

Present-day and future contributions of glacier runoff to summertime flows in a Pacific Northwest watershed: Implications for water resources

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[1] While the impacts of long-term climate change trends on glacier hydrology have received much attention, little has been done to quantify direct glacier runoff contributions to streamflow. This paper presents an approach for determining glacier runoff contributions to streamflow and estimating the effects of increased temperature and decreased glacier area on future runoff. We focus on late summer streamflow (when flow is lowest and nonglacier contributions to flow are minimal) of a small glacierized watershed on the flanks of Mount Hood, Oregon, United States. Field and lab measurements and satellite imagery were used in conjunction with a temperature-index model of glacier runoff to simulate potential effects of increased temperature and reduction in glacier area on late summer runoff in the watershed. Discharge and stable isotope data show that 41–73% of late summer streamflow is presently derived directly from glacier melt. Model simulations indicate that while increased temperature leads to rapid glacier melt and therefore increased streamflow, the consequences of glacier recession overcomes this effect, ultimately leading to substantial declines in streamflow. Model sensitivity analyses show that simulation results are most sensitive to degree day factor and less sensitive to uncertainties in debris-covered area and accumulation area ratio. This case study demonstrates that the effects of glacier recession on streamflow are a concern for water resource management at the local scale. This approach could also be extended to larger scales such as the upper Columbia River basin where glacier contributions to late summer flows are also thought to be substantial.

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1. Introduction

1.1. Significance and Motivation

[2] Glacier runoff contributions to streamflow provide critical water supply in many mountainous regions [e.g., Singh and Singh, 2001; Barnett *et al.*, 2005]. Historical records and future climate projections point to the loss of midlatitude glaciers throughout the world [Oerlemans, 2005; Lemke *et al.*, 2007], resulting in significant changes to both total annual and summer streamflow downstream [Chen and Ohmura, 1990; Barnett *et al.*, 2005; Hock *et al.*, 2005; Juen *et al.*, 2007]. Glacier runoff supplies fresh water to numerous communities throughout the world and is highly sensitive to changes in temperature [Chen and Ohmura, 1990]. Warmer temperatures cause increased glacial melt but as glaciers recede, their potential contributions to water supplies are diminished [Barnett *et al.*, 2005; Hock *et al.*, 2005]. Glaciers also modulate intra and interannual flow variability by storing water in the form of ice during years of high

precipitation and releasing meltwater during seasons and years of high temperature [Fountain and Tangborn, 1985].

[3] The hydrologic properties of glacierized watersheds differ from glacier-free watersheds in several ways. Glaciers release an estimated two to ten times more water than neighboring catchments of equal area and altitudes in the United States [Mayo, 1984]. Runoff variability in glacierized watersheds is controlled primarily by surface energy fluxes whereas runoff variability in glacier-free watersheds is dominated by precipitation patterns [Jansson *et al.*, 2003]. There is a lag effect caused by glacial storage and the delayed networking of englacial and subglacial conduits [Jansson *et al.*, 2003] such that runoff from glacier melt is delayed until later in the summer, when other contributions to streamflow are much reduced. Glacier melt decreases streamflow variation, bolsters late season runoff, and is especially important in drought years [Fountain and Tangborn, 1985]. Under negative mass balance conditions, glaciers discharge a greater volume of water than is input in the form of precipitation and this “excess discharge” can be substantial, even for watersheds having less than 15% glacier coverage [Lambrecht and Mayer, 2009].

[4] In the northwestern United States, glaciers diminished throughout the 20th century and model simulations suggest this trend will continue through the next 100 years [Dyrgerov and Meier, 2000; Hall and Fagre, 2003]. However, there has not been any research performed on glacier

