Impact of glacier loss on annual basin water yields

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Abstract. We couple a glacier flow model to a simplified landscape model to investigate the effects of glacier dynamics, climate, and vegetation succession on annual basin runoff during glacier retreat. Basin runoff initially increases as water is released from glacier storage but eventually decreases to below preretreat levels due to increases in evapotranspiration and altitudinal losses in precipitation. Peak basin runoff and the time to peak basin runoff are primarily determined by glacier

5   dynamics, with shallow sloping continental glaciers experiencing the largest increases in basin runoff (up to 62%) and longest time until peak basin runoff (up to 142 years), compared to 14% and 54 years for steep maritime glaciers subjected to the same rate of climate change. These differences in peak basin runoff and time to peak basin runoff can be characterized by the glacier response time: glaciers with long response times are pushed farther out of equilibrium for a given climate forcing and produce larger variations in basin runoff than glaciers with short response times. After peak basin runoff is reached, vegetation plays an

10   increasingly important role, with basin runoff decreasing considerably faster for heavily vegetated landscapes than for rocky landscapes and ultimately reaching values that are over 50% lower than preretreat levels. Our results demonstrate that glacier dynamics and landscape evolution should receive roughly equal attention when assessing the impacts of glacier mass loss on water resources.

1   Introduction

15   Glacier runoff is a dominant control on the timing and magnitude of runoff from glacierized watersheds (Hock, 2005). Short term water storage within glaciers impacts the diurnal characteristics of runoff, while intermediate term storage influences the seasonality of runoff by heavily concentrating runoff in summer months (Jansson et al., 2003) when glaciers can provide a substantial portion of streamflow even at very low levels of catchment glacierization (Stahl and Moore, 2006; Nolin et al., 2010; Huss, 2011). On annual time scales, basin runoff from glacierized watersheds can be more than double that of comparable

20   nonglacierized watersheds (Hood and Scott, 2008).

Globally, more than a billion people live in watersheds that receive runoff from glaciers (Kaser et al., 2010). Within these watersheds, glacier runoff provides a wide variety of ecosystem services, including agricultural and municipal water supplies, hydroelectric power generation, stream temperature modulation, biodiversity, and fisheries (Milner et al., 2017; Cheesbrough et al., 2009; Gaudard et al., 2016; Fellman et al., 2014; Dorava and Milner, 2000). Moreover, changes in runoff from glaciers

25   have wide ranging implications for the ecological function of downstream aquatic ecosystems (Milner et al., 2009; Jacobsen et al., 2012). As a result, developing a quantitative understanding of how runoff from glaciers and their watersheds will be

altered as glaciers continue to thin and recede is critical for predicting how the ecosystem services associated with glacier runoff will change in the future. Glacier runoff is controlled by the energy balance at the glacier surface, and as a result it is highly vulnerable to future climate warming compared to other components of the terrestrial water budget.

As watersheds undergo deglacierization, annual water yields are hypothesized to show a transient increase followed by a decrease to a new, lower baseline value as glaciers are lost (e.g. Jansson et al., 2003; Moore et al., 2009). The magnitude of this change in basin water output can be substantial. For example, annual runoff from the Hofsjökull and southern Vatnajökull ice caps in Iceland is expected to increase roughly 50 percent by the end of the century (Aðalgeirsdóttir et al., 2006). In contrast, late summer water yields in glacierized basins in British Columbia demonstrate widespread negative trends in recent decades (Stahl and Moore, 2006). The direction of the glacier runoff driven change in water yields is roughly a function of watershed glacier coverage with increasing water yields in heavily glacierized basins and decreasing water yields in catchments with diminished glacier coverage ($\lesssim$10%; Casassa et al., 2009). On a global scale, this trend is reflected in the fact that regional glacier runoff is projected to increase in the Arctic, Canada, and Russia and decrease sharply in lower latitude mountain basins in Asia, Europe, and South America (Bliss et al., 2014).

Previous efforts to evaluate the impact of glacier loss on water yields have focused on measuring and modeling individual catchments (Moore et al., 2009; Huss et al., 2008; Huss, 2011; Stahl and Moore, 2006) and regions (Stahl and Moore, 2006; Huss and Hock, 2018; Nolin et al., 2010; Baraer et al., 2012). While valuable, these case studies do not elucidate the broader geomorphological and glaciological controls that govern the hydrological responses of watersheds to ongoing glacier recession. A number of studies have also focused on glacier runoff (e.g. Bliss et al., 2014), the amount of discharge from the receding glacier terminus, rather than basin runoff from a fixed gauging station. The latter is critical from a water resources standpoint because of the static nature of hydroelectric and water collection infrastructure. Efforts to understand how glacier change will impact streamflow at the basin scale are also confounded by the fact that recently deglaciated landscapes are eco-hydrologically dynamic as a result of changes in evapotranspiration associated with recolonization of vegetation following the loss of glacier ice.

Here, we couple a glacier flow model to a simplified landscape model in order to evaluate how glacier recession and subsequent vegetation succession will impact the timing and magnitude of annual basin runoff from glacierized watersheds. To investigate how differences in the morphology and climate of glacierized basins influence hydrological responses to glacier loss, we vary a variety of parameters within our model including climate regime (maritime vs. continental), basin slope, vegetation rates, and vegetation types. Our findings provide insight into the fundamental controls on the hydrologic response of glacierized basins to a changing climate.

2 Methods

Assuming that non-glacier changes in water storage within a basin are negligible on annual timescales, the annual basin runoff at the watershed outlet, Q_s, is given by

$$Q_s = Q_g + Q_n = (P_g - Q_b) + (P_n - ET),\tag{1}$$
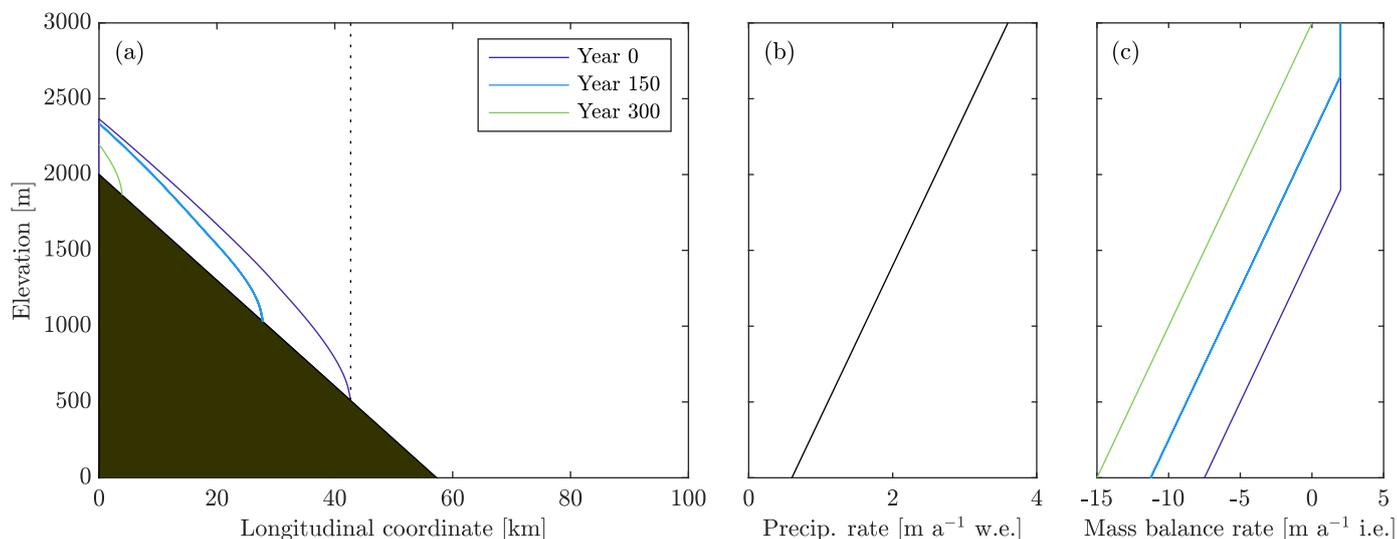
Figure 1. (a) Glacier thickness profile at various stages of glacier recession with a 2° basin slope, continental climate, and RCP8.5 climate scenario. The dotted vertical line demarcates basin extent. (b) Precipitation rate with altitude, which is held constant throughout glacier retreat. (c) Specific mass balance rate with altitude at various times during glacier recession.

where Q_g and Q_n are the glacier runoff (definition #5 in O'Neel et al., 2014) and nonglacier runoff, P_g and P_n are the precipitation fluxes (solid plus liquid) into the glaciated and nonglaciated portions of the basin, Q_b is the glacier-wide mass balance flux (accumulation minus ablation), and ET is the evapotranspiration flux. For consistency, all fluxes are expressed in water equivalent units.

