resulting in the surface velocity u_s. Abbreviations: A, ablation; C, accumulation; g, acceleration due to gravity; L_{in}/L_{out}, incoming/outgoing longwave radiation; $Q_E/Q_H/Q_R$, latent/sensible/rain heat flux; ρ, density of ice; S_{in}/S_{out}, incoming/outgoing shortwave radiation; u_b/u_s, sliding/surface velocity.

accumulation in the lower sections. The equilibrium line divides the accumulation and the ablation zones (Cogley et al. 2011) (**Figure 4**). For land-terminating glaciers that do not have iceberg calving, the ablation processes depend almost entirely on the surface energy balance. On temperate glaciers, ablation is dominated by melt, and accumulation mostly results from snowfall associated with the passage of fronts and low-pressure systems as well as from orographic processes and the subsequent redistribution of snow by wind and avalanching.

Ablation: process that removes mass from a glacier from melt, sublimation, iceberg calving, wind scour, and avalanche loss

2.1. Why Are Glaciers Sensitive to Temperature Change?

Accumulation and ablation processes are both sensitive to air temperature. Owing to this high sensitivity of glacier mass balance to air temperature, glaciers require large changes in precipitation (~40%) to balance a small change in temperature (~1°C) (Oerlemans 2001). Accumulation is sensitive to temperature via its control on the snow/rain threshold (T_r), which separates precipitation falling as rain or snow (usually in the range of $0°C < T_r < 3°C$). In maritime environments, this observation, combined with the fact that (*a*) much of the precipitation on glaciers falls at temperatures close to 0°C (e.g., Cullen & Conway 2015) and (*b*) precipitation rates are generally high, means that small changes in air temperature can cause large changes in accumulation.

Ablation at the glacier surface is controlled by the surface energy balance (**Figure 4**):

$$Q_m = S_{in} + S_{out} + L_{in} + L_{out} + Q_H + Q_E + Q_R + Q_G.$$

Incoming shortwave radiation (S_{in}) varies over glacial-interglacial cycles (Berger & Loutre 1991, Eisenman & Huybers 2006, Milankovitch 1941). However, over years to decades, variability in incoming shortwave radiation is relatively small. The amount of shortwave radiation absorbed by a glacier ($S_{in} + S_{out}$) is strongly affected by its surface albedo ($\propto$), with ice (typical $\propto = 0.34$) absorbing two to three times as much shortwave radiation as fresh snow (typical $\propto = 0.75$) (Oerlemans & Knap 1998).

Cloud cover reduces incoming shortwave radiation, but the presence of clouds usually coincides with a higher (and thus compensating) incoming longwave radiation (L_{in}). Outgoing longwave radiation (L_{out}) depends on the surface temperature T_s, which for a melting surface can be assumed to equal 0°C. Rain (Q_R) and surface (Q_G) heat fluxes are relatively minor energy balance terms in most situations.

The dominant influence of air temperature on the surface energy budget can be illustrated by linearizing the budget about the state $T_s = T_a$, in which the surface temperature (T_s) and the air temperature (T_a) are identical. The net surface energy flux (Q_s), taking into account radiative, sensible, and latent heat fluxes, can be written

$$Q_s = Q_0 + b(T_a - T_s),$$

where Q_0 is the heating when $T_a = T_s$ and b is a coupling coefficient dependent on air temperature and other atmospheric conditions (see section 6.5 in Pierrehumbert 2011). The behavior of Q_0 is complicated, and the reader is referred to Pierrehumbert (2011) for a discussion. Under melting conditions, T_s is pinned at freezing so the surface warming given by the second term increases with air temperature, with proportionality constant b. For air temperatures somewhat above freezing, a typical value of b is 40 W/m^2/°C, with roughly equal contributions from latent and sensible heat flux and lesser contributions from longwave radiation (see table 6.2 in Pierrehumbert 2011). If wind is too weak to overcome the stable stratification of the boundary layer typical over melting ice, turbulent transfer is suppressed and radiation takes on a more dominant role, especially under cloudy conditions (Ohmura 2001). With the stated value of b, each additional degree of air temperature is sufficient to cause an additional 30 cm/month (water equivalent) of melt, based on the latent heat of melting of ice. This is, in essence, the physical underpinning of the temperature index method.

For the reasons explained above, many mass balance modeling studies do not calculate the full surface energy balance. Rather, temperature index or positive degree-day models rely on either an observed relationship between air temperature (in particular positive temperature sums) and ablation (e.g., Braithwaite 1981, 1995) or a combination of this temperature-based melt-index method and shortwave radiation (e.g., Hock 1999). Including shortwave radiation in calculations helps to account for nonlinear effects on mass balance, which relate to changes in accumulation and surface albedo. For example, loss of snow cover (because of reduced accumulation as the temperature rises above the snow/rain threshold) results in lowering of surface albedo and the

Equilibrium line: separates the accumulation and ablation zones, usually coinciding with the elevation of the snowline at the end of summer

Temperate glacier: glacier where snow and ice are at or close to the melting point throughout the year

Temperature index method (or positive degree-day method): empirical method for calculating ablation as a function of near-surface air temperature influence of incoming longwave radiation (L_{in}), and turbulent heat fluxes (Q_H and Q_R), which are air-temperature dependent, create a feedback into melt from shortwave radiation. For further information about glacier mass balance concepts and modeling approaches, we refer the reader to Hock (2005) and Cogley et al. (2011).

Mass balance sensitivity: change in mass balance due to a change in a climatic variable such as air temperature (or sometimes precipitation)

Mass balance gradient: rate of change of mass balance with altitude

2.2. Climatic Setting and Glacier Mass Balance Sensitivity

The overall sensitivity of glacier mass balance to climate change has some dependence on climate setting, with high-precipitation glaciers being more sensitive to temperature change than those in more arid climates (e.g., Oerlemans 1992). Mass balance sensitivity also depends on topographic factors, including the glacier's elevational range and the presence or absence of a surface debris cover. A thin (typically <2 m) surface debris cover is common on glaciers in tectonically active mountain ranges (Anderson & Mackintosh 2012) and on glaciers that are retreating rapidly (e.g., Pellicciotti et al. 2014). Debris cover reduces mass balance sensitivity. The mass balance gradient, which reflects both climatic setting and topographic influences on glaciers, is a better predictor of mass balance sensitivity than precipitation alone (Anderson & Mackintosh 2012).

We estimated glacier mass balance sensitivity for all glaciers in the global Randolf Glacier Inventory dataset (Pfeffer et al. 2014) using the empirical parameterizations of Anderson & Mackintosh (2012) (**Figure 5**). **Figure 5*a*** shows that the world's most sensitive glaciers are found in the Southern Hemisphere, where the southern westerlies deliver very large amounts of precipitation (>10 m/year) to the western slopes of the Andes and Southern Alps. High-precipitation glaciers on the western margins of Scandinavia and North America also show high mass balance sensitivity. In contrast, low-precipitation glaciers in continental, Arctic Canada show very low mass balance sensitivity. **Figure 5*b*,*c*** further illuminates these relationships: **Figure 5*b*** shows that glacier mass balance sensitivity scales linearly with annual precipitation, whereas **Figure 5*c*** indicates a slightly more complicated, inverse relationship between mass balance sensitivity and continentality.

Some glaciers exist in very cold environments where the near-surface mean annual air temperature at the equilibrium line altitude (ELA) remains below zero (**Figure 5*d***). These glaciers are found poleward of 60°S (e.g., 77°S in the McMurdo Dry Valleys, Antarctica; see Fountain et al. 1998), in the high mountains of central Asia (~30°–40°N) and South America (10°–40°S), and on isolated high peaks such as Mount Kilimanjaro (see figure 10 in Pfeffer et al. 2014). Although these glaciers accumulate in very cold environments and generally have low mass balance sensitivity (**Figure 5*d***), they commonly descend to elevations where surface melt occurs. Owing to the strong dependence of surface melt on air temperature, these melting glaciers will still dominantly advance and retreat in response to changing air temperature.

For glaciers where cold conditions prevail over their entire extent, ablation occurs mostly as a result of sublimation. Sublimation is only weakly related to air temperature, largely because for sublimating glaciers, surface temperature is not pinned at freezing and can increase with air temperature. In situations in which sublimation dominates, the attribution of recent glacier retreat to temperature change is more complicated and uncertain than is the case for temperate glaciers. For example, it has been argued that glaciers on Mount Kilimanjaro (5,000–6,000 m above sea level) retreated during the twentieth century in response to reducing precipitation (Mölg et al. 2009, 2010).

Projected climate changes (e.g., IPCC 2013) mean that in coming decades to centuries the melt zone will generally extend upward on high mountains and poleward at high latitudes. Even on Mount Kilimanjaro, unusual melt conditions may have begun in very recent times (Thompson et al. 2009). Remaining cold climate glaciers are thus likely to transition to temperate glaciers at some point in the future.