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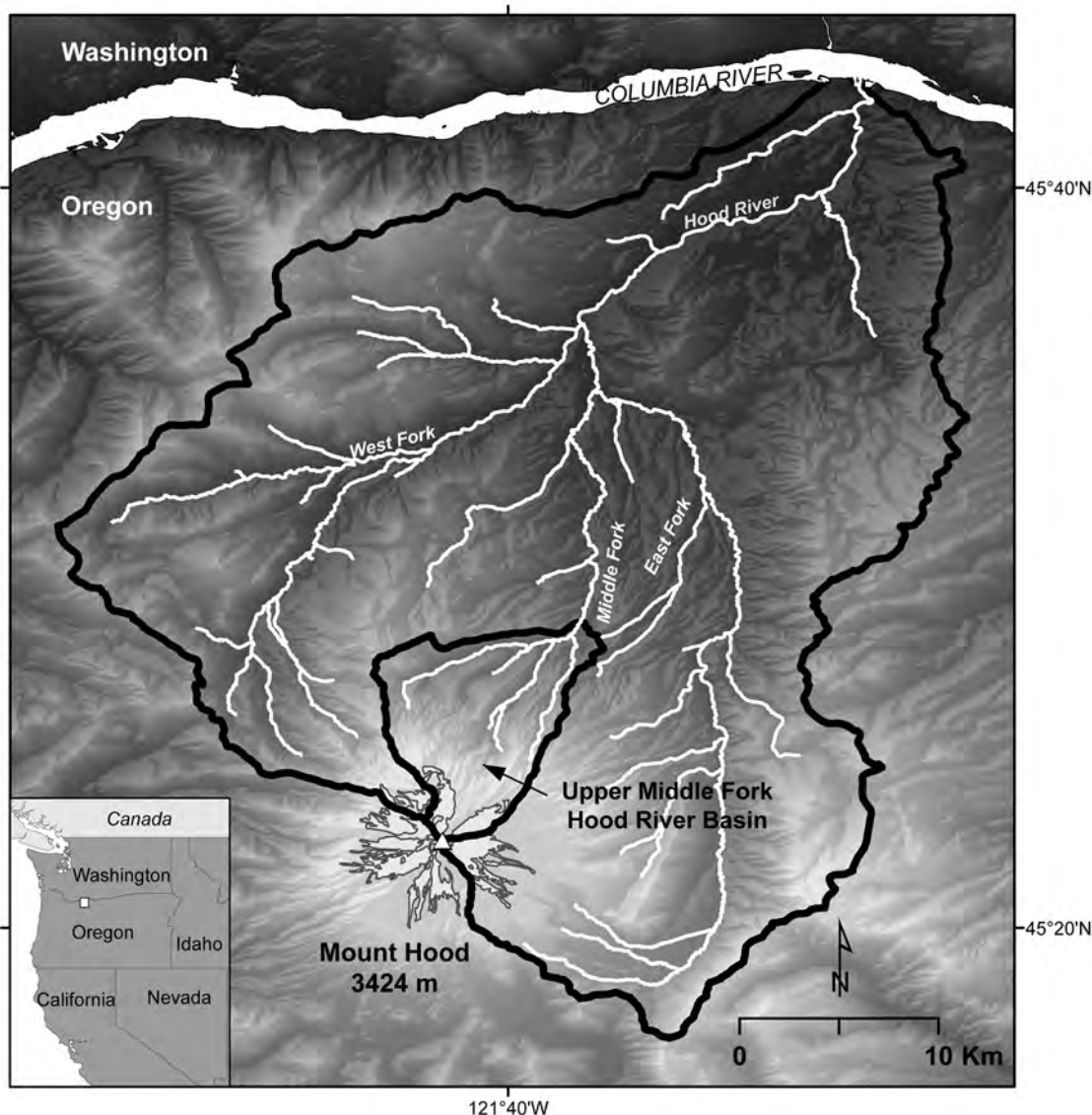


Figure 1. Hood River basin, Oregon, United States, showing Mount Hood glacier cover (gray) and Upper Middle Fork study area.

runoff contributions to streamflow in the Pacific Northwest, United States. Our research focuses on Mount Hood, the glacier-capped composite volcano that is Oregon's highest peak and the source of the Hood River, which feeds the acclaimed agricultural industry in the Hood River valley. Five irrigation districts along the East, Middle, and West Forks of the Hood River rely on late summer snow and ice melt from six glaciers, along with reservoir storage of winter rains, to meet high water demands during the dry summer months, while maintaining sufficient in-stream flows and cool water temperatures for threatened anadromous fish. Recent studies document that Mount Hood's glaciers have decreased as much as 61% over the past century [Lillquist and Walker, 2006]. However, there are no historical or current monitoring programs that provide measurements of the contributions of Mount Hood glaciers to streamflow, and therefore, no way to estimate the potential impact of their loss.