5   We calculate the runoff components in Equation (1) with a depth-integrated glacier flow model and a simplified landscape model. We assume that the precipitation and mass balance rates depend on elevation and that the evapotranspiration rates are a function of time since deglaciation. The glacier flow model adjusts the elevation and length/surface area of the glacier in response to the glacier's mass balance, and the landscape model tracks the evolution of the deglaciated landscape.

The model domain consists of a parallel-sided valley that has a constant width of 4000 m and a constant downvalley slope,
10   with the bedrock reaching a peak elevation of 2000 m (Fig. 1a). The glacier is assumed to flow from an ice divide at the upper reaches of the valley, to span the width of the valley at all times, and to initially fill the entire length of the valley. Thus, the model results tend to overemphasize the relative importance of glacier runoff on basin runoff. While simplified, the model is fast, making it possible to run numerous simulations for various parameter combinations. In particular, we explore the effect that slope, vegetation rates, and vegetation types have on basin runoff over decadal time-scales under two different climate
15   types (maritime vs. continental) and two different climate change scenarios.

2.1 Precipitation

Precipitation rates are needed for calculating both the glacier and nonglacier runoff. We assume that precipitation varies linearly with altitude, such that

$$\dot{P}(z) = \dot{P}_0 + \frac{\mathrm{d}\dot{P}}{\mathrm{d}z} z, \tag{2}$$

5   where $\dot{P}$ is the width-averaged precipitation rate and $\dot{P}_0$ is the precipitation rate at sea level (Fig. 1b). For all simulations we set $\mathrm{d}\dot{P}/\mathrm{d}z = 0.001\ \mathrm{a}^{-1}$ (Immerzeel et al., 2015). The precipitation at sea level is chosen to ensure that the precipitation at elevation always exceeds glacier accumulation rates (see Section 2.4).

2.2 Glacier model and glacier runoff

Glacier runoff is calculated by integrating the precipitation and specific surface mass balance rates over the glacier surface, i.e.,

$$Q_g = P_g - Q_b = \int_{\Omega_g} \left(\dot{P} - \frac{\rho_i}{\rho_w} \dot{B} \right) \mathrm{d}\Omega_g, \tag{3}$$

where $\rho_i = 917\ \mathrm{kg\ m^{-3}}$ and $\rho_w = 1000\ \mathrm{kg\ m^{-3}}$ are the densities of ice and water, $\dot{B}$ is the width-averaged specific surface mass balance rate (in units of ice equivalent) and Ω_g is the glacier surface area (in map view). The precipitation rate is given in Equation (2), and the balance rate is prescribed by using a constant mass balance gradient and imposing a maximum balance

15   rate, $\dot{B}_{\mathrm{max}}$. In other words,

$$\dot{B} = \min \left(\frac{\mathrm{d}\dot{B}}{\mathrm{d}z}(z - \mathrm{ELA}), \dot{B}_{\mathrm{max}} \right), \tag{4}$$

where ELA is the elevation of the equilibrium line altitude (Fig. 1c). We use an initial ELA of 1500 m, consistent with high and mid-latitude glaciers (Huss and Hock, 2015). In our simulations, we vary the climate type by adjusting the balance gradient and the maximum balance rate (e.g., maritime glaciers have high balance gradients and high accumulation rates) and parameterize

20   climate change by varying the ELA (see Section 2.4).

From Equation (3) it is clear that glacier runoff depends on glacier hypsometry, which evolves in response to the glacier mass balance. To model the evolution of glacier hypsometry, we invoke a commonly used one-dimensional, depth- and width-integrated flow model (see Fig. 1a for example longitudinal cross-section; Nick et al., 2009; Enderlin et al., 2013). The model is based on conservation of momentum, which requires that the glaciological driving stress is balanced by gradients in longi-

25   tudinal stress, lateral drag, and basal drag (van der Veen, 2013), such that

$$2 \frac{\partial}{\partial x} \left(H \nu \frac{\partial U}{\partial x} \right) - \frac{H}{W} \left(\frac{5U}{2AW} \right)^{1/3} - \tau_b = \rho_i g H \frac{\partial h}{\partial x}, \tag{5}$$

where H is ice thickness, ν is the depth- and width-averaged viscosity, U is the depth- and width-averaged velocity, W is glacier width, A is the flow rate factor, g is gravitational acceleration, and τ_b is the basal shear stress. The viscosity depends on

the strain rate:

$$\nu = A^{-1/3} \left| \frac{\partial U}{\partial x} \right|^{2/3}. \tag{6}$$

We assume a constant flow rate factor of $A = 2.4 \times 10^{-24}$ Pa^{-3}s^{-1}, consistent with that of temperate ice (Cuffey and Paterson, 2010), and a constant basal shear stress of 10^5 Pa as is commonly observed beneath valley glaciers (e.g., Brædstrup et al., 2016). Thus, we assume that the basal shear stress is at the yield stress for ice (Cuffey and Paterson, 2010). A velocity of $U = 0$ is prescribed at the ice divide ($x = 0$), and a velocity gradient is applied at the terminus by inserting the depth-averaged deviatoric stress into Glen's Flow Law. The latter is necessary because in the model the ice must maintain some finite thickness at the terminus. Finally, after each time step the glacier surface is updated with a depth- and width-integrated mass continuity equation (van der Veen, 2013), in which

$$\frac{\partial H}{\partial t} = \dot{B} - \frac{1}{W} \frac{\partial (UHW)}{\partial x}, \tag{7}$$

and the glacier length is updated by removing any ice from the terminus that is thinner than 0.1 m.

2.3 Landscape model and nonglacier runoff

The nonglacier runoff is calculated by assuming that the evapotranspiration is some fraction of the precipitation, such that

$$Q_n = P_n - ET = \int_{\Omega_n} C \dot{P} \, \mathrm{d}\Omega_n, \tag{8}$$

where $0 \leq C \leq 1$ is the runoff ratio (the ratio of precipitation to runoff over an area of land) and Ω_n is the area of the deglaciated landscape. The runoff ratio of a particular deglaciated area will vary based on the time since deglaciation. Thus, in order to calculate the nonglacier runoff, our landscape model tracks the area exposed during glacier retreat as well as changes in the surface cover as it transitions through progressively more vegetated surface types (Crocker and Major, 1955; Burga et al., 2010; Chapin et al., 1994). The model is based on two simple assumptions. First, we assume that the catchment becomes increasingly vegetated following deglaciation and that the type of vegetation only depends on time since deglaciation. Second, as areas of the catchment become colonized, the rate at which water is evapotranspired increases until reaching a maximum value representative of the climax vegetation state.