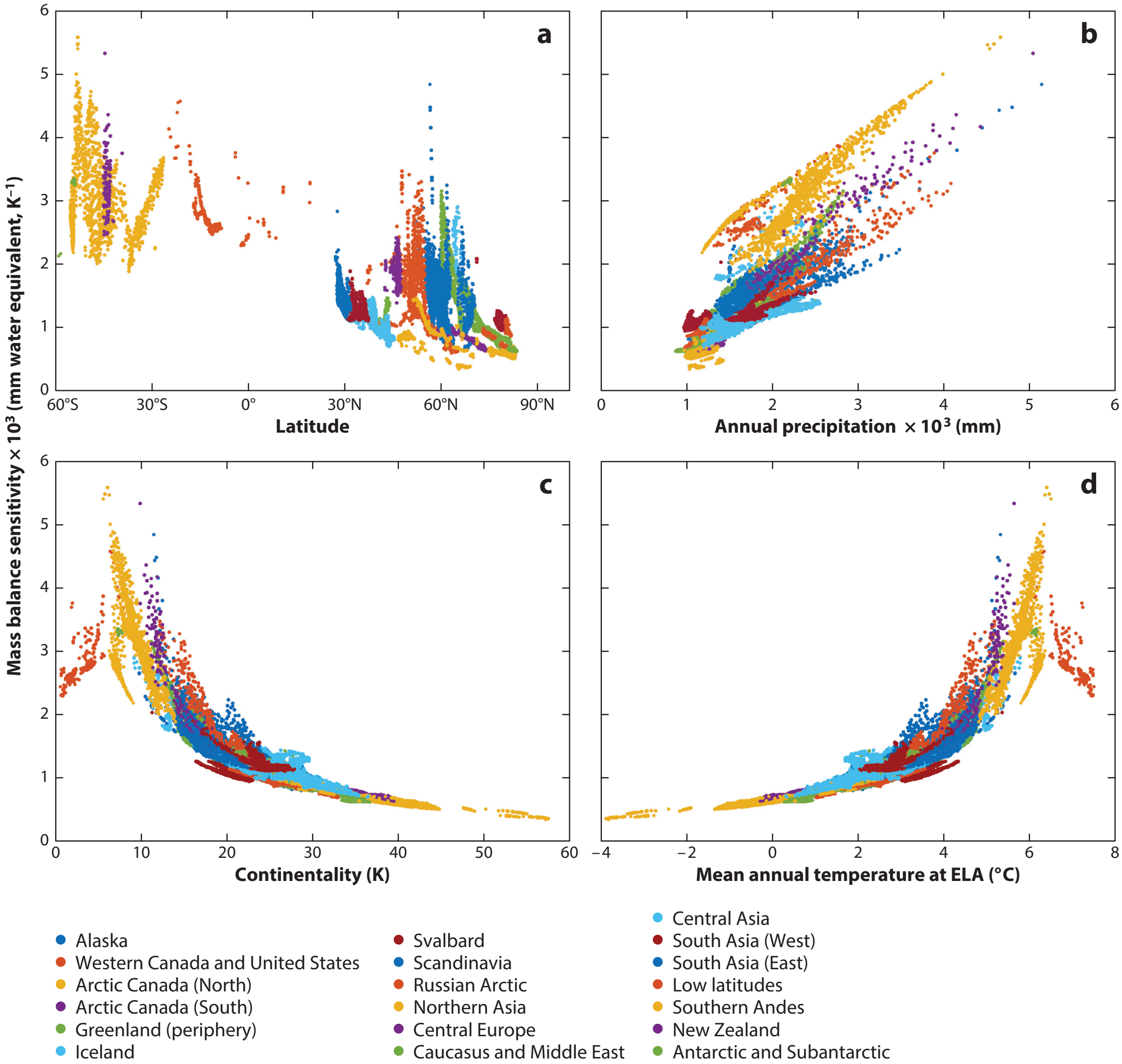

Figure 5

Mass balance sensitivity for a global dataset of glaciers (Pfeffer et al. 2014) based on empirical relationships between mass balance sensitivity and the mass balance gradient (Anderson & Mackintosh 2012). Mass balance sensitivity is plotted against (*a*) latitude, (*b*) annual precipitation, (*c*) continentality (estimated from the difference between the warmest and coolest month at each glacier site), and (*d*) mean annual temperature at the equilibrium line altitude (ELA). Climate data are from New et al. (2002). Note that panels *c* and *d* are mirror images, as temperatures at the ELA and continentality are anticorrelated (maritime glaciers have higher temperatures at the ELA, and vice versa).

3. WHAT DO GLACIERS TELL US ABOUT LARGE-SCALE CLIMATE?

The potential effects of an amplification of the global warming signal in mountain regions needs to receive greater attention from researchers and international funding agencies.

—Bradley et al. (2009, p. 3)

Temperature lapse rate: rate at which atmospheric temperature decreases with increasing elevation in the troposphere; routinely measured by radiosondes (weather balloons)

3.1. Temperature

Long before there were weather balloons or aircraft, mountains provided a window into the vertical structure of the atmosphere. The mountaineer and polymath Horace Bénédict de Saussure invented a heliothermometer in his quest to determine why temperature decreases with increasing altitude along a mountain slope; pondering the operation of that instrument was one of the things that led Joseph Fourier to conceive of the greenhouse effect (Archer & Pierrehumbert 2011). In a sense then, as mountain glaciers have become key indicators of climate change due to anthropogenic increase in the greenhouse effect, the field has come full circle.

We now know that the temperature decrease with height seen on mountain slopes is representative of the vertical structure of the troposphere in general and arises from the adiabatic cooling of air parcels when they are lifted and expand. The lifting that produces the temperature decrease does not rely on air elevating as it flows over the mountain; otherwise it would be seen along the mountain slope only when the wind is blowing upslope. The primary cause of the temperature gradient along the slope is the general air mass of the troposphere that bathes the mountain, and in this sense the mountain serves as nature's meteorological tower for sampling the atmosphere. The record of past temperatures is preserved in the form of glacial geology (see Section 5).

We have seen in Section 2 that air temperature exerts a primary controlling influence on glacier mass balance. As depicted in **Figure 6**, the temperature profile $T_{\mathrm{m}}(z)$ driving the glacier is to a first approximation the free-tropospheric temperature $T_{\mathrm{free}}(z)$ measured by a weather balloon some distance from the mountain. Glacial geology records the temperature aloft, in the general vicinity of the ELA. Estimates of past temperature at glacier elevation can be valuable, as they can be used to probe theories of the vertical structure of the atmosphere using the behavior in past climates, if independent proxy data on sea level temperature are available. More often, though, one wishes to use the glacier data to evaluate sea surface temperature (SST) changes, as an independent check of SST proxy data. Errors in early reconstructions of tropical SST at the Last Glacial Maximum were first detected using this independent check (Rind & Peteet 1985).

When reconstructing SST using glacier extent from past stadials, a substantial sea level drop relative to today caused by the growth of ice sheets must be taken into account (Betts & Ridgway 1992). Even if it is then assumed that the air temperature driving glacier mass balance exactly follows the free-tropospheric temperature at the corresponding altitude, transferring glacier-level temperature change to sea level temperature requires an assumption about the vertical temperature profile (temperature lapse rate) and the way it changes as climate changes. It has long been recognized that the vertical structure of the tropical atmosphere is controlled by moist convection and closely follows the moist adiabat, whereas the midlatitude temperature profile is governed by mixing due to large-scale synoptic eddies (Stone & Carlson 1979). Subsequent work has confirmed moist adiabatic control of the temperature lapse rate in the tropics. Because $T(z)$ for the moist adiabat is not a straight line and changes its shape as temperature and moisture content change, the temperature change at glacier level is amplified relative to the temperature change at sea level, which must be accounted for when inferring changes in sea level temperature (Betts & Ridgway 1992). The effect is illustrated in **Figure 7**, in which we plot the altitude of the 0°C isotherm versus surface temperature, assuming the vertical profile follows the moist adiabat. If the lapse rate was

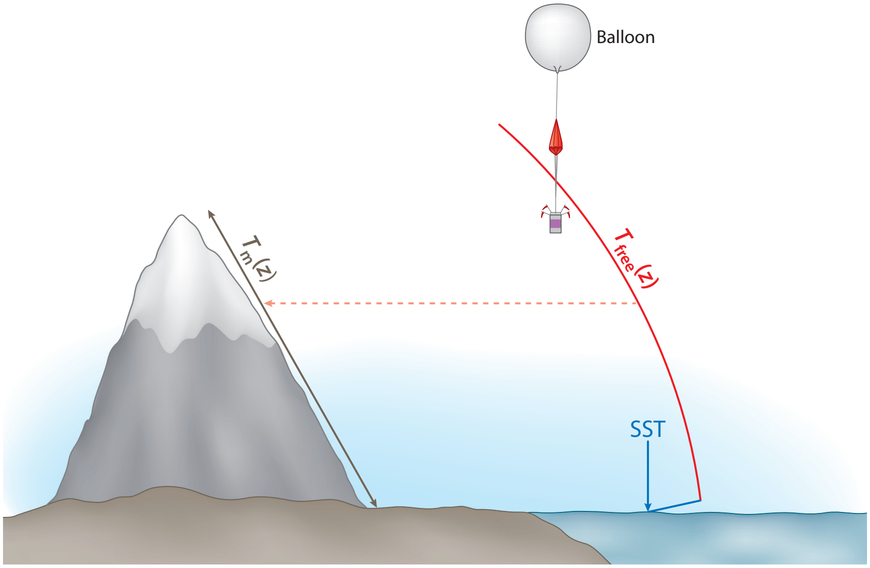

Figure 6

Schematic of the relationship between the temperature profile measured along a mountain slope (T_m) and the free-tropospheric temperature measured by a weather balloon (T_{free}), both of which depend on height above sea level, z. This sketch also illustrates the influence of temperature structure in the boundary layer, which can cause the sea surface temperature (SST) to deviate considerably from the free-tropospheric temperature at the upper edge of the boundary layer.

independent of height and remained fixed as surface temperature changes, the graph would be a straight line. In reality there is pronounced curvature, because neither condition is satisfied. As a result, a 4°C sea level temperature cooling starting from a sea level temperature of 26°C would result in a 1,300 m drop in the 0°C isotherm, from 6,000 to 4,700 m above sea level, but the same 4°C cooling applied to an initial sea level temperature of 10°C would result in only a 744 m drop in the 0°C isotherm.