[5] The objectives of this investigation were to quantify glacier runoff contributions in an ungaged basin, to model projected impacts of warming temperatures and decreased

glacier area, and to provide a methodological template for potential future investigations of glacier runoff contributions to streamflow in similar basins.

[6] In section 2 we describe a combination of in situ and isotopic methods for directly determining glacier runoff contributions to late summer streamflow and a modeling approach for estimating future glacier runoff based on projections of increased temperature and glacier recession. Section 3 details the results of the measurements and modeling and in section 4 we address sources of uncertainty. In section 5 we discuss the implications of this work to small, glacierized watersheds as well as the potential application to larger-scale watersheds such as the upper Columbia River basin.

1.2. Description of the Study Area

[7] Located on the north side of Mount Hood, Oregon, the Hood River basin covers an area of 882 km² and drains to the Columbia River (Figure 1). The basin ranges in elevation from 26 to 3424 m, and glaciers are found above 1900 m. The

Table 1. Upper Middle Fork Hood River Watershed and Stream Characteristics

	Watershed Area (km ²)	Stream Length (km)	Glacier Area (%)	Elevation Range (m)
Eliot Creek	9.6	8.3	18.9	821–3424
Coe Creek	17.5	8.0	8.7	833–3271
Clear Creek	14.9	7.2	0	892–2132
Pinnacle Creek	7.0	5.4	0	892–1853
Upper Middle Fork Hood River	50.6	28.9	6.6	606–3424

region has a Mediterranean climate with cool, wet winters and warm, dry summers. The majority of precipitation falls from November to March (200–250 mm month⁻¹) with very little precipitation occurring from June to September (40–80 mm month⁻¹). In winter, high streamflow is produced by rainfall at lower elevations, and discharge remains high through the spring as the snowmelts progressively upward in elevation on Mount Hood. By July, seasonal snowpacks have typically melted away, causing declines in streamflow. The Hood River basin contains approximately 650 km of perennial streams, of which 150 km are spawning grounds for anadromous fish [Coccoli, 2004]. Water from the Hood River irrigates more than 53 km² of commercial pear, apple, and peach orchards. The basin as a whole has a very low percentage of glacierized area (<0.1%); however we focus on an upper portion of the basin (6.6% glacierized area), where water for irrigation is withdrawn seasonally from the river and its tributaries.

[8] Of the three main branches, the East, Middle and West Forks, of the Hood River, this research focuses on the Upper Middle Fork (50.6 km²), which is heavily utilized for irrigation in the region, and highly relevant for water management purposes (Figure 1). The headwaters of the Upper Middle Fork are on the northern flanks of Mount Hood, where glaciers cover about 3.4 km², and are drained by four streams: Eliot, Coe, Clear and Pinnacle creeks (Figure 1 and Table 1). Eliot and Coe are glacier fed, whereas Clear and Pinnacle are sourced from permanent snowfields and groundwater inputs during the summer dry season. Extraction of water from these mountain streams is especially important in the late summer as the harvest period for apples and pears approaches (D. Compton and C. DeHart, Middle Fork Irrigation District, personal communication, 2007).

[9] Eliot Glacier terminates in a relatively narrow channel, which contains all of the glacier runoff and forms the headwaters of Eliot Creek. In contrast, the Coe drainage consists of one main ice body (Coe Glacier) and additional smaller ice bodies and snowfields in the Compass Creek subbasin that melt and drain into multiple channels, ultimately entering Coe Creek about 4200 m downstream from the terminus of Coe Glacier (Figure 2). Quantification of glacier runoff contributions to the Upper Middle Fork Hood River was accomplished using a combination of discharge measurements, isotope sampling and application of a two-component mixing model.

2. Methods

2.1. Stream Discharge and Stable Isotope Measurements

[10] Stream discharge measurements are a direct means of determining contributions from glacier melt to streamflow. Discharge at the outlets of Eliot and Coe Glaciers was

measured from 10 August to 7 September 2007. To determine the contribution of this glacier runoff to streamflow at the subbasin outlet (where water is diverted for irrigation and hydropower purposes), discharge was also measured immediately upstream of the four diversions for the Upper Middle Fork Hood River on Eliot, Coe, Pinnacle, and Clear creeks (Figure 1). At each of the six sites, water height was recorded at 15 min intervals using OdysseyTM capacitance water height recorders, and discharge was measured 6–14 times during the study period using a Marsh-McBirney velocity meter following the procedure of Carter and Davidian [1968]. Stage-discharge rating curves were developed to calculate discharge from the continuous water height data. Discharge measurements from the glacier outlets were compared to measurements at the diversion sites in order to calculate the contribution of glacier melt to total streamflow.