We express the change in landscape cover that occurs during plant colonization through a step-wise parameterization of the runoff ratio. Runoff ratios range from 0.5 (forest) to close to 1 (ice) and depend on the vegetation type (Zhang et al., 2001; Filoso et al., 2017). We parameterize vegetation type using four runoff ratios, such that

$$C_i = \begin{cases} C_1 & 0 \leq t \leq T_1 \\ C_2 & T_1 < t \leq T_2 \\ C_3 & T_2 < t \leq T_3 \\ C_4 & t > T_3 \end{cases}, \tag{9}$$

where C_i is the runoff ratio associated with the vegetation type i, $t = 0$ is the time at which a portion of the catchment is deglaciated, and T_i indicates the time at which there is a transition in surface type. Thus, the runoff ratio is a function of time but varies spatially, and consequently Equation (8) can alternatively be expressed as

$$Q_n = \sum_{i=1}^{4} C_i \int_{\Omega_{n_i}} \dot{P} \, \mathrm{d}\Omega_{n_i}, \tag{10}$$

5   where Ω_{n_i} represents the nonglacier surface area that has runoff ratio C_i.

2.4 Simulations

We use our model to test the effect that bed slope, vegetation type, and vegetation rates have on basin runoff for two different climate types and two different climate change scenarios by considering a range of parameter values. We varied the bed slope of the basin (including beneath the glacier) from shallow ($2°$) to steep ($10°$) and used six different sets of four runoff

10   ratios, $C = \{C_1, C_2, C_3, C_4\}$, and six different sets of three vegetation timings, $T = \{T_1, T_2, T_3\}$ (see Eq. 9). The runoff ratios ranged from a rocky high elevation or latitude environment with no vegetation, $C = \{1, 1, 1, 1\}$, to a low elevation or latitude environment with heavy vegetation, $C = \{0.95, 0.8, 0.7, 0.5\}$, and the vegetation rate ranged from rapid, $T = (5\,\mathrm{a}, 10\,\mathrm{a}, 25\,\mathrm{a})$, to slow, $T = (50\,\mathrm{a}, 100\,\mathrm{a}, 250\,\mathrm{a})$. With the exception of the no vegetation scenario, the runoff ratio always decreased with time since deglaciation, consistent with increased evapotranspiration associated with vegetation succession. Consequently, the

15   runoff ratio decreases in the downvalley direction until the landscape has reached climax vegetation. Our model vegetation types and their corresponding runoff ratios span the range of reported values for the process of vegetation succession following glacier retreat, which can be highly spatially variable even within a given climate (e.g., Crocker and Major, 1955; Burga et al., 2010; Chapin et al., 1994).

The two climate types that we define are designed to roughly mimic the climates that are experienced by maritime and

20   continental glaciers. The climates are defined by a glacier's surface mass balance gradient and maximum surface mass balance (Eq. 4) and by the precipitation rate at sea level (Eq. 2). For the maritime climate, we set $\mathrm{d}\dot{B}/\mathrm{d}z = 0.01\,\mathrm{a}^{-1}$, $\dot{B}_{\max} = 4\,\mathrm{m\,a}^{-1}$, and $\dot{P}_0 = 2.4\,\mathrm{m\,a}^{-1}$, whereas for the continental climate these values are $\mathrm{d}\dot{B}/\mathrm{d}z = 0.005\,\mathrm{a}^{-1}$, $\dot{B}_{\max} = 2\,\mathrm{m\,a}^{-1}$, and $\dot{P}_0 = 0.55\,\mathrm{m\,a}^{-1}$ (see Cuffey and Paterson, 2010, for example mass balance curves); note that the mass balance curves are defined in units of ice equivalent while the precipitation curves are defined in units of water equivalent.

25   In each simulation, a constant climate is used to spinup the model to a steady-state, defined as being reached when the rate of terminus advance or retreat is less than $2\,\mathrm{m\,a}^{-1}$. After reaching steady-state, the climate is changed by steadily raising the ELA (e.g., Fig. 1c). We consider two climate change scenarios that are roughly based on expected changes in ELA (Huss and Hock, 2015) under two different Representative Concentration Pathways (RCP2.6 and RCP8.5, which correspond to increases in radiative forcing of 2.6 and 8.5 W/m^2). In the RCP2.6 scenario we prescribe the ELA (in meters) according to

30    $$\mathrm{ELA} = 1500 + 158(1 - e^{-t/28}), \tag{11}$$
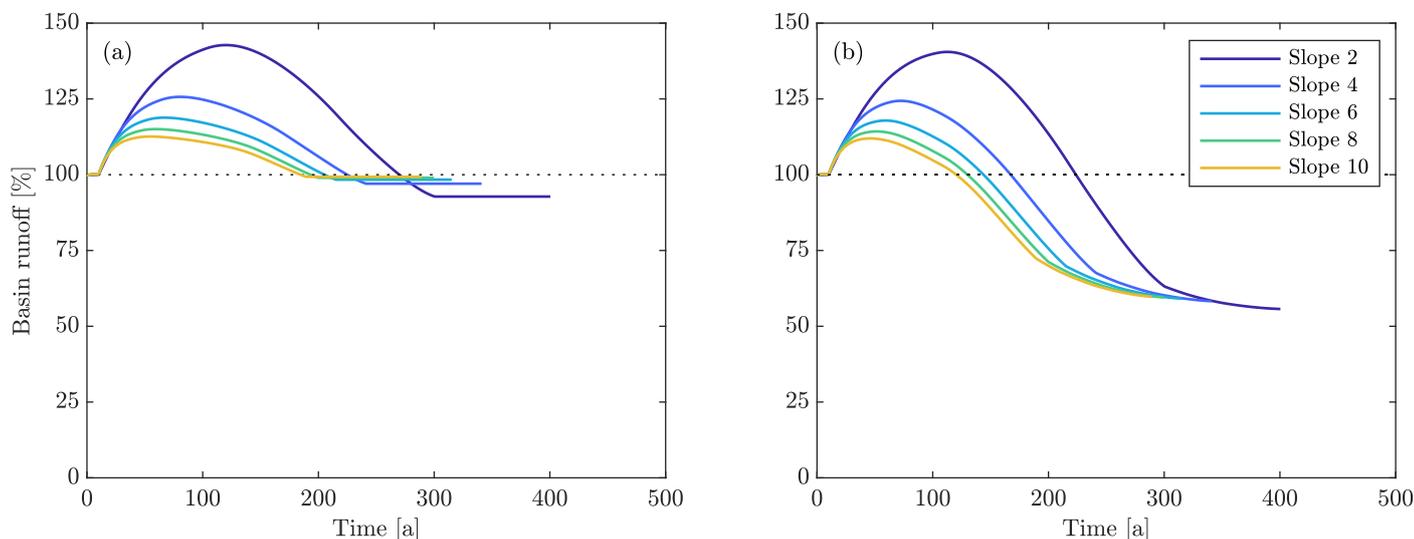
Figure 2. Variations in basin runoff (a) without vegetation and (b) with vegetation. For both (a) and (b) we use a maritime climate and the RCP8.5 climate change scenario. In (b) we set $T = \{20, 50, 100\}$ and $C = \{1, 0.9, 0.8, 0.6\}$.

resulting in an asymptotic increase in the ELA of about 150 m over a 100 year period. In contrast, during the RCP8.5 scenario we raise the ELA linearly with time:

$$\text{ELA} = 1500 + 5t. \tag{12}$$

We hold the climate constant for the first 10 years of the model run before initializing changes in the ELA according to Equations 11 and 12. Note that we neglect any temporal variations in precipitation and are assuming that changes in glacier mass balance are primarily due to warming (Fig. 1c; Van de Wal and Wild, 2001).