The temperature gradient across the marine boundary layer further complicates the transfer of glacial-level temperature to SST (Betts & Ridgway 1992). As sketched in **Figure 6**, SST can be quite different from the free-tropospheric temperature at the outer edge of the boundary layer. Especially in regions of strong cold upwelling, the SST can be considerably below the low-level free-tropospheric temperature, though it is difficult for the SST to much exceed the free-tropospheric near-surface temperature, because positive temperature excess drives strong convection, which tends to wipe out the gradient. Nonetheless, tropical freezing altitude correlates very well with tropical SST. Bradley et al. (2009) analyzed the relationship of high-altitude surface station temperatures with tropical-mean SST changes and found that a 1°C reduction in mean SST yields a 253 m depression of the freezing line altitude. This is notably less than would be inferred from the moist adiabat for typical tropical SSTs (i.e., 325 m per degree according to **Figure 7**). It is not known at present whether this moderate mismatch reflects a true deviation of the tropical vertical structure from the moist adiabat or a failure of the local temperature driving the glacier to exactly follow the free-tropospheric temperature of the corresponding height.

For the midlatitudes the situation is much more unsettled, because there is no simple quantitative theory accounting for the way synoptic eddies determine the vertical structure. Frierson (2008) discusses some of these complexities. As an example of the deviation of the midlatitude

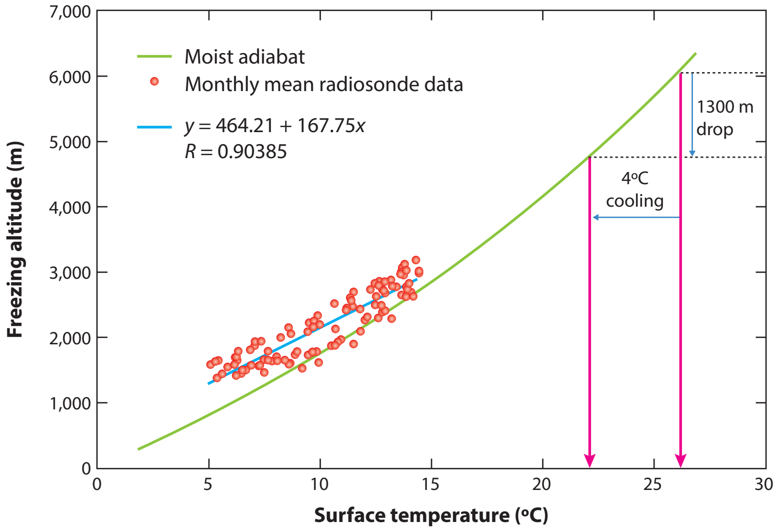

Figure 7

Relation of the freezing line altitude to sea level temperature. The solid green line indicates results assuming the vertical temperature structure is governed by the moist adiabat. Red dots are for monthly mean data from the Invercargill radiosonde data (46.4°S, 168.4°E) on the coast of New Zealand's South Island for 2010–2016. The blue line is the linear regression of the data, corresponding to a 168 m depression of the 0°C isotherm for each degree of cooling of the surface temperature.

temperature lapse rate from the moist adiabat, in **Figure 7** we plot the monthly mean freezing height versus monthly mean 1,000 hPa (surface) temperature based on the Invercargill radiosonde data for 2010–2016 on the coast of the South Island of New Zealand. Variations in this record primarily reflect the seasonal cycle but also to some extent reflect interannual variability. The freeze line clearly does not conform to the moist adiabat and has a lower slope versus surface temperature than the moist adiabat for the corresponding range of temperatures.

To make matters worse, it cannot be assumed that the regression of glacier-level temperature on sea level temperature one infers from the seasonal cycle is valid for other causes of climate change (e.g., CO_2 changes or Milankovitch orbital forcing). So what is to be done? Until a usable theory of midlatitude lapse rate comes along, we think the best approach is to use general circulation models of climate change forced by specific mechanisms (e.g., the combination of CO_2 and Milankovitch and ocean heat transport changes that led to the substantial Southern Ocean cooling during the Last Glacial Maximum) and their resulting temperatures to drive glacier models and then to compare the inferred glacial extent with the moraine record (see Section 7). There is considerable intermodel variability in how lapse rate changes under various kinds of climate change, so ideally, such a study would be done with a multimodel ensemble and provide an estimate of the uncertainty of the results.

Two additional questions arise when inferring large-scale climate properties from glacial geology. The first key question is the spatial extent of the region one can characterize by study of glacier length fluctuations. Assuming temperature to be the main controlling influence, the characteristic scale is given by the Rossby radius of deformation, which characterizes the spatial coherence scale of temperature on a rotating planet. This length is NH/f, where N is the stability frequency, H is the scale height of the atmosphere, and f the Coriolis parameter. In the midlatitudes (around 45°N or 45°S), this is on the order of 750 km, so midlatitude glaciers are indicative of temperatures in the

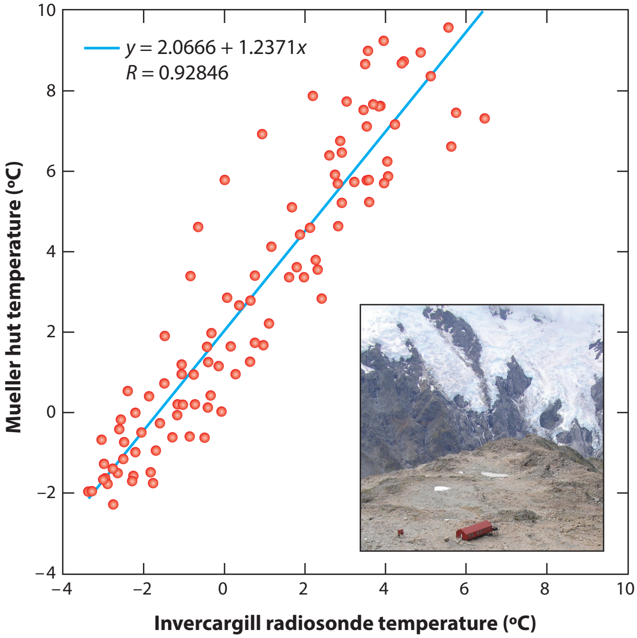

Figure 8

Regression of monthly mean surface temperature data (°C) at the Mueller Hut, South Island of New Zealand (43.7°S, 170°E, 1,818 meters above sea level) against monthly mean free-tropospheric temperature (°C) at the same altitude at Invercargill, approximately 400 km distant. Data are for 2010–2016, minus periods in 2012 when Mueller Hut data were not available.

surrounding few hundred kilometers (also see Section 5). In contrast, the radius of deformation near the equator becomes unbounded, and the whole tropics has weak temperature gradients in the horizontal (Williams et al. 2009). This makes tropical mountain glaciers especially powerful indicators of climate, as they reflect the temperature over the entire tropics (Bradley et al. 2009).

The second key question is the extent to which the temperature profile along the mountain deviates from the free-tropospheric temperature at the corresponding height. As an illustration of the issues, in **Figure 8** we plot the monthly mean station data at Mueller Hut (1,818 m above sea level) in the New Zealand Southern Alps against the free-tropospheric data from the closest available radiosonde (Invercargill). The mountain station temperature correlates well with the free-tropospheric temperature, but clearly the two temperatures are not equal. The regression shows that fluctuations in the mountain station data are amplified by approximately 24% relative to the free-tropospheric temperature and are approximately 2°C warmer. The offset may just represent a temperature gradient owing to Mueller Hut being further equatorward than Invercargill, but the slope indicates that something else is going on. Examination of the diurnal and seasonal cycle at Mueller Hut suggests the amplification is due to greater continentality of the station data, which respond rapidly to local solar heating. This is an example from only one case, but it illustrates the principle that in relating glacial geology to free-tropospheric temperature change, one must not fall prey to the fallacy that temperature driving the glacier is equal to the free-tropospheric temperature change, though it is sound to assume the two temperatures are strongly correlated. A proper inference of ambient temperature requires an identification of the physical processes governing this correlation and an assessment of how they might respond to various mechanisms forcing climate change. This represents fairly unexplored territory.

3.2. Precipitation

When it comes to temperature, mountains "make their own weather" only to a slight extent, and the temperature driving a glacier is quite representative of the temperature in a large surrounding region. This is not the case for precipitation, which is crucially affected by processes in the immediate vicinity of the mountain. Precipitation in mountainous regions can be orders of magnitude greater than precipitation in immediately adjacent flat areas. The study of orographic precipitation has a long and deep history and still engages many unresolved issues.