[11] Stable oxygen isotopes in glacier melt are typically depleted relative to seasonal snow cover and summer precipitation. Isotopic analysis has been successfully used to investigate glacier melt contributions to streamflow where logistical difficulties associated with maintaining gage sites have been prohibitive [e.g., Theakstone and Knudsen, 1996; Rietti-Shati et al., 2000; Gooseff et al., 2006]. Discharge measurements, hydrochemical samples, and a two-component mixing model of oxygen isotopes were used to calculate a glacier runoff contribution of 30–45% of the total annual discharge for watersheds in the Cordillera Blanca, Peru [Mark and Seltzer, 2003] and to show that glacier melt contributions increased over a 3 year period as the glaciers receded [Mark and McKenzie, 2007].

[12] We use synoptic sampling of water oxygen isotopes in a two-component mixing analysis to compare with discharge-based calculations of glacier contributions to streamflow. Water samples were collected at seven locations during three sampling trips between August and October 2007 (Table 2). These samples were collected at each of the four discharge measurement sites and at three springs that source water for tributaries of Coe and Eliot creeks between the glaciers and the diversions. The springs were sampled as a representation of groundwater contributions to streamflow. Springs in the Oregon Cascades have been shown to have isotope ratios that are constant throughout the year [Jefferson et al., 2006], so the three samples collected on two dates should adequately characterize the isotopic composition of the groundwater. Clear and Pinnacle creeks are not glacier fed so they were not part of the isotopic study. Small ice bodies in the Coe Creek watershed were not individually sampled, but assigned the same isotopic composition as Coe Glacier. It was not necessary to sample snowmelt, as all water samples were collected after the snowmelt season.

[13] Samples were analyzed for the ratio of ¹⁸O to ¹⁶O at the Isotope Ratio Mass Spectrometer Facility at Oregon State University (Corvallis, Oregon) using a ThermoFinniganTM Delta Plus XL (dual inlet). Values are reported in the standard notation ($\delta^{18}\text{O}$) as permil (‰) deviations relative to Vienna Standard Mean Ocean Water (VSMOW), with a precision of $\pm 0.03\text{‰}$. Standard equations for a two-component isotopic mixing model [Sklash and Farvolden, 1979] were developed with glacier runoff and groundwater as the end-member components

$$Q_{\text{glacermelt}} = \left(\frac{\delta^{18}\text{O}_{\text{stream}} - \delta^{18}\text{O}_{\text{groundwater}}}{\delta^{18}\text{O}_{\text{glacermelt}} - \delta^{18}\text{O}_{\text{groundwater}}} \right) * Q_{\text{stream}}, \quad (1)$$