During the simulations the basin is initially filled with ice (i.e., the length of the basin is defined as the initial steady-state length of the glacier and thus the basins do not have the same lengths). As the glacier recedes the portion of vegetated area increases and previously exposed portions mature, moving through progressive vegetation types. The simulations continue until the glacier has reached a new steady-state or disappeared altogether and the newly exposed landscape has reached the final vegetation state.

3 Results

3.1 Basin runoff

Following the initiation of climate change in our simulations, glaciers retreat and thin steadily until either disappearing completely (for the RCP8.5 scenarios) or reaching a new steady-state (for the RCP2.6 scenarios). Glaciers in the RCP2.6 climate scenario lose between 16–25% of their area and 19–26% of their volume before reaching a new steady state, with steep glaciers
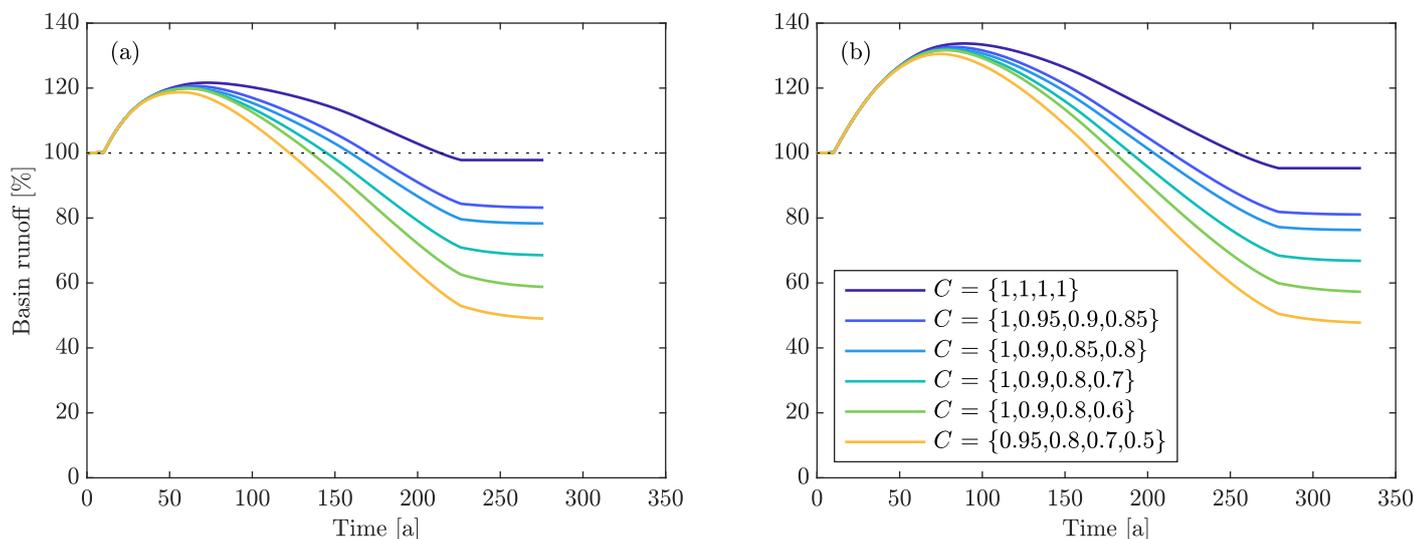
Figure 3. Variations in basin runoff for differing types of catchment vegetation in (a) maritime and (b) continental climates. The basin slope ($5°$), climate change scenario (RCP8.5), and vegetation timing ($T = \{15a, 30a, 50a\}$) are the same in both panels.

showing higher fractional volume and area losses. For a given slope, fractional volume and area changes are similar (within ~1%) between continental and maritime climates. As glaciers retreat, basin runoff in the maritime climate under the RCP8.5 scenario experiences a transient increase of about 10–40% over a time period of 20–100 years (Fig. 2), with shallow sloped basins experiencing substantially higher and later peak basin runoff than steep basins. Basin runoff subsequently decreases over the next 100–200 years.

For simulations that include no vegetation, end basin runoff (the final steady-state runoff) is slightly below preretreat levels due to orographic losses in precipitation since glacier volume loss results in a decrease in basin elevation (Fig. 2a). This result is more pronounced for shallow basins than for steep basins (-14% vs -1% for the RCP8.5 scenario) because shallow sloped basins contain longer, thicker glaciers that undergo more surface lowering. When vegetation is included, end basin runoff can fall below 50% of the preretreat basin runoff (e.g., Fig. 3). Increases in evapotranspiration partially offset increases in basin runoff driven by glacier volume loss, although this effect is small compared to the impact of vegetation on end basin runoff (Fig. 3). Vegetation rate has no impact on the magnitude of end basin runoff because the vegetation eventually reaches a mature state regardless of the rate of change (Fig. 4). Overall, the magnitude and timing of peak basin runoff as well as the magnitude of end basin runoff are smallest when vegetation occurs rapidly and progresses to a heavily forested state (low runoff ratio).

Variations in annual basin runoff also depend on climate type. Continental basins, i.e., those with low precipitation rates and mass balance gradients, experience peak basin runoffs that are about 10–20% higher and 10–20 years later during RCP8.5 (with a greater difference for shallow basins) than comparable maritime basins, regardless of vegetation type (Fig. 3) or vegetation rate (Fig. 4). Changing from a maritime basin to a continental basin under RCP2.6 has a similar trend in effect as during RCP8.5, except with smaller changes in magnitude and timing. Continental basins have ~3% higher peak basin runoff and 7–
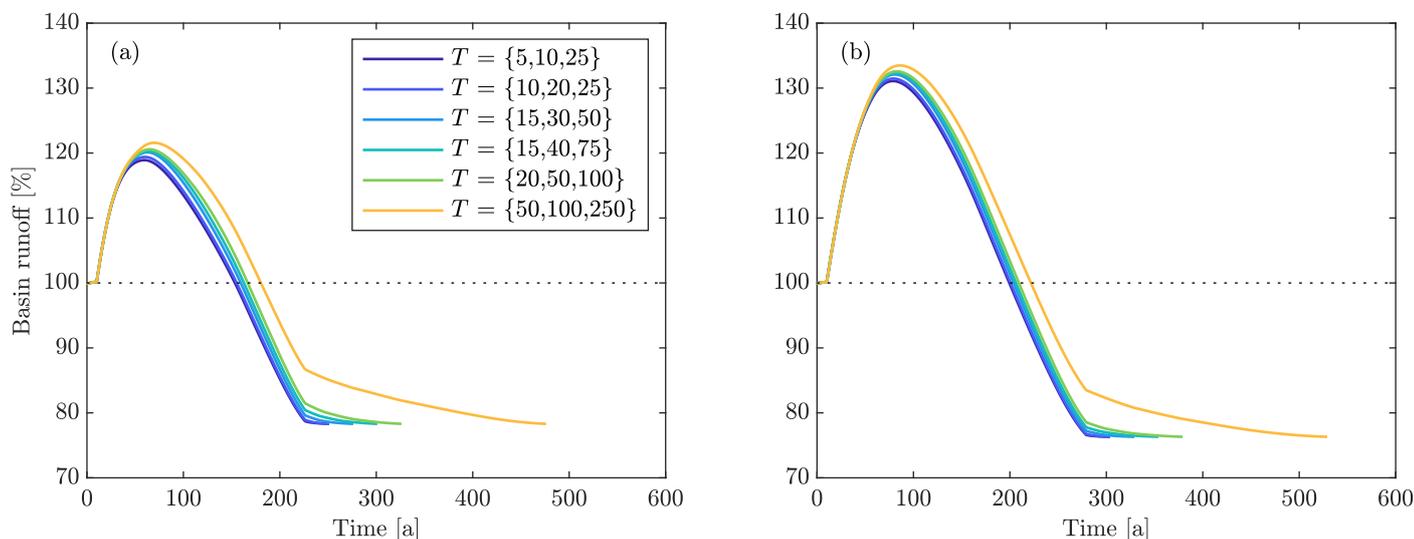
Figure 4. Variations in basin runoff for differing rates of catchment vegetation in (a) maritime and (b) continental climates. The basin slope (5°), climate change scenario (RCP8.5), and vegetation type ($C = \{1, 0.95, 0.85, 0.8\}$) are the same in both panels.