The fundamental reason that mountains amplify precipitation is that moist air impinging on the mountains is lifted and cooled, causing condensation. This can lead to a great deal of precipitation relative to what can be produced by synoptic eddies or convective systems over flat surfaces, because a strong wind can bring in a steady and essentially inexhaustible supply of moisture. This insight is the basis of the upslope precipitation model discussed in Roe (2005) and Smith (2006). In the upslope model, it is assumed that the entire column of moist air impinging on the mountain with uniform air speed U is lifted and cooled adiabatically, allowing for latent heat release from condensation, so that water vapor must be precipitated out as rain or snow to keep the air from becoming supersaturated. The net result is that the precipitation rate along a mountain with height profile $h(x)$ is proportional to $q(x)U\mathrm{d}h/\mathrm{d}x$, where $q(x)$ is the column-integrated amount of water remaining (in units of $\mathrm{kg/m^2}$) at position x along the mountain. $q(x)$ goes down as the mountain crest is approached, because water is progressively removed by precipitation, so the two factors at play are the slope of the mountain (which enhances the precipitation) versus the progressive loss of water as high elevations are approached. For this reason, upslope models tend to yield maximum precipitation partway up the slope of high mountains, with marked reduction toward the crest. Despite their simplicity, upslope models do fairly well at capturing the general magnitude of orographic precipitation. They indicate that orographic precipitation is sensitive to synoptic and larger-scale circulation patterns mainly through the wind direction and is largely decoupled from the precipitation mechanisms that would be yielded by these systems over a flat surface. In particular, there is little point in driving a glacier model with the precipitation yielded by a large-scale climate model with smoothed out orography; it would generally be better to throw away the precipitation produced by the climate model and reconstruct it from the wind, temperature, and moisture fields using some variant of an upslope precipitation model (e.g., Clarke et al. 2015). Alternatively, one might drive a glacier model with very high-resolution climate model output (e.g., Mölg & Kaser 2011; see also Section 7). For further discussion of the upslope model and its various refinements, and the extent to which it can be reconciled with the microphysical processes involved in orographic precipitation, the reader is referred to Roe (2005) and Smith (2006).

Orographic precipitation in past climates represents a considerable challenge to the endeavor of reconstructing past climate from glacial geology, though the challenge is mitigated by the fact that it usually takes a very large proportionate change in precipitation to have the same effect on mass balance as a rather modest change in temperature. Recall, moreover, that temperature still exerts a controlling effect on the glacial mass balance associated with a given precipitation rate by controlling the altitude of rain versus snow. Further, given some estimate of the large-scale wind field (e.g., from a climate model), the general physical principle that cold air holds less water in saturation can help constrain likely changes in precipitation rates. Finally, in some cases it may be possible to directly constrain precipitation rates in past climates (for more on both possibilities, see Section 7).

4. GLACIER LENGTH CHANGES

At the core of the mystery and awe surrounding glaciers is their changeability.

—Carey (2007, p. 501)

4.1. Glacier Flow

Glacier flow acts to move mass from the upper part of the glacier, where net accumulation is positive (i.e., more snow falls than can melt over a year), to the lower part of the glacier, where net ablation is positive (**Figure 4**). Hence the most basic control on glacier flow is that of mass conservation $\nabla \cdot \boldsymbol{u} = M$, where $\boldsymbol{u}$ is the ice velocity field and M is the mass balance. The flow field $\boldsymbol{u}$ depends on the geometry of the ice mass and its rheology and can be treated as the flow of a viscous material that, with some simplifying assumptions, is governed by the Navier-Stokes equations, which can be written with the body force as gravity and in terms of the pressure (p) and deviatoric stress as

$$0 = -\nabla p + \nabla \cdot \tau' + \rho g.$$

Glacier response time: timescale for a glacier to undergo approximately two-thirds of its volume adjustment following a step change in mass balance

The deviatoric stress τ' tensor is calculated according to the constitutive relation, which links stress and strain rates (ϵ) and is often taken as Nye's generalization of Glen's flow law (with the flow exponent $n = 3$):

$$\epsilon_{ij} = A\tau_{\mathrm{eff}}^2 \tau'_{ij},$$

where the effective stress τ_{eff} is the second invariant of the deviatoric stress tensor. The Stokes equation is still computationally expensive to solve, and so a series of further simplifications can be made. These simplifications arise from the observation that ice sheets (and most glaciers) are generally much thinner than their areal extent. A natural aspect ratio $\upsilon = d/l$, where d is the characteristic depth and l is the characteristic length, is applied in the expectation that $\upsilon \ll 1$. The Stokes equations can then be written as a series expansion, and an approximation can be made by using only a finite number of terms from the expansion. The approximation that is commonly known as the shallow ice approximation (SIA) is formally termed the zeroth-order SIA (Cuffey & Patterson 2010). This approximation neglects stress terms that are important for the flow of glaciers and ice sheets near their margins. The stresses most important in these areas are the longitudinal deviatoric stresses τ_{xx} and τ_{yy}.

A consequence of neglecting the full stress field is that at equilibrium, the zeroth-order SIA produces a glacier that is slightly shorter and thicker than Stokes models (Le Meur et al. 2004). For time-dependent simulations, little difference is seen in retreat scenarios when ice flow plays a lesser role in glacier response but significant differences are seen during advance phases when flow dominates (Adhikari & Marshall 2013, Egholm et al. 2011).

4.2. Glacier Response Time

Glaciers integrate year-to-year variations in surface mass balance over their characteristic response time, which is a function of their mass balance and geometry. The response time is distinct from the reaction time, which is the time lag between a climate forcing and response at the glacier terminus. Glacier reaction times are easy to measure (e.g., Purdie et al. 2014), but they are not very useful for characterizing glaciers. This is because reaction times are not a physical constant, as they depend on glacier geometry and the history of mass balance forcing (Oerlemans 2001). Glacier response times are more useful. However, as response times require a step change in climate that is maintained indefinitely, they cannot be directly calculated from field data.

John Nye developed our understanding of the basic physics of how a glacier dynamically adjusts its length in response to climate forcing (e.g., Nye 1951, 1960, 1965). Later, Tómas Jóhannesson and others used a geometric simplification of the dynamic equations governing glacier response to show that response time (τ) can be approximated as

$$\tau = H/b_t,$$

where H is the ice thickness and b_t is the ablation rate at the glacier terminus (Jóhannesson et al. 1989). According to this formula, the typical response time for glaciers is several decades, consistent with observations that they change their length on multiannual to decadal timescales. Although this simple formula does not consider some aspects of glacier response (e.g., the feedback between glacier elevation and surface mass balance), it can be quickly calculated from field data. Another simple approach, proposed by Oerlemans (2001), takes the mean length (L) and a characteristic velocity (u) to estimate a response time as $\tau = L/u$, based loosely on the kinematic wave theory of Nye (1960). However, these simple approaches provide somewhat idealized results, and response time is still estimated most robustly by using numerical models.

The response time of glaciers does not increase linearly with glacier size, as one might expect. Large glaciers do tend to have long response times because of their lower surface slope (and thus their greater ice thickness), compared with short glaciers. However, relatively long but steep glaciers can have shorter response times than smaller, gently sloping cirque glaciers (Bahr et al. 1998). In Section 5 we discuss how glacier response time influences moraine formation and efforts to reconstruct climate from glaciers.

5. MORAINES AND THEIR PALEOCLIMATE SIGNIFICANCE

> *The fact that climatic changes of the last millennium are recorded by numerous moraine systems but pass undetected in most other paleoclimatic records suggests that alpine-glacier fluctuations afford a valuable monitor of high-frequency climatic variations.*
>
> —Denton & Karlén (1973, p. 157)

5.1. Cosmogenic Exposure Dating of Moraine Sequences

Lateral and terminal moraines, along with erosional trimlines, are commonly preserved in the landscape, marking the former maximum extent and thickness of glaciers in recent geological time (e.g., Benn et al. 2003) (**Figures 1** and **9**). Until recently, moraines could be dated only indirectly, using relative-age indicators (e.g., soil and vegetation development, lichenometry, weathering rinds; see Haeberli et al. 2003), or in situations in which organic material associated with moraines could be radiocarbon dated (e.g., wood incorporated in moraines as a glacier readvanced over forested terrain; see Denton & Hendy 1994, Roethlisberger & Schneebeli 1979). In the last few decades, cosmogenic exposure dating of moraine boulders has provided direct ages for moraine deposition and abandonment in most of the world's mountain ranges (Balco 2011). Cosmogenic dating can also inform us about periods when glaciers were smaller than they are today (e.g., Goehring et al. 2011) (**Figure 1**). Furthermore, visualization tools such as Google Earth now make it possible to rapidly search for moraines to target for dating. Although the application of cosmogenic ages requires careful landscape mapping and sample selection (e.g., Putnam et al. 2012) with consideration of the various geological factors that may affect ages (Applegate et al. 2012, Gosse & Phillips 2001, Putkonen & Swanson 2003) (see also Section 5.2), this technique has already transformed our understanding of past glacier fluctuations and associated climate changes. For example, in New Zealand, the application of beryllium-10 (^{10}Be) dating at a number of moraine sites has produced high-precision ages that extend from the nineteenth century to at least 130,000 years ago (e.g., Kaplan et al. 2010, Putnam et al. 2013b).