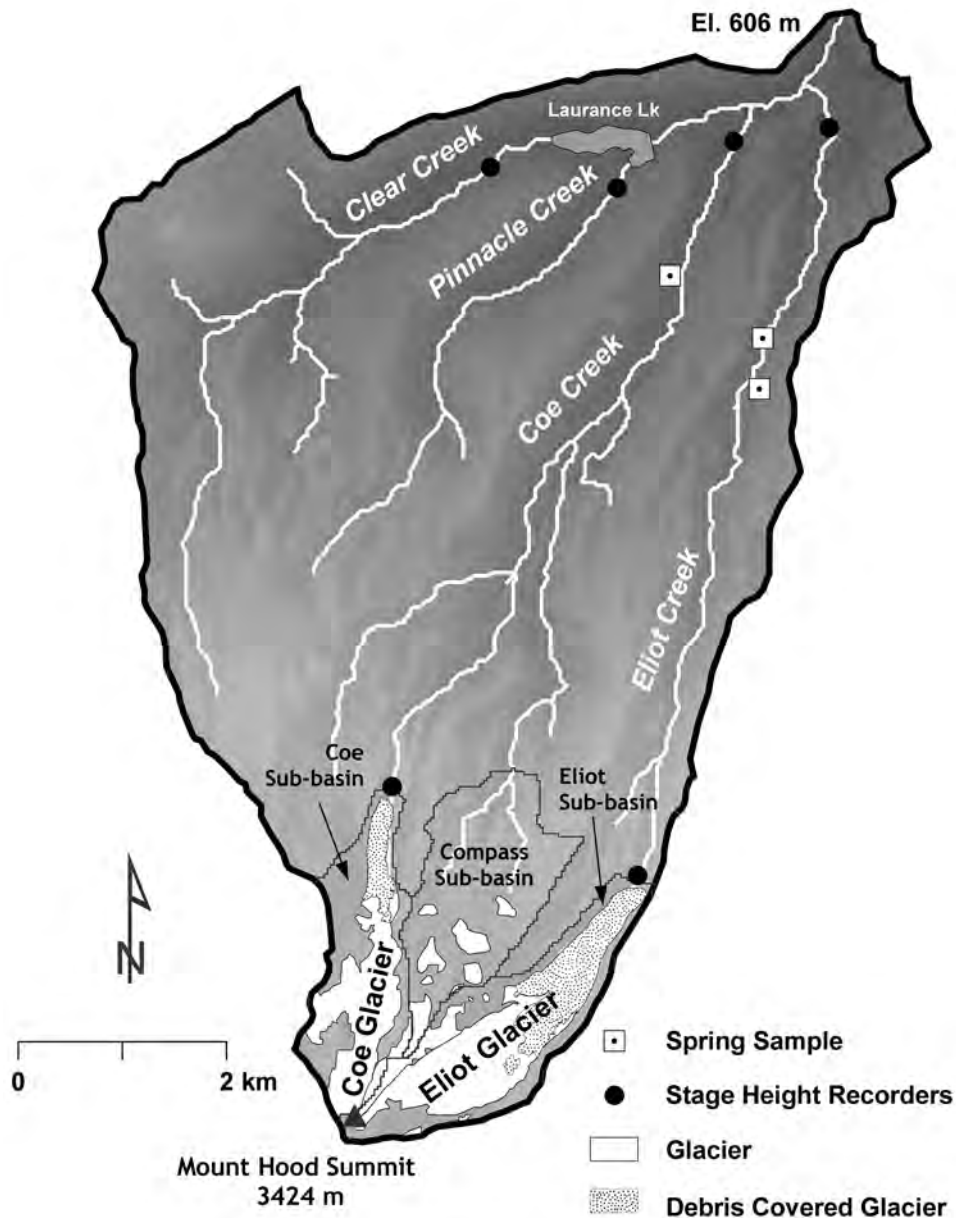


Figure 2. Digital elevation model of the Upper Middle Fork Hood River watershed indicating study sub-basins and creeks, glacier locations (white), debris-covered glacier extent (stippled), water height recorders (circles), and spring water sample locations (squares with center dots).

where Q is discharge and the subscripts *stream*, *groundwater*, and *glaciermelt* represent water from the streams, groundwater, and glacier melt, respectively.

2.2. Glacier Melt Modeling

[14] Projected future glacier runoff contributions to streamflow are qualitatively estimated using a temperature-index model calibrated with present-day discharge. We assess the model's sensitivity to various input parameters to bound our estimates. Projected changes in glacier area and temperature are then used to investigate glacier melt contributions. This study uses the Snowmelt Runoff Model (SRM) [Martinec, 1975; Rango and Martinec, 1995] instead

of a more complex energy balance model because the study area lacks sufficient physical data to calculate energy fluxes. Furthermore, SRM has obtained excellent results in high altitude terrain [Ferguson, 1999] and has recently been validated for use in glacier melt computations [Schaper et al., 2000]. There is also a physical justification for using air temperature as an index for calculating melt, because the primary heat sources for melt, radiation and sensible heat flux, are highly correlated with temperature [Ohmura, 2001; Kuhn, 1993]. Wind speed, the energy balance input least correlated with temperature, is a very small contributor to melt [Ohmura, 2001].