15 years later peak runoff than comparable maritime basins during RCP2.6 (data not shown). Thus, changing from a maritime climate to a continental climate has a comparable effect to decreasing the slope of a basin. The dependence of peak basin runoff on slope and climate type and is related to the glacier response to climate change, which we discuss in Section 4.

To further quantify the effects of model parameters on basin runoff, we characterize the impacts of glacier recession on basin
5 runoff with four key metrics: peak basin runoff, time to peak basin runoff, time to preretreat basin runoff, and end basin runoff. We evaluate the relative effect that each parameter has on basin runoff by selecting a canonical set of parameters, and then varying each parameter individually around that parameter set. For the canonical set we use a maritime climate, basin slope of 5°, vegetation type of $C = \{1, 0.9, 0.8, 0.6\}$, and vegetation timing of $T = \{15\,\mathrm{a}, 30\,\mathrm{a}, 50\,\mathrm{a}\}$.

Within a given climate regime, the magnitude and timing of peak basin runoff is most strongly influenced by basin slope
10 (Fig. 5a, b). In the RCP8.5 scenario, peak basin runoff from shallow basins is ∼30% higher and occurs 70 years later than for steep basins. The timing and extent of catchment vegetation have minimal impact on peak basin runoff timing and magnitude, due to the limited amount of newly vegetated land that is present at peak basin runoff, regardless of the vegetation scenario. The vegetation type and vegetation rate become increasingly important as time progresses. The time to preretreat basin runoff is affected almost equally by slope and vegetation type, with vegetation rate playing a smaller but still substantial role (Fig.
15 5c). The end basin runoff is almost entirely dependent on the runoff ratio of the final vegetation state (Fig. 5d).

Variations in model parameters consistently have a smaller impact on model output in the RCP2.6 scenario than in the RCP8.5 scenario but exhibit similar trends. This is consistent with the fact that the complete loss of glacier ice in the RCP8.5 scenario has a larger range of hydrologic impacts compared to the partial loss of glacier ice associated with the RCP2.6 scenario. One exception is the effect of varying slope on end basin runoff, which has the opposite trend for the RCP8.5 scenario than
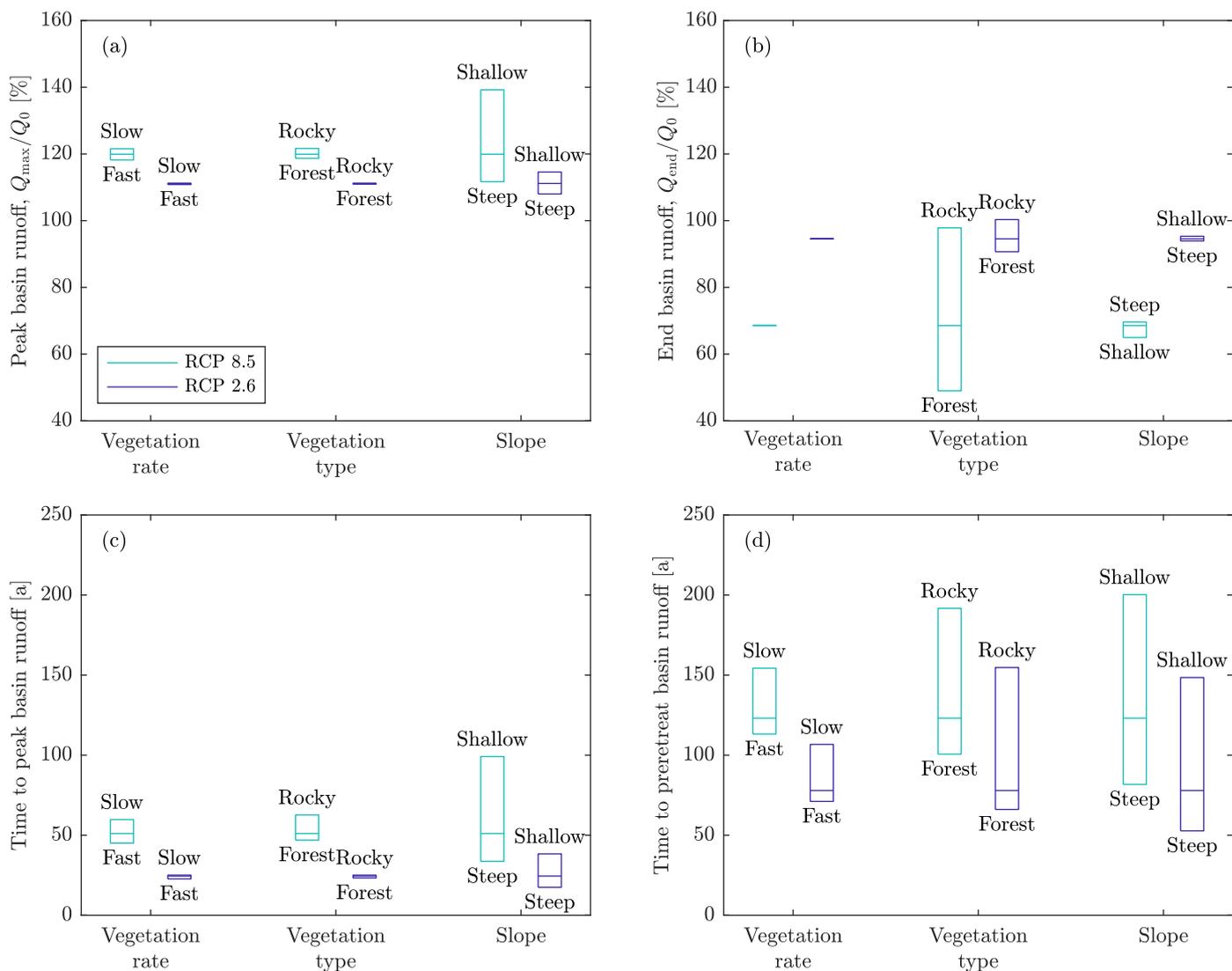
Figure 5. Influence of catchment vegetation and basin slope on four key basin hydrologic metrics: (a) peak basin runoff, (b) end basin runoff, (c) time to peak basin runoff, and (d) time to preretreat basin runoff. The line within each box represents the value for the canonical simulation (maritime climate, basin slope of $5°$, $C = \{1, 0.9, 0.8, 0.6\}$, and $T = \{15\,\mathrm{a}, 30\,\mathrm{a}, 50\,\mathrm{a}\}$). Boxes represent the range of influence for change in an individual parameter, and the direction of change for vegetation rate (slow to fast), vegetation type (unvegetated/rocky to forest), and basin bed slope (shallow to steep) are noted for each box.
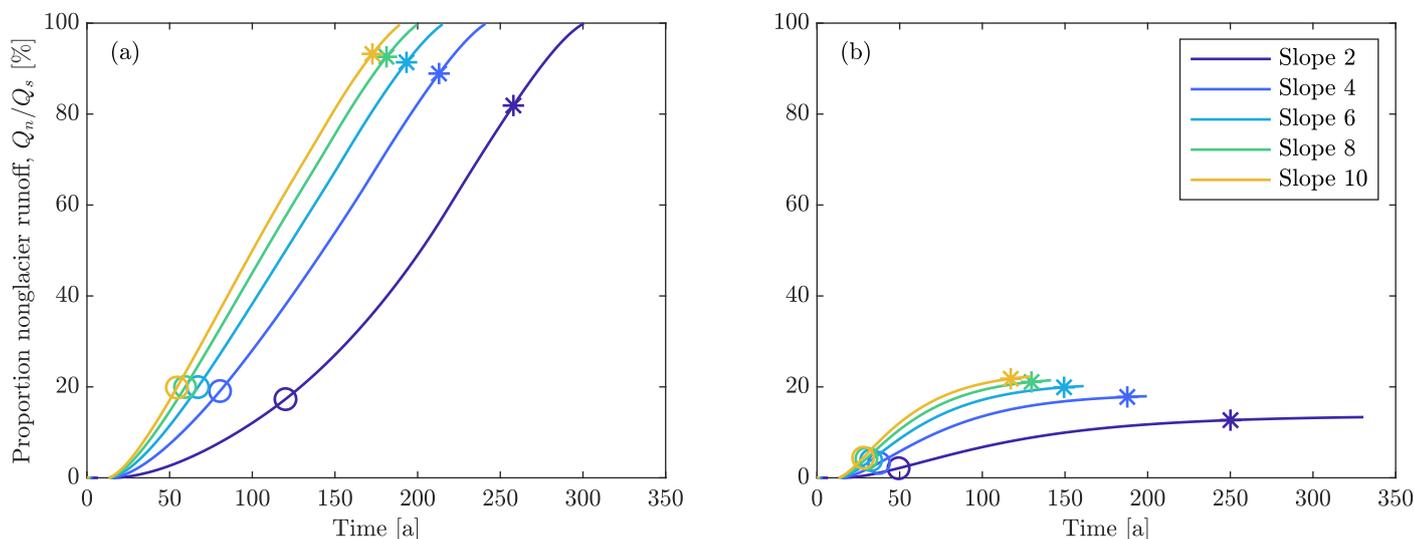
Figure 6. Percent contribution of nonglacier runoff to basin runoff without vegetation for a maritime climate during (a) RCP8.5 and (b) RCP2.6 climate change scenarios. Circles and asterisks indicate the time to peak basin runoff and to return to preretreat levels of basin runoff, respectively.