5.2. Glaciological and Geological Controls on Moraine Formation

Glacier response time is an important factor influencing moraine formation. Fast-responding glaciers may react to both high-frequency (e.g., interannual to decadal) and low-frequency (e.g.,

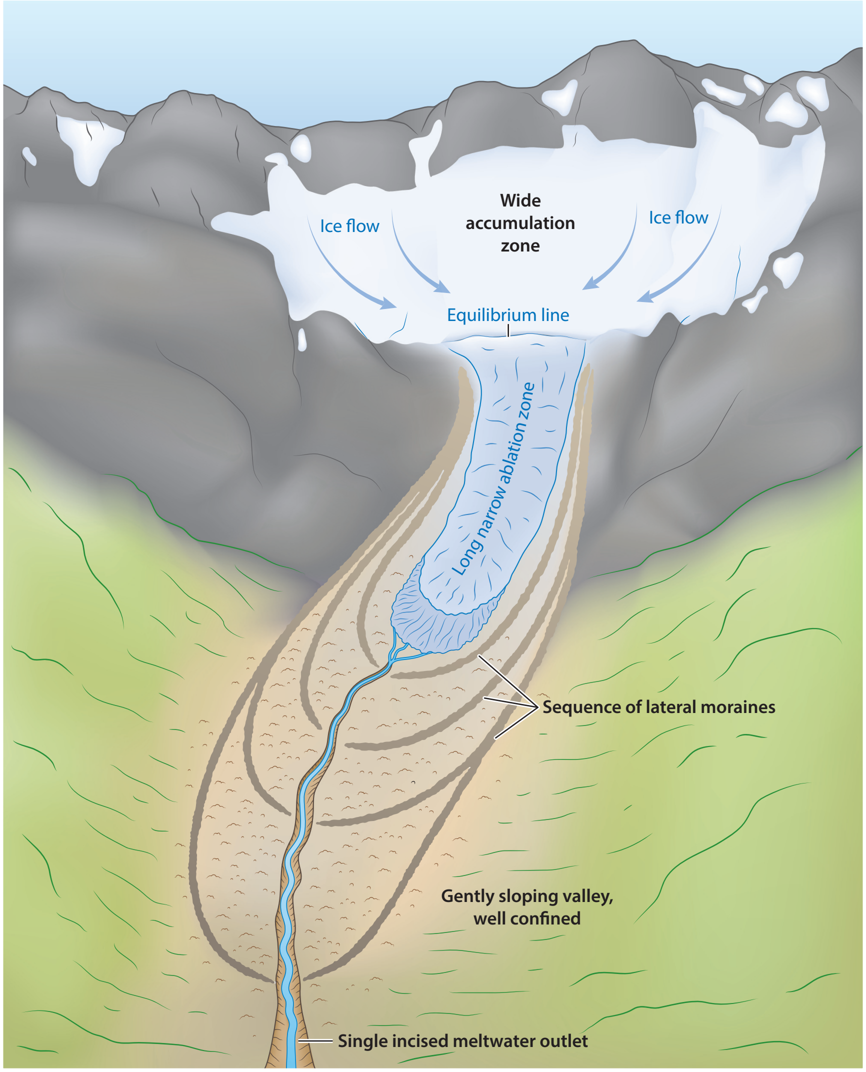

Figure 9

Glaciological and geological factors that promote deposition and preservation of a detailed moraine sequence. The wide accumulation area and long narrow snout, terminating in a confined narrow valley, promote a sensitive glacier length response. An incised meltwater stream allows moraines to be preserved rather than buried or washed away by fluvial activity.

multidecadal to centennial) climate changes. Slow-responding glaciers may respond only to low-frequency climate changes, but they still exhibit broadly similar behavior to adjacent glaciers with shorter response times. This is why moraine sequences show regional similarities over distances of hundreds of kilometers (also see discussion of the Rossby radius of deformation in Section 3). **Figure 10** illustrates how, in an idealized modeling experiment, both Franz Josef (response time ~20 years) and Brewster Glaciers (response time ~50 years) are forced by an identical climate sequence with a warming trend. Both glaciers retreat; however, Franz Josef Glacier shows

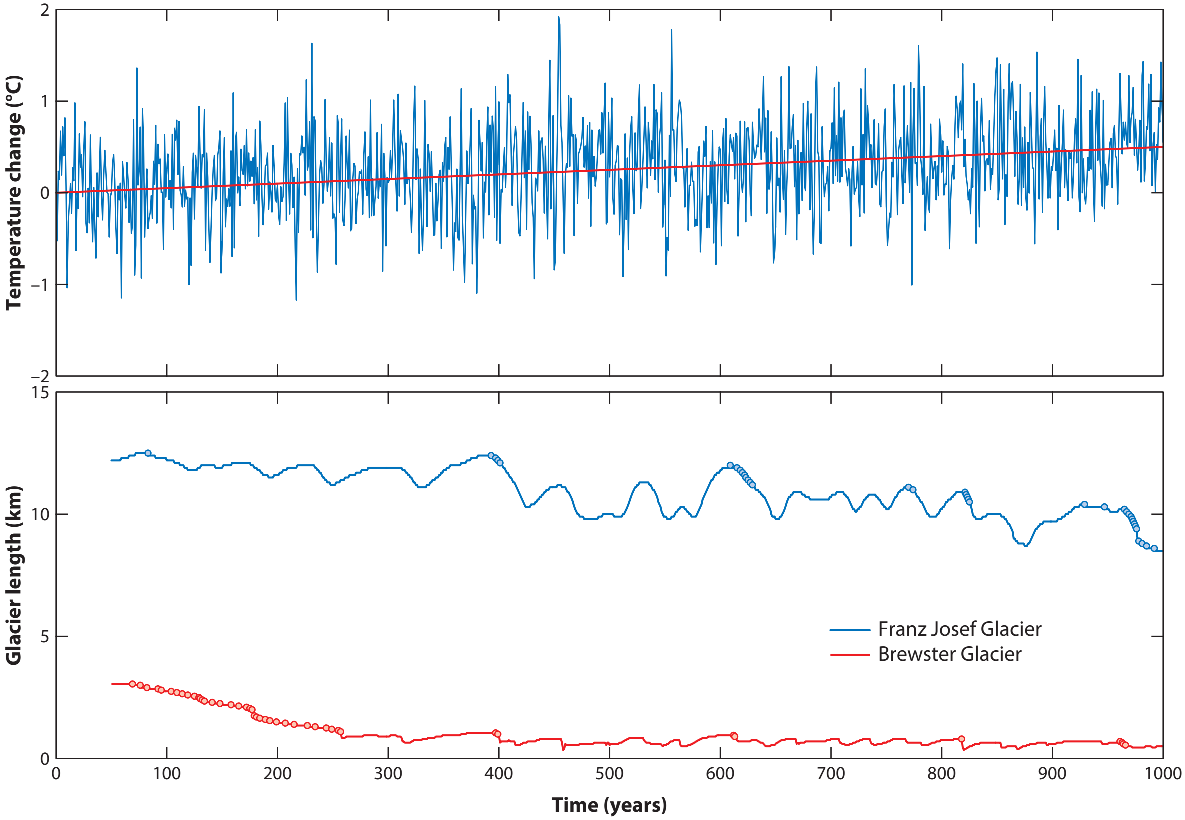

Figure 10

The modeled response of two different glaciers highlights the potential for glacier response time to influence moraine formation. In this case Franz Josef Glacier (response time ~20 years) and Brewster Glacier (response time ~50 years) are forced with an identical 1,000-year temperature sequence, which includes random interannual variability (standard deviation = 0.5°C) superimposed on a 1°C warming trend (this is superficially similar to the climate of the last millennium). Moraines (marked by red and blue circles) have the potential to be deposited when a glacier retreats from an advanced position and does not subsequently overrun that position during the simulation. Franz Josef Glacier has the potential to form moraines at approximately 12 different intervals. Brewster Glacier experiences approximately 5 potential moraine-building intervals. The timing of the (~5) major moraine-building intervals, however, is similar for both glaciers.

more frequent fluctuations (also see **Figure 3**) and has the potential to deposit a more detailed moraine sequence, whereas Brewster Glacier retreats almost monotonically.

Glacier length sensitivity: change in glacier length due to a set change in a climatic variable such as air temperature (or sometimes precipitation)

Glacier length sensitivity is another important glaciological factor influencing moraine formation. Glaciers with a high length sensitivity exhibit large terminus fluctuations in response to climate forcing, for example, Nigardsbreen in Norway (Oerlemans 1997). Such glaciers are likely to spread their moraine sequences over a great distance along a valley floor (**Figure 9**). In contrast, glaciers that terminate at a break in slope, and/or in positions where valley widening occurs (e.g., Mueller Glacier, New Zealand), tend to deposit large, compound moraine sequences through the repeated occupation of a site during multiple advance/retreat cycles (Roethlisberger & Schneebeli 1979). Both types of moraine sequences offer potential for dating past glacier fluctuations (e.g., Putnam et al. 2012, Schaefer et al. 2009).

It is tempting to infer that large moraines represent extended periods when glaciers were in equilibrium (i.e., achieved a steady state) with climate. However, sediment supply is also an important factor influencing moraine size. The sediment flux to a glacier terminus depends on the dynamic characteristics of a glacier (erosion at the bed depends on the sliding velocity to the second power; see Herman et al. 2015) as well as the contribution of sediment from surrounding valley walls. Although the basal erosion rate can be calculated using a numerical model, sediment input from valley walls is highly variable in space and time and can confound attempts to relate moraine size to time. Postdepositional moraine degradation or erosion (e.g., Applegate et al. 2012) is an additional source of uncertainty that potentially increases with moraine age. Further modeling could be carried out to improve our understanding of these relationships.