Table 2. The $\delta^{18}\text{O}$ Analysis and Resulting Proportions of Glacier Runoff

Source	Sample Collection Date and Local Time	$\delta^{18}\text{O}$ (‰)	Glacier Runoff Contribution (%)
Spring 1, Coe	10 Sep 2007 1410	-11.63	-
Spring 2, Eliot	13 Sep 2007 1406	-11.53	-
Spring 3, Eliot	13 Sep 2007 1440	-11.63	-
Eliot Glacier	24 Aug 2007 1606	-13.22	-
	14 Sep 2007 0847	-13.74	-
	13 Oct 2007 1900	-14.23	-
Lower Eliot Creek	24 Aug 2007 1712	-13.02	88 ± 4
	14 Sep 2007 0727	-13.26	78 ± 3
	13 Oct 2007 2157	-13.59	76 ± 3
Coe Glacier	24 Aug 2007 1330	-12.82	-
	11 Sep 2007 1640	-12.74	-
Lower Coe Creek	24 Aug 2007 1330	-12.67	88 ± 5
	11 Sep 2007 2210	-12.40	70 ± 6

[15] SRM uses a degree day approach to calculate total ice and snowmelt [Kustas *et al.*, 1994] and computes daily runoff using the following equation [after Martinec *et al.*, 2007]:

$$Q_{n+1} = [c_{Sn} \cdot a_n(T_n + \Delta T_n)G_n + c_{Rn}P_n] \frac{A \cdot 10000}{86400} \cdot (1 - k_{n+1}) + Q_n k_{n+1}, \quad (2)$$

where

- Q average daily discharge (m^3s^{-1});
- n model day number;
- c runoff coefficient expressing losses as the ratio of runoff to precipitation (c_S refers to runoff derived from snowmelt and c_R refers to runoff derived from rainfall);
- a degree day factor ($\text{cm} \text{ } ^\circ\text{C}^{-1} \text{ d}^{-1}$) indicating the water equivalent snowmelt depth resulting from 1 degree day;
- T number of degree days ($^\circ\text{C d}$);
- ΔT the adjustment by temperature lapse rate when extrapolating the temperature from the station to the hypsometric average of the model elevation zone ($^\circ\text{C d}$);
- G ratio of the glacier covered area to the total area of the basin or elevation zone (modified from snow covered area from Martinec *et al.* [2007]);
- P precipitation contributing to runoff (cm);
- A area of the basin or elevation zone (km^2);
- k recession coefficient showing the decrease in discharge during a time period during which there is no snowmelt or rainfall.

[16] The model was run for the period 1 August 2007 to 30 September 2007. Meteorological data from three stations (Table 3) were used as daily input to SRM. Mount Hood Meadows daily temperature data were used to calibrate the model for the study period. A 25 year record of daily temperature and a 27 year record of daily precipitation from the Mount Hood SNOTEL site were used to generate daily climatological mean values of temperature and precipitation, which were used for the SRM simulations. Precipitation from the Red Hill and Mount Hood SNOTEL sites were used to

compute a precipitation lapse rate. Precipitation during the study period was minimal, with two short events producing <4 cm of rain in total. The model calibration period was typical of late summer climate for the study area with maximum and minimum temperatures within 0.45 standard deviations of the mean (25 year record; Mount Hood Meadows; Table 3) and precipitation within 0.06 standard deviations of the mean (27 year record; Mount Hood SNOTEL; Table 3). Thus, the model is calibrated on conditions that are very similar to those of the climatological mean temperature and precipitation.