for RCP2.6 scenario (Fig. 5d). In all simulations, glacier thinning causes a decrease in orographic precipitation that is most pronounced in shallow basins. In the RCP8.5 simulations where glaciers disappear completely, this process determines the impact of slope on end basin runoff. The situation is different for the RCP2.6 simulations because the glaciers do not disappear. There, the steep glaciers experience a larger fractional retreat than the shallow sloping glaciers, which exposes more land for vegetation and ultimately results in a lower end basin runoff than for shallow basins.

Overall, peak basin runoff, time to peak basin runoff, and time to preretreat basin runoff are all smaller in the RCP2.6 scenario than in the RCP8.5 scenario. Of the four key hydrologic metrics, only the end basin runoff is higher in the RCP2.6 scenario, and this occurs because the basins do not fully deglaciate in that climate scenario. The results are similar for a continental climate, but with slightly longer times and larger peak basin runoff (data not shown).

3.2 Glacier and nonglacier runoff

In our simulations, peak basin runoff and time to peak basin runoff are most strongly influenced by basin slope and climate type (Fig. 5a, c), whereas landscape evolution plays an increasingly important role in the later stages of retreat (Fig. 5b, d). The largest possible contribution of nonglacier runoff to total basin runoff occurs when vegetation and associated evapotranspiration is assumed to be negligible ($C = \{1, 1, 1, 1\}$). Under these conditions, nonglacier runoff contributes 10–20% (RCP8.5; Fig. 6a) and 1–5% (RCP2.6; Fig. 6b), of the basin runoff during peak basin runoff. By the time that basin runoff has returned to preretreat levels, the contribution from nonglacier runoff has increased to 70–95% and 9–22%, for RCP8.5 and RCP2.6 respectively (lower bounds of peak basin runoff and preretreat basin runoff are from the continental simulations and are not
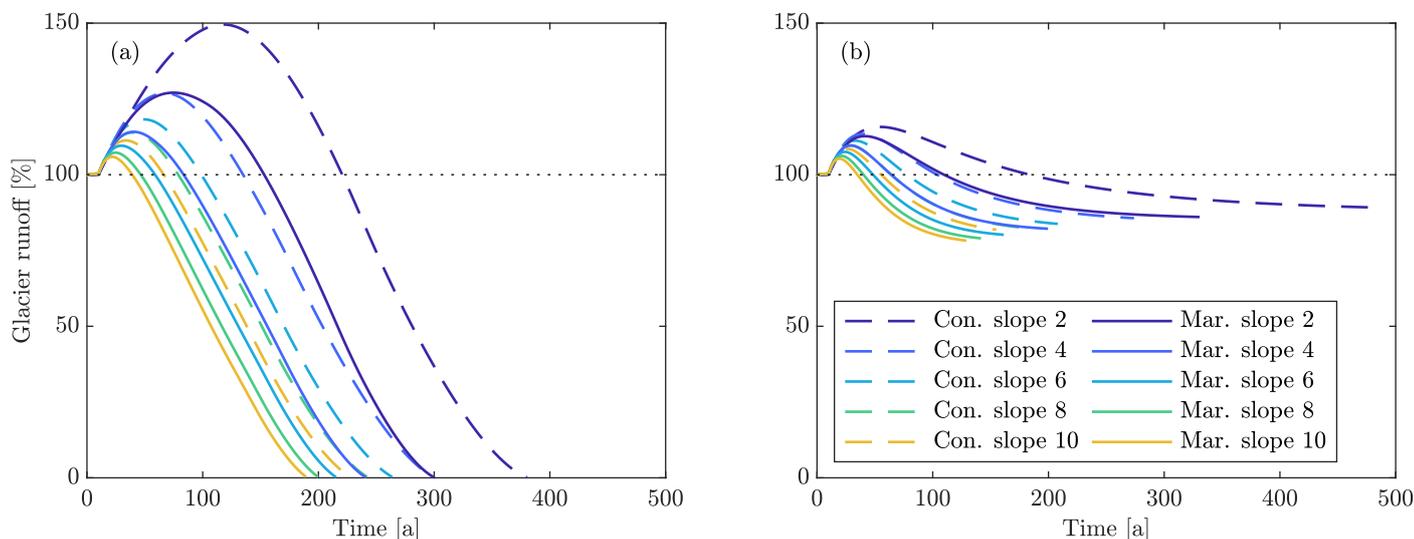
Figure 7. Variations in glacier runoff for differing slopes and climates during (a) RCP8.5 and (b) RCP2.6 climate change scenarios.

shown). The smaller contributions of nonglacier runoff in the RCP2.6 scenario reflect the smaller amount of glacier retreat that occurred during those simulations.