Some of the geological factors that can result in destruction or preservation of moraine sequences are illustrated in **Figure 9**. This figure is not comprehensive; we refer the reader to Benn & Evans (2010) for a full treatment of glacial landsystems. **Figure 9** shows that the routing of meltwater streams is critical: Glaciers that terminate on outwash (sandur) plains tend not to deposit terminal moraines, as most debris arriving at the terminus is preferentially incorporated into streams (e.g., Benn et al. 2003). However, such glaciers often deposit detailed lateral moraine sequences (e.g., Sólheimajökull, Iceland, or Tasman Glacier, New Zealand) that can still be targeted for moraine dating (e.g., Schaefer et al. 2009). Where meltwater is removed from the terminal environment within bedrock channels or gorges, moraines have the greatest potential to be deposited and preserved (**Figure 9**). Moraine sequences in such topographic settings can be very detailed, offering the potential for deriving a rich climate history (e.g., Jomelli et al. 2014, Kaplan et al. 2013, Putnam et al. 2012).

6. CLIMATE RECONSTRUCTION FROM GLACIERS

The use of individual glacier records for climate reconstruction was first given a mathematical basis by [Nye (1965)], but the method has hardly been used.

—Oerlemans (2001, p. 93)

6.1. Equilibrium Line Altitude Reconstructions and Early Insights into Ice Age Cycles

In the 1970s, Stephen Porter pioneered a simple technique to reconstruct climate from geological evidence of past glacial extent (e.g., Porter 1975). The method relies on the fact that accumulation above the ELA must match ablation below the ELA. Many studies have shown that, for this to occur, the ratio of the accumulation area to the total area of the glacier (the accumulation area ratio, or AAR) must be an average of approximately 0.66, although this value results from a combination of glacier geometry and mass balance gradient and can be expected to be different for each reconstructed glacier.

By mapping glacial landforms in a valley previously occupied by ice and reconstructing the geometry of the former glacier, it is possible to estimate the ELA using this AAR. To compare against modern climate, the present-day end of summer snowline (averaged over many years) is taken as the ELA (Ohmura et al. 1992). The difference between these reconstructed and modern ELAs then represents the climatic shift associated with the change in glacier geometry. This powerful insight provided a basis for assessing ice age climate changes along entire mountain chains. Broecker & Denton (1990) applied this method to a transect extending from Antarctica to the Arctic Ocean along the American Cordillera. They showed that Last Glacial Maximum equilibrium lines were depressed on average by 1,000 m (see Section 3 for a discussion on what this means for associated sea level temperature reductions).

6.2. From Equilibrium Line Altitude Reconstructions to Physics-Based Glacier Models

Researchers sometimes assume that glacier-based climate reconstructions derived from numerical models are more precise or reliable than those carried out using ELA reconstruction methods. This is not necessarily the case; well-constrained ELA reconstructions and numerical modeling output should produce similar, if not identical, results for steady-state reconstructions of geometrically simple glaciers, for which the AAR can be assumed to take a standard value (Doughty et al. 2013, Kaplan et al. 2013).

Even though the theory and numerical approach for reconstructing climate from glaciers was developed in the 1960s and 1970s (Budd & Jenssen 1975, Nye 1965), it wasn't until the 1980s that we began to develop numerical models for a number of glaciers with the explicit aim of reconstructing past climate (e.g., Greuell 1992, Oerlemans 1986). Model-based reconstructions rely on a mass balance model, which translates climatic variations into mass balance variations, and a flow model, which uses the principles of mass conservation and ice flow to translate changes in mass balance to changes in glacier geometry. Models offer several potential improvements over manual ELA reconstructions:

1. Glacier models require no assumptions about glacier shape, a predefined AAR value, or a mass balance gradient.
2. Glacier models allow the local topographic effects, such as shading from shortwave radiation, to be directly considered (Plummer & Phillips 2003).
3. Glacier models allow for sensitivity tests that can lead to a fuller analysis and quantification of the uncertainties inherent in glacier-climate reconstructions. Uncertainties typically relate to lack of knowledge of climate variables (e.g., past precipitation levels) or model parameters (degree-day factors and temperature lapse rates or, in the case of energy balance models, roughness lengths for turbulent fluxes and albedo values for snow and glacier ice; e.g., Doughty et al. 2013, Machguth et al. 2008).
4. Glacier models can be used to carry out hierarchical experiments, for which the aim is to identify and rank the most important climate variables influencing glacier response. For example, Anderson & Mackintosh (2006) evaluated whether temperature or precipitation drove a late-glacial advance of Franz Josef Glacier to the Waiho Loop moraine in New Zealand.
5. Glacier models can be used for transient experiments, whereby glacier length changes are forced by an independent climate proxy reconstruction, such as a paleo-ecological reconstruction or ice core record (Doughty et al. 2013, Hubbard 1999), and the output is then compared with moraine records. If successful, such experiments can provide confidence in the climate proxy record used to force the simulation, the model parameterization, and the moraine chronology.
6. Glacier models can be driven by climate models. In such cases, glacier modeling experiments might help to evaluate climate model performance (e.g., in locations where well-dated moraine sequences exist and modern glacier-climate relationships are well understood).
7. Glacier models have the option of refining or including new processes as our understanding of glacier-climate physics improves.
8. Model processes can also be switched on or off; such heuristic experiments can lead to a greater understanding of the glacier-climate system.

The glacier modeling approach that one uses should vary in complexity, depending on the question being asked and the data available. Simplified linear equations that relate glacier length to a particular climate state are suited to climate reconstructions using hundreds of glaciers (Oerlemans 2005) or simulations of glacier response over millennial timescales (Oerlemans 2001, Roe 2011). At their most complex, Stokes models of glacier flow that calculate the full stress fields (e.g., Cuffey & Patterson 2010) can be driven by ocean-atmosphere general circulation models. Such models are useful for predicting future glacier extent (e.g., Réveillet et al. 2015), but they are less suitable for reconstructing past climate. This is because climate reconstructions are often based on long model integrations, requiring significant computational resources. Fortunately, Stokes models are probably not needed for most climate reconstructions.

We favor the intermediate approach of using a temperature index or energy balance model to calculate mass balance, coupled to an SIA flow model, as a flowline in either a single dimension (e.g., Anderson & Mackintosh 2006, Malone et al. 2015) or two dimensions (Doughty et al. 2013, Eaves et al. 2016). The SIA is solved on a digital elevation model or on parameterized valley width and depth, in the case of a flowline model. Temperature index and energy balance models are driven by climate fields from individual weather stations (Anderson et al. 2006), interpolated station data (Anderson & Mackintosh 2012), and/or climate gridded datasets including reanalyses (e.g., Birkel et al. 2012). Temperature index models require only temperature and precipitation as inputs, whereas energy balance models have greater data requirements, needing precipitation, wind speed, relative humidity, cloud cover, and solar radiation.

An advantage of 2D or 3D models is that they are directly comparable with field evidence, preserved either in the landscape (Birkel et al. 2012, Egholm et al. 2011, Golledge et al. 2012, Kessler et al. 2006, Plummer & Phillips 2003, Putnam et al. 2013a) (**Figure 11**) or in offshore sedimentary basins (e.g., Pollard & DeConto 2009). Such climate reconstructions have been particularly successful in geometrically simple catchments with well-mapped and dated moraine sequences (e.g., Doughty et al. 2013; Kaplan et al. 2010, 2013) (**Figure 11**) or in models where many catchments have been simultaneously reconstructed (e.g., Birkel et al. 2012, Golledge et al. 2012).

Model evaluation is a necessary step in all glacier modeling simulations, because glaciers are very sensitive natural systems (e.g., Oreskes et al. 1994) and glacier models include several processes that are parameterized (e.g., turbulent flux calculations and glacier sliding). Present-day glacier measurements of mass balance and flow on the glacier can be used to calibrate and test glacier models (e.g., Anderson & Mackintosh 2006, Hubbard et al. 1998). When these measurements are not available, less direct measurements such as the extent of present-day glaciers or the longitudinal profiles of paleoglaciers can be used to constrain or test model parameters (Oerlemans 1997). However, even for the best-constrained experiment there is still a need to test the sensitivity of the final output of the model (e.g., a reconstruction of past temperature change relative to present-day climatology) against the uncertainties in each parameter. Typically, the parameters that the models are most sensitive to, and therefore that most effort should go into constraining, are the surface albedo (Section 2), temperature lapse rate (Section 3), and sliding parameters within the ice flow model (Section 4) (e.g., Doughty et al. 2013).