[17] In this study, glacier-covered area (G in equation (2)) takes the place of the standard snow covered area in SRM. A 30 m resolution image from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) satellite, acquired on 10 September 2006, was used to delineate the debris-free portions of Eliot and Coe Glaciers as well as the smaller ice bodies and permanent snowfields of the Compass Creek subbasin, which drains into Coe Creek (Figure 2). We used the method of Taschner and Ranzi [2002] to map glacier-covered area by computing a ratio of ASTER bands 3 (0.76–0.86 μm) and 4 (1.60–1.70 μm). This method is effective for mapping the upper debris-free portions of Eliot and Coe Glaciers and the small ice bodies and snowfields but does not suffice for the debris-covered lower sections of the Eliot and Coe Glaciers. Instead, the debris-covered portions of these two glaciers were delineated using a combination of Global Positioning System (GPS) readings taken on 14 September 2007 using a Garmin e-Trex Legend handheld unit and analysis of 1 m resolution orthorectified aerial photographs acquired in June 2005. If glacier area was computed based solely on the ASTER band ratio method, the area of the Eliot Glacier would have been underestimated by 40%. It is interesting to note that there was substantial error in the United States Geological Survey (USGS) Quads 1956 depiction of Eliot and Coe Glaciers. These maps, derived from aerial photos acquired at an unknown time of summer did not include more than 60% of the debris-covered glacier areas of Eliot and Coe Glaciers. They also overestimated glacier coverage in other areas, likely due to inclusion of seasonal snow and snowfields as well as changes in glacier extent since the acquisition of the baseline aerial photos. Our mapping of the full extent of glaciers in the Upper Middle Fork Hood River basin includes the debris-covered lower portions of the Eliot and Coe Glaciers (Figure 2). Glacier-covered area was assumed to be constant through the 2 month modeling period.

[18] A standard temperature lapse rate of $0.65^\circ\text{C}/100 \text{ m}$ was used to spatially distribute the temperature data over the full elevation range [Barry, 1992]. A precipitation lapse rate of $6.4\%/100 \text{ m}$ was computed using the differences in the 10 year mean August and September precipitation between the Red Hill SNOTEL site (1341 m) and the Mount Hood SNOTEL site (1646 m). This lapse rate was applied to the hypsometric mean elevation of each 200 m elevation zone

Table 3. Meteorological Stations

Station Name	Altitude (m)	Geographic Coordinates	Period of Data Extraction
Mount Hood SNOTEL	1637	45.32°N, 121.71°W	1981–2007 (Aug–Sep)
Mount Hood Meadows (Mesowest ID MHM52)	1600	45.33°N, 121.60°W	2007 (Aug–Sep)
Red Hill SNOTEL	1341	45.47°N, 121.70°W	1998–2007 (Aug–Sep)

in the model. As noted above, there was only a small amount of precipitation during the study period (3.8 cm as measured at the Mt. Hood SNOTEL site, <http://www.or.nrcs.usda.gov/snow/maps/sitepages/21d08s.html>, last accessed 23 June 2010).

[19] The degree day factor (DDF, a in equation (2)) is typically (1) measured using ablation stakes or a snowmelt lysimeter, (2) calculated from an energy-balance equation [Zhang *et al.*, 2006a], or (3) computed according to its relationship to snow density [Martinec, 1960]. In the absence of such data, we used a mean of DDF values from 21 previously published studies measuring DDF of snow and ice on glaciers [see Singh *et al.*, 2000; Zhang *et al.*, 2006a]. The mean DDF for snow ($4.4 \text{ mm } ^\circ\text{C}^{-1} \text{ d}^{-1}$) was applied to all zones above our estimated equilibrium line altitude (ELA). The ELA represents the elevation on the glacier where accumulation is roughly equal to ablation and is often represented by the snow-ice boundary on the glacier at the end of summer. The mean DDF for ice ($7.1 \text{ mm } ^\circ\text{C}^{-1} \text{ d}^{-1}$) was applied to the ablation zone above the debris-covered section of the glacier. This method of employing the estimated ELA as a boundary for the DDF was successfully used in a temperature-index model by Zhang *et al.* [2006b]. Here, we use the end of summer snowline as an estimate of ELA. Analysis of in situ photographs and high-resolution orthorectified aerial photographs from August 2005 generated an ELA for Eliot Glacier at an elevation of approximately 2300 m. In a similar manner, the ELA of Coe Glacier was determined to be 2230 m, and this value was extrapolated to neighboring small ice bodies for use in SRM.

[20] The DDF for debris-covered glaciers can be difficult to determine [Hochstein *et al.*, 1995]. The thickness of debris cover has been shown to be inversely related to the glacier ablation rate [Kayastha *et al.*, 2000]. Since DDF is equal to the ablation rate divided by the number of degree days, we calculated a spatially distributed DDF for the debris-covered sections of Eliot Glacier using ablation rate measurements from Jackson [2007] and Lundstrom [1992]. These empirically derived DDF values for the debris-covered glacier range from 0.06 to $0.38 \text{ mm } ^\circ\text{C}^{-1} \text{ d}^{-1}$. Debris thickness data were not available for Coe Glacier so, based on data from Eliot Glacier we assumed a mean debris thickness (36 cm) over the debris-covered portion of Coe Glacier and applied a single DDF to that portion of the glacier. A spatially weighted mean DDF was calculated for SRM elevation zones that contained more than one DDF value. The sensitivity of the model to uncertainties in the values of DDF, ELA and debris cover extent was assessed with additional models runs.