Basin runoff is clearly controlled by variations in glacier runoff during the early stages of retreat. Glacier runoff, Q_g (Eq. 1), undergoes a transient increase followed by a decrease to below preretreat levels (Fig. 7). This pattern is consistent for all

5 glacier geometries and climate variations. Glaciers with steep basin (bed) slopes experience lower fractional peaks in glacier runoff compared to glacier with shallow basin slopes, with variations in slope eliciting a 35% difference in peak fractional glacier runoff for the RCP8.5 scenario (Fig. 7a) and an 8% difference in peak fractional glacier runoff for the RCP2.6 scenario (Fig. 7b). The smaller peak glacier runoff of glaciers in steep basins is also associated with an earlier peak glacier runoff. For example, in a continental climate glaciers in steep basins experience peak glacier runoff roughly 80 and 30 years before shallow

10 glaciers for the RCP8.5 and RCP2.6 scenarios, respectively. Furthermore, continental glaciers have higher peak glacier runoff, reach peak glacier runoff later, and exhibit greater variations in glacier runoff between low and high slope basins than maritime glaciers. Peak glacier runoff occurs before peak basin runoff due to decreases in precipitation on glaciated area. The impacts of basin slope and climate type on peak glacier runoff timing and magnitude are similar to those observed for basin runoff.

4   Discussion

15 ### 4.1   Glacier-driven hydrological change

Our projections of the long term impacts of glacier volume loss on annual basin runoff agree closely with previous conceptual models suggesting that basin runoff will increase sharply at the onset of glacier recession, peak as glacier coverage in the basin diminishes, and then return to a steady state below the preretreat basin runoff level when the basin becomes deglacierized (Jans-

son et al., 2003; Moore et al., 2009). Moreover, our model results provide insights into how climate and basin characteristics influence variability in the timing and magnitude of the hydrologic response among glacierized basins. In particular, glacier slope exerts a strong control on the timing and magnitude of annual basin runoff, with steeper glaciers having a shorter time to peak basin runoff and a lower peak basin runoff compared to shallower sloped glaciers. Unlike peak basin runoff, final steady

5 state basin runoff following glacial recession is strongly influenced by the rate and type of vegetation that colonize ice-free landscapes within a basin.

Our model results also suggest that climate regime is an important control on basin hydrological response to glacier loss, with basins in continental climates experiencing a later and proportionally larger annual basin runoff response to glacier volume loss. This finding is in agreement with field measurements from a paired basin study in Alaska showing that glacier volume

10 change has a strong impact on annual basin water yield in a continental environment in part because of the fact that glacier volume change accounts for a larger proportion of annual streamflow in interior mountain ranges (O'Neel et al., 2014). The rate of climate warming in a basin can similarly impact the hydrological response. In particular, applying the RCP8.5 climate scenario to our model elicited a stronger and more variable response in annual basin runoff compared to the more moderate RCP2.6 climate scenario. Much of the difference in response is due to the fact that the glacier completely disappears from the

15 basin in the RCP8.5 scenario, resulting in a longer response time, a higher peak in annual basin runoff, and a substantially lower end basin runoff regardless of individual basin characteristics.

It is difficult to directly validate our model results because of a lack of discharge data for glacierized watersheds that span timescales comparable to those we modeled. However, comparisons to studies from individual glacier basins provide insight into whether the timing and magnitude of changes in annual runoff we modeled are consistent with previous runoff projections

20 from individual basins across a range of climate conditions. At the Hofsjökull ice cap in Iceland, basin runoff is projected to peak in about a century at 50% above current basin runoff levels (Aðalgeirsdóttir et al., 2006), which agrees well with our model results indicating that shallow sloping maritime glaciers reach a peak basin runoff of 43% after 110 years of climate warming under RCP8.5. In Alaska, increases in summer basin runoff in continental and maritime glacierized catchments over a 30+ year period are similar in magnitude to our model results for annual basin runoff with increases in basin runoff on the

25 order of 15–25% (O'Neel et al., 2014). Our model results for annual basin runoff are generally lower than both Aðalgeirsdóttir et al. (2006) and O'Neel et al. (2014) possibly because both of these studies account for long term increases in precipitation that are not included in our more general modeling.

In the Pacific Northwest of North America, Nolin et al. (2010) modeled the relationship between changes in glacier extent and end glacier runoff for Eliot Glacier and found that end glacier runoff was reduced by 0.9% for every 1% decrease in glacier

30 area. This finding is highly consistent with our simulations for maritime glaciers during RCP 2.6, which showed that glacier area losses of 16–24% corresponded with decreases in end glacier runoff of 14–22%. Model predictions for changes in basin runoff in Peru's Cordillera Blanca show that end basin runoff after glaciers fully disappear decreases ∼30% from present day values (Baraer et al., 2012). Present day values for nearly all basins in Baraer et al. (2012) are past peak discharge (∼30–40 years). Their model results for decreases in present day basin runoff to end basin runoff fall within the range of values that we

35 predict for decreases from peak basin runoff to end basin runoff after a basin is deglacierized, ∼20–90%.

In a global scale analysis of future basin runoff in 56 large-scale glacier basins, the increase in basin runoff to peak basin runoff averaged 26% for the RCP2.6 scenario and 36% for the RCP8.5 scenario (Huss and Hock, 2018), which is in broad agreement with our results. Moreover, the findings of Huss and Hock (2018) suggest that (i) there is a significant positive correlation between glacier area (shallow sloped glaciers in our study) and time to peak basin runoff and (ii) increases in the

5   strength of the warming scenario result in a later and higher peak basin runoff. The similarity in our findings across a wide range of basin characteristics provides confidence in the trends elucidated by our model results. The delay in peak basin runoff evident in basins with larger glaciers and basins that undergo stronger warming scenarios highlights the roles of glacier surface area and overall melt rates for determining the long term hydrological response in basins that contain glacier ice.

In many regions, glaciers have receded to the point where glacier and basin runoff have passed the peak runoff tipping points

10   and are exhibiting declines in annual water output (Stahl and Moore, 2006; Bliss et al., 2014; Huss and Hock, 2018). In these basins, the variation in glacier response will no longer be the major driver of variation in basin runoff and further examination of the ecohydrological impact of vegetation colonization is warranted. In glacierized basins with declining annual runoff, increased evapotranspiration and canopy interception resulting from vegetation will become an increasingly important driver of long-term variation in annual basin runoff. This finding suggests that decreases in annual basin water yields associated with

15   glacier loss may be especially pronounced in regions such as Patagonia, New Zealand, and coastal Alaska where productive forests can rapidly recolonize newly exposed landscapes following the loss of glacier ice (e.g. Crocker and Major, 1955; Chapin et al., 1994).

We acknowledge that our study focused on annual basin runoff and did not explore climate-driven changes in the seasonality of basin runoff, which may be substantial even in basins where annual runoff remains largely unchanged. In particular,

20   discharge data and model results from glacierized basins suggest that late summer basin runoff may decrease substantially with the continued loss of glacier ice (Huss and Hock, 2018; Kaser et al., 2010; Stahl and Moore, 2006; Nolin et al., 2010). Nevertheless understanding future changes in annual basin runoff is useful for understanding the overall hydrologic response of glacierized basins and how the wide range of ecosystem services these basins provide (e.g. Milner et al., 2017) may respond to future warming. Our findings suggest that changes in annual basin runoff with glacier loss may vary on regional scales as a

25   result of differences in climate regime (maritime vs. continental) and regional differences in the strength of the climate warming signal. However, sub-regional variation in the hydrologic response may also be considerable as a result of catchment-scale differences in aspect, elevation, slope, and latitude, all of which influence rates of glacier ice loss and subsequent colonization. The impact of basin characteristics on ice loss, regionally and subregionally, is primarily through their influence on glacier response time.