There are several limitations in reconstructing climate from standalone glacier modeling experiments. Disentangling the effects of temperature and precipitation in these simulations is not easy, and using glacier model–based reconstructions of past climate to infer wider changes in the atmosphere and ocean is challenging (see Section 3). Improving our understanding of the changes in the total climate system that drive glaciers requires more explicit coupling of glacier and climate models. This integration remains a challenge because of scale issues; either climate models need to be applied at the resolution of individual glaciers, a computationally expensive task requiring

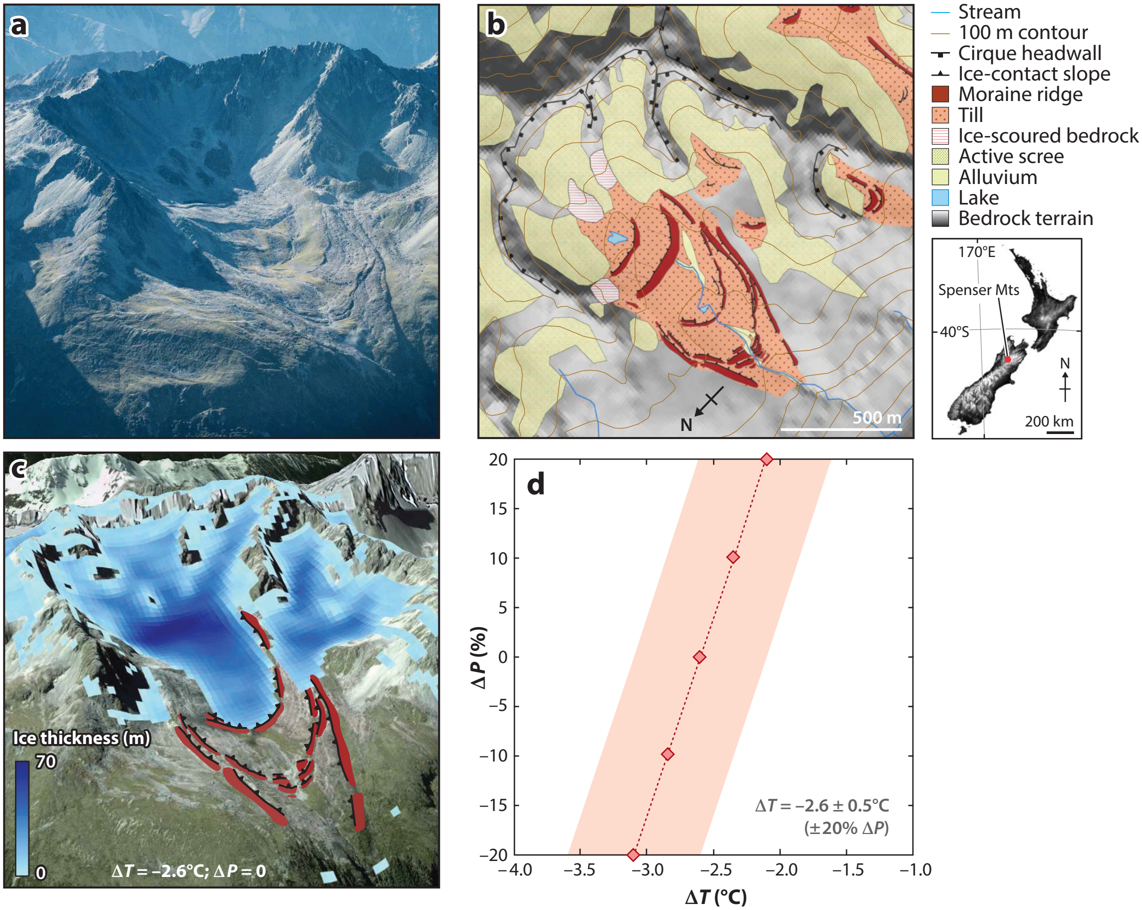

Figure 11

Example of how (*a*,*b*) a mapped moraine sequence of the Spenser Mountains, New Zealand, can be used to constrain a model-based reconstruction of past temperature. (*c*) A relatively simple ice flow model, in this case, a two-dimensional shallow ice approximation (2D-SIA) model coupled to an energy balance model (Eaves et al. 2016), is used to reconstruct (*d*) a cooling of -2.6 ± 0.5°C relative to present for moraines dating to the Late Glacial (approximately 15,000–11,000 years ago). The uncertainty in this temperature reconstruction includes imposed precipitation variability of ± 20% as well as parametric uncertainty (*pink area* in panel *d*). Abbreviations: *P*, precipitation; *T*, temperature.

well-understood boundary conditions, or climate downscaling and bias correction techniques need to be applied to coarser-resolution climate models. These issues are covered further in the next section.

7. CHALLENGES AND FUTURE DIRECTIONS IN RECONSTRUCTING CLIMATE FROM GLACIERS

Models can only be evaluated in relative terms, and their predictive value is always open to question.

—Oreskes et al. (1994, p. 641)

7.1. Disentangling Temperature and Precipitation

One of the limitations of reconstructing past climate from glaciers is the nonunique aspect of mass balance forcing. The principle of equifinality applies, because a given end state (glacier length) can be reached by many potential means (changing temperature, precipitation, and other variables that influence glacier mass balance). Multiple steady states are also possible in response to the same climate forcing, owing to nonlinearities introduced via the interactions between glaciers and their beds (e.g., McKinnon et al. 2012).

Uncertainties related to past precipitation variability are not as large as one might imagine, because temperate glaciers are more sensitive to temperature changes than any other meteorological variable (see Section 2). Possible ranges in past precipitation variability can be estimated from present-day climatology (e.g., Anderson & Mackintosh 2006) or from paleoclimate model reconstructions (e.g., Braconnot et al. 2012) (see Section 3). Repeated glacier model simulations can then be carried out using many possible combinations of past temperature and precipitation, so that our uncertainty in past precipitation can be expressed as uncertainty in the resulting temperature reconstruction. Our previous work shows that a 40% precipitation change (relative to modern levels) introduces a $\pm 0.5°C$ uncertainty in paleotemperature reconstructions, although this figure is likely to depend somewhat on the climatic setting and geometry of the glacier in question (Anderson & Mackintosh 2006, Doughty et al. 2013, Eaves et al. 2016).

The uncertainties in temperature reconstructions can potentially be reduced if precipitation levels have been independently constrained from proxy data. For example, past precipitation can be reconstructed from lake levels in simple closed drainage basins (e.g., Barth et al. 2016). If a local ice core is available, past accumulation rates can be directly constrained (e.g., Davies et al. 2014). In such situations, directly accounting for past precipitation variability (with appropriate uncertainties) in glacier modeling experiments results in tighter paleotemperature reconstructions.

7.2. The Effect of Climate Noise on Glacier Fluctuations

Glaciers can exhibit significant length changes in response to year-to-year mass balance variability within an otherwise stationary climate (e.g., Oerlemans 2001, Roe 2011) (**Table 1**). Such random walks, which reflect stochastic climate noise rather than natural climate variability or anthropogenic climate change, may render the climatic interpretation of glacier records difficult (Lüthi 2009). This behavior also makes it challenging, for example, to disentangle the effects of natural and anthropogenic climate drivers of glacier length changes over the last few decades.

The magnitude of length variations that a glacier might experience as a consequence of stochastic forcing is a function of the topographic and climatic setting, but the details of these controls are not yet well understood. Some model dependence on existing results is also likely, as studies of glacier length variability in response to stochastic variability have been carried out with a range of model types and forcings (**Table 1**). A carefully constructed model intercomparison with the full range of models and forcings is needed to tease out these differences.

To illustrate the dependence of length variability on glacier slope and mass balance forcing, we ran a set of experiments using a geometrically idealized flowline model based on equations 31 and 32 of Le Meur et al. (2004), with the only modification being that the flowline length was extended to 7,500 m to accommodate all length variations. The results (**Figure 12**) show that glacier length variability increases with larger imposed mass balance variability and steeper surface slope (which generally results in shorter glacier response time). This explains, in part, the different findings shown in **Table 1** because glaciers with the smallest response to random interannual forcing (e.g.,

Table 1 Calculated standard deviations of glacier length (σ_L) driven by white noise forcing of equilibrium line altitude (σ_E), temperature (σ_T), precipitation (σ_P), and/or mass balance (σ_M) for various glaciers

Glacier	σ_E (m)	σ_T (K)	σ_P (m)	σ_M (m w.e. a^{-1})	σ_L (m)	Model type	Source
Nigardsbreen	75	NA	NA	NA	658	Linear model	Oerlemans 2001
Nigardsbreen	NA	NA	NA	NA	360[a]	SSC/flowline model	Reichert et al. 2002
Rhonegletscher	NA	NA	NA	NA	180[a]	SSC/flowline model	Reichert et al. 2002
Nigardsbreen	NA	NA	NA	1.05	610	SSC/flowline model	Oerlemans 2001
Rhonegletscher	NA	NA	NA	0.75	240	SSC/flowline model	Oerlemans 2001
Franz Josef	NA	NA	NA	1.05	280	Flowline model	Oerlemans 2001
Quelccaya Ice Cap	NA	0.52	0.2	NA	134	EBM/flowline model	Malone et al. 2015
Cameron Glacier	NA	0.4	0.45	0.67	50	EBM/2D-SIA model	Doughty et al. 2017
Cameron Glacier	NA	0.4	0.45	0.63	190	EBM/2D-SIA model	Doughty et al. 2017
Mount Baker Glaciers	NA	0.8	1.0	NA	360	Linear model	Roe 2011
Mount Baker Glaciers	NA	0.8	1.0	NA	324	Flowline model	Roe 2011
Mount Baker Glaciers	NA	0.8	1.0	NA	314	3-stage linear model	Roe & Baker 2014
Nigardsbreen	NA	0.9	0.7	0.92	1,063[b]	Flowline model	Roe & Baker 2014
Nigardsbreen	NA	0.9	0.7	0.92	1,222	3-stage model	Roe & Baker 2014
Nigardsbreen	NA	0.9	0.7	0.92	1,501	Linear model	Roe & Baker 2014
Fall Creek Glacier	NA	NA	NA	NA	370	Flowline model	Anderson et al. 2014
Fall Creek Glacier	NA	NA	NA	NA	415	Linear model	Anderson et al. 2014
Rakaia Glacier	NA	0.8	0.9	NA	850	EBM/2D-SIA model	Rowan et al. 2014
Rakaia Glacier	NA	0.8	0.9	NA	1,200	EBM/2D-SIA model	Rowan et al. 2014

[a]Mass balance white noise forcing experiment (σ_m) not specified in article.
[b]When integrated around a mean length of 14 km as in Oerlemans (2001), Roe & Baker (2014) give $\sigma_L = 650$ m.
Abbreviations: EBM, energy balance model; m w.e. a^{-1}, meters of water equivalent per year; NA, not available; SSC, seasonal sensitivity characteristic; 2D-SIA, two-dimensional shallow ice approximation.

Malone et al. 2015, Doughty et al. 2017) are generally located in less maritime climates with lower mass balance variability and have relatively shallow bed slopes.

Figure 12 illustrates ways in which the impact of glacier length variability can be taken into consideration when reconstructing past climate from glacial geomorphology. In a given climate setting, glaciers with small bed slopes show limited length variability because of the approximately inverse relationship between bed slope and response time (smaller bed slope gives a longer response time as shown by equation 3c in Roe & Baker 2014 and equation 4 in Leclercq & Oerlemans 2012). Likewise, glaciers with small mass balance (and climate) variability also show small length variability.

Glaciers in maritime climates are likely to experience more year-to-year mass balance variability than glaciers in low-precipitation, continental climates. However, this does not, in our view, compromise their ability to be used as paleoclimate indicators. Not all short-term mass balance variability is noise, and maritime glaciers are influenced by natural modes of climate variability such as the El Niño Southern Oscillation, which operates over a timescale of 1–5 years. For example,

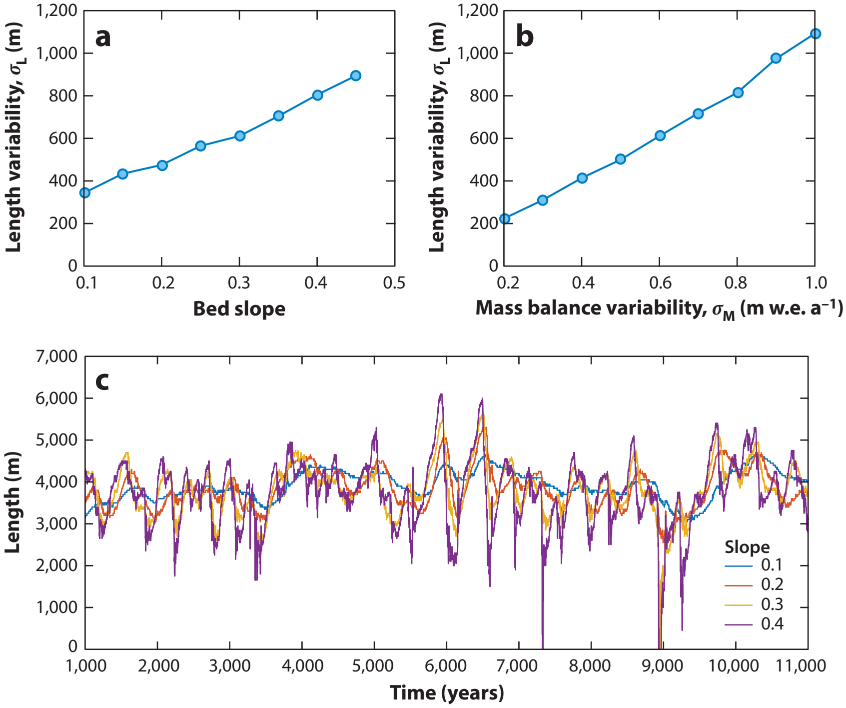

Figure 12

Length variability of glaciers depends on (*a*) glacier bed slope and (*b*) imposed mass balance variability, which is a function of the climate forcing. Panel *c* shows the dependency of glacier length variability on bed slope (slope ranges 0.1–0.4). To calculate the variability values for each data point, the flowline model is integrated for 11,000 years, and the first 1,000 years is discarded. Abbreviation: m w.e. a^{-1}, meters of water equivalent per year.

Franz Josef Glacier, a fast-responding maritime glacier in New Zealand, recently experienced a 25-year period of advance from 1983 to 2008 (**Figure 3**) as a consequence of regional-scale variability in atmospheric temperature and precipitation (Mackintosh et al. 2017). Such high-frequency glacier fluctuations and associated moraine records (if preserved; see **Figures 9** and **10**) provide information about the climate system that is additional and complementary to the smoothed-out response provided by slow-responding glaciers in more continental settings.

Over longer timescales, it is unlikely that random climate variability is a significant driver of change for the majority of glaciers. A glacier does not act like a reservoir subject to filling by randomly fluctuating input. Rather, glacier length corresponds to a steady mass forcing of any given magnitude, and the response time determines how quickly this equilibrium is approached. The glacier acts more like a low-pass filter, with length fluctuations driven by fluctuations in the average of mass balance over the response time, rather than the sum.

Observations and moraine records documenting glacier response to climate change support this interpretation. For example, the clear secular trend in glacier length of hundreds of glaciers during the last few centuries (Zemp et al. 2015) (see Section 1 and **Figure 3**) shows that glaciers dominantly responded to climate warming (Oerlemans 2005, Roe et al. 2017) rather than to random year-to-year variability (or regional variations in precipitation) during this period. The large-scale retreat of glaciers in both hemispheres following the Last Glacial Maximum provides another example of a clear global climatic response (e.g., Schaefer et al. 2006, Shakun et al. 2015).

7.3. Glacier Simulations Driven by Climate Models

Although it is now common to drive ice sheet models with climate models, it is still relatively novel for glaciers. The problem is that the spatial resolution of most climate models is not fine enough to resolve glaciers. Temperature and (especially) precipitation fields cannot be directly used as input to glacier mass balance models without some kind of downscaling and/or bias correction (e.g., Jarosch et al. 2012). Owing to the computational costs involved, only a few studies have simulated glacier mass balance directly using output from very high-resolution climate models (e.g., Mölg & Kaser 2011). A practical approach has been to combine temperature downscaling using a temperature lapse rate or statistical relationships, which can then be used to force a degree-day model (e.g., Bravo et al. 2015), with up-slope modeling of snow accumulation using an orographic precipitation model (Clarke et al. 2015, Kessler et al. 2006) (see Section 3).

Ultimately, driving glaciers with climate models, or in some cases developing fully coupled climate-glacier models, will have clear benefits: (*a*) climate models provide insight into the physical climate processes that force glaciers; (*b*) estimating climatic parameters such as former temperature or precipitation lapse rates will not be needed in coupled climate-glacier modeling experiments; (*c*) fully coupled climate-glacier models allow the boundary-layer effects associated with glacier changes to be fed back into climate models (e.g., Collier et al. 2015); (*d*) transient climate modeling experiments (e.g., He et al. 2013) can be used to force glacier simulations through time; (*e*) the availability of well-dated landscape and sedimentary records to compare with simulated glacier fluctuations provides an additional means for evaluating climate model performance, which may be particularly helpful in situations in which paleoclimate models show divergent behavior (e.g., Rojas et al. 2009); and (*f*) driving glaciers with large-scale climate models forced with specific hypothesized climate change mechanisms alleviates the difficulties regarding the behavior of mid-latitude lapse rates discussed in Section 3, especially when done with multimodel ensembles.

7.4. Global-Scale Glacier Reconstructions: Constraints on Climate Sensitivity and Polar Amplification of Climate?

In recent years, several groups have compiled global databases of ^{10}Be ages from terrestrial glacial sequences. These data help to constrain, for example, the timing of the Last Glacial Maximum (Clark et al. 2009) or the timing of glacier retreat with respect to temperature (Denton et al. 2010) and CO_2 forcing (Shakun et al. 2015). There remains, however, the potential to use these datasets to reconstruct past global temperature changes using glacier models. Glacier models provide the unifying physical framework required to carry out this reconstruction. If one type of glacier model is used, or if model-to-model differences are well understood, then it should be possible to reconstruct past temperature at the Last Glacial Maximum (and for other time periods) in all of the world's glaciated mountain ranges.

With the caveats of relating glacier-derived past temperatures to global climate kept in mind, we believe that a global-scale reconstruction of past temperature from glaciers at the Last Glacial Maximum would be useful for a number of reasons. (*a*) It would provide an independent check on the fidelity of other terrestrial (e.g., Bartlein et al. 2011) and marine (e.g., MARGO 2009) temperature reconstructions, much like early ELA reconstructions helped to assess the quality of tropical SST reconstructions (Rind & Peteet 1985). (*b*) It would provide an ideal dataset for evaluating climate model simulations (e.g., Braconnot et al. 2012). This approach is particularly valuable because glaciers tend to form in areas where we have data gaps in climate proxy records—either at high elevation or in remote parts of the world. (*c*) It would provide further constraint on the Earth's climate sensitivity to CO_2 forcing (e.g., Hansen et al. 2013, Schmittner et al. 2011).

(*d*) Reconstructions from mountain ranges such as the Andes would help to assess how Last Glacial Maximum temperature varied with latitude. This assessment is important because we have very few paleoclimate constraints on polar amplification of climate from the southern midlatitudes to high latitudes (IPCC 2013).

8. SUMMARY

Glaciers are sensitive environmental indicators, and they advance and retreat in response to a changing climate. They are relatively simple physical systems that can be described by a set of equations, which can be easily implemented in a numerical model. After appropriate evaluation, glacier models can be powerful tools for reconstructing past climate from dated moraine sequences and other geological data. Research groups who bring together high-precision dating of glacier landform systems, modern observations on glaciers, and numerical modeling of glaciers have the potential to significantly improve our understanding of the climate system. This research might include providing tighter constraints on the climate sensitivity of the Earth to CO_2 forcing as well as polar amplification of the climate system.

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Errata

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