30   **4.2 Glacier response times**

Our results indicate that the initial hydrological responses to glacier recession are dominated by variations in glacier runoff, which are themselves a result of glacier dynamic feedbacks. Thus, the peak basin runoff and time to peak basin runoff can be understood in terms of the time that it takes a glacier to respond to climate change. Theoretical work (Harrison et al., 2001)
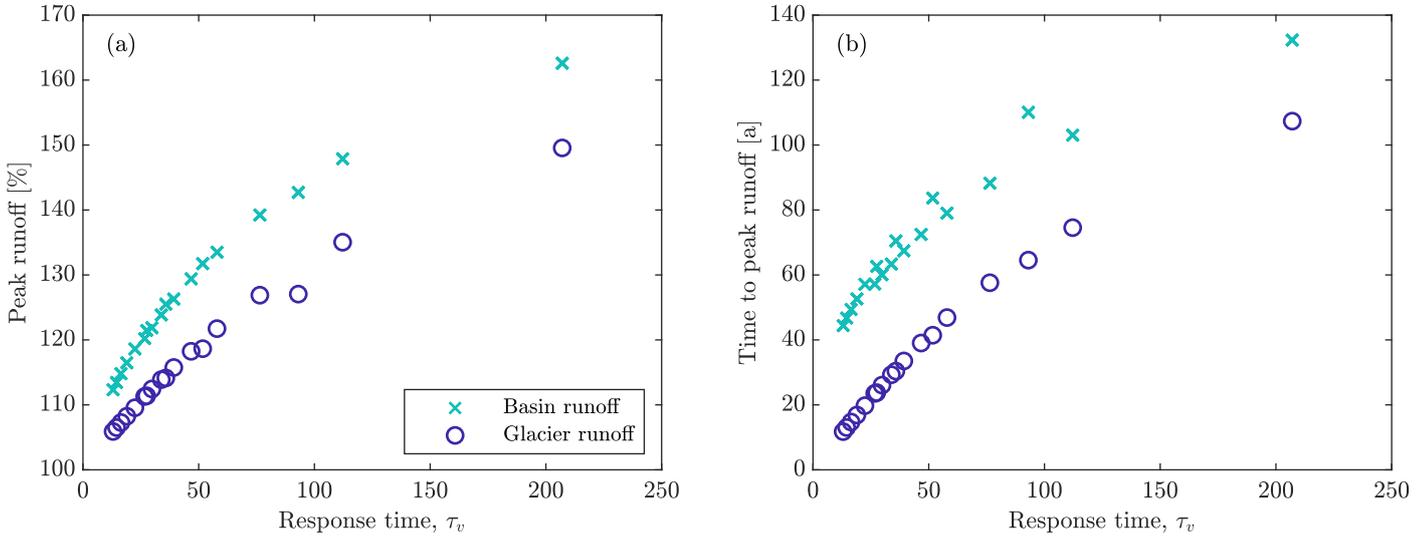
Figure 8. Relationships between glacier response time and (a) peak basin and glacier runoff and (b) time to peak basin and glacier runoff for the RCP8.5 climate change scenario. The peak basin runoff calculations assume a rocky landscape (with no vegetation).

suggests that the glacier volume response time, τ_v, is given by

$$\tau_v = \frac{1}{-\dot{b}_e/H^\star - \dot{G}_e} \, , \qquad (13)$$

where $\dot{b}_e$ is the effective specific mass balance rate in the vicinity of the terminus, $\dot{G}_e$ is the effective gradient of the specific mass balance rate with elevation, and $H^\star = \frac{\mathrm{d}V}{\mathrm{d}\Omega_g}$ is a thickness scale in which V is volume and recalling that Ω_G is glacier sur-

5 face area. The volume response time is the e-folding time for the volume of a glacier to evolve from one steady-state to another following a step change in climate. Equation 13 is derived from mass continuity arguments and characterizes both the timing and magnitude of the volume response of a glacier (longer response times result in larger changes in volume; Harrison et al., 2001), but depends on the assumption that $H^\star$ is constant and therefore that changes in volume/climate are small. Nonetheless, the glacier response time is a useful tool for understanding how glacier volume and glacier runoff might be expected to evolve

10 in a changing climate. In particular, Equation 13 indicates that glacier response times will be largest for thick glaciers (i.e., those that occur on shallow slopes) in a continental climate. The first term in the denominator, $-\dot{b}_e/H^\star$, will always be positive, and the larger this value the shorter the response time. Glaciers in continental climates typically have relatively small values of $-\dot{b}_e$ and thus this term tends to be small (indicating long response times). The second term in the denominator, $\dot{G}_e$, is positive and acts to increase the response time by accounting for the impact of climate on mass redistribution from high elevations to

15 low elevations. Mass balance gradients are smaller in continental climates than in maritime climates. However, the $-\dot{b}_e/H^\star$ term is significantly smaller in continental climates, and thus the denominator is about half as small (i.e., the glacier response time is twice as long) for continental glaciers than it is for similarly sloped maritime glaciers.

Our modeling results do not follow the assumption of small changes in climate, but nonetheless are broadly consistent with the notion of glacier response time. We calculate the glacier response time using the initial balance rate at the terminus, the balance gradient for the respective climate, and the thickness scale $H^\star$ by calculating an average value of $dV/d\Omega_g$ during the first quarter of each simulation. We find that peak glacier runoff is highest and occurs latest for shallow sloped glaciers in a

5 continental climate (i.e. those that have long response times; Fig. 8). Similarly, because variations in basin runoff are strongly influenced by glacier runoff in the early stages of retreat, glacier response time is also a useful predictor of peak basin runoff and time to peak basin runoff. We find nearly linear relationships between response time and both peak runoff and time to peak runoff for small response times (for both glacier runoff and basin runoff). The relationships deviate from linear because the response time calculation (i) does not directly indicate when the rate of volume loss is at a maximum, (ii) does not account

10 for changes in basin/glacier runoff due to precipitation (it is only a statement about glacier evolution), and (iii) is based on the assumption that changes in volume are small and therefore that $H^\star$ is a constant, which breaks down as the response time increases. In this context, a key result is that peak basin runoff and time to peak basin runoff are largest for basins containing glaciers that have long response times.

5   Conclusions

15 Basin runoff varies during glacier recession due to release of water from glacier storage and subsequent colonization of deglaciated land. Rapid glacier mass loss during the early stages of retreat drives an increase in basin runoff, which eventually decreases as a glacier shrinks and the landscape becomes increasingly vegetated. Peak basin runoff and time to peak basin runoff are largely driven by glacier response to climate change due to the major contribution of glacier runoff to basin runoff during the initial stages of retreat. Basins with glaciers that have fast response times (i.e., steep and maritime) have

20 lower and earlier peak basin runoff because those glaciers respond rapidly to climate warming. Slow responding glaciers (i.e., shallow sloping and continental) are unable to stay in step with climate variations and consequently experience high sustained rates of volume loss well after the initiation of climate warming, resulting in high and late peak basin runoff. In the later stages of retreat, nonglacier runoff becomes an increasingly significant contributor to basin runoff. The time at which basin runoff falls below preretreat levels is heavily influenced by the colonization of the basin. Basins with fast and high levels of vegetation

25 have earlier and lower peak runoff and reach preretreat levels of basin runoff substantially earlier than those with low levels of vegetation. Basin runoff in the late stages of glacier recession is primarily determined by the vegetation in the basin because the glacier runoff becomes small or negligible (depending on the degree of climate warming that has occurred).

The model simulations that we performed in order to explore variations in basin runoff were highly idealized and aimed at elucidating the fundamental controls on basin runoff over annual time scales and longer. In particular, we assumed constant

30 glacier width, uniform basin slope, and simplified parameterizations of climate, climate change, and vegetation succession. We also note that we began all simulations with 100% glacier cover and a glacier in a near steady-state, and consequently our simulations tend to overemphasize the impacts of glacier recession on basin runoff. Future work should explore in more detail the effect of basin hypsometry on basin runoff and also incorporate more sophisticated climate and hydrological models.

Increased model complexity would be required to address the impact of climate change on the timing of runoff on seasonal and shorter time scales, which is critical for water resource management since glacier runoff tends to be highly focused and highly variable during the summer months.

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