3. Results

3.1. Present-Day Glacier Contributions to Streamflow

3.1.1. Discharge Measurements

[21] Discharge measurements in Eliot and Coe creeks over the period from 10 August to 7 September 2007 exhibited a strong diurnal signal (Figure 3), which is likely due to air temperature patterns affecting glacier melt rates. With no glacier runoff inputs, Clear Creek showed minimal variations in the daily cycle of discharge (Figure 3). A lag in peak daily discharge between the glacier terminus sites and their respective lower gaging sites in Eliot and Coe creeks is observed. Computing the root mean squared (RMS) difference between discharge at the Eliot Glacier gage and the

lower Eliot Creek gage (7.7 km downstream) produced an estimated overall lag of 1 h 50 min, ranging from 1 h 30 min to 2 h 40 min over the sampling period. The computed time lag between the Coe Glacier terminus and the corresponding downstream gage (7.1 km) was also calculated to be 1 h 50 min with variations from 1 h 4 min to 2 h 30 min. Groundwater contributions and flow paths with in-stream and hyporheic transient storage probably account for the slower recession in the downstream gages and the imperfect alignment of hydrographs when lagged by 1 h 50 min.

[22] On a volumetric basis, discharge from Eliot Glacier contributed 87% of the discharge measured at the downstream Eliot Creek site over the total 28 day sampling period. Coe Glacier contributed 31% of the measured volumetric flow to the downstream Coe Creek site. As described earlier, the headwaters of Coe Creek are derived from the main ice body of Coe Glacier and several smaller ice bodies and snowfields in the Compass Creek subbasin. Thus, this value serves as a lower bound for the glacier runoff contribution to Coe Creek. Another means of estimating the unmeasured glacier melt contributions from Compass Creek is to scale the measured discharge from Coe by the relative area of the Compass Creek ice bodies. Combining the estimated discharge from the Compass ice bodies with the measured discharge from Coe Glacier gives an estimated an overall glacier contribution of 38% to Coe Creek.

[23] When we consider all surface runoff, including that from the nonglacierized Pinnacle and Clear creeks, we found that Eliot and Coe creeks comprised 26% and 59%, respectively, of the volumetric discharge for the Upper Middle Fork Hood River over the sampling period. While their total combined contribution to discharge for the Upper Middle Fork Hood River is 85%, their combined areas are only 55% of the Upper Middle Fork Hood River watershed. Nonglacierized Pinnacle and Clear creeks, which together drain 45% of the total watershed area, contributed just 11% and 4%, respectively, to discharge in the Upper Middle Fork Hood River during this study.

3.1.2. Oxygen Isotope Measurements

[24] Isotopic data from the three sampling periods in August–October 2007 suggest that glacier runoff is significantly more depleted in heavy isotopes than the springs. The three spring samples were very similar, with an average of -11.6‰ and a total range of only 0.1‰ (Table 2), while glacier runoff ranged averaged -12.6‰ for Coe and -13.7‰ for Eliot. Values from the downstream samples are intermediate between the glacier runoff and the spring samples. The 1.1‰ difference in isotopic composition of glacial runoff between Eliot and Coe Glaciers is likely due to isotopic interactions between ice and meltwater as melt travels through englacial conduits, and is much smaller than the range of isotopic changes that typically occur in snowmelt [Taylor *et al.*, 2002].

[25] The isotopic composition of precipitation events varies causing snow layers to have varying isotopic compositions. Postdepositional processes such as sublimation will alter the isotopic composition of snow, firn, and glacier ice. Mixing of meltwater from different layers will lead to variations in isotopic composition. The pathways will vary from glacier to glacier thus meltwater exiting one glacier can have a different isotopic composition than meltwater from a nearby glacier.

[26] Applying the two-component mixing model (equation (1)) to the downstream samples indicates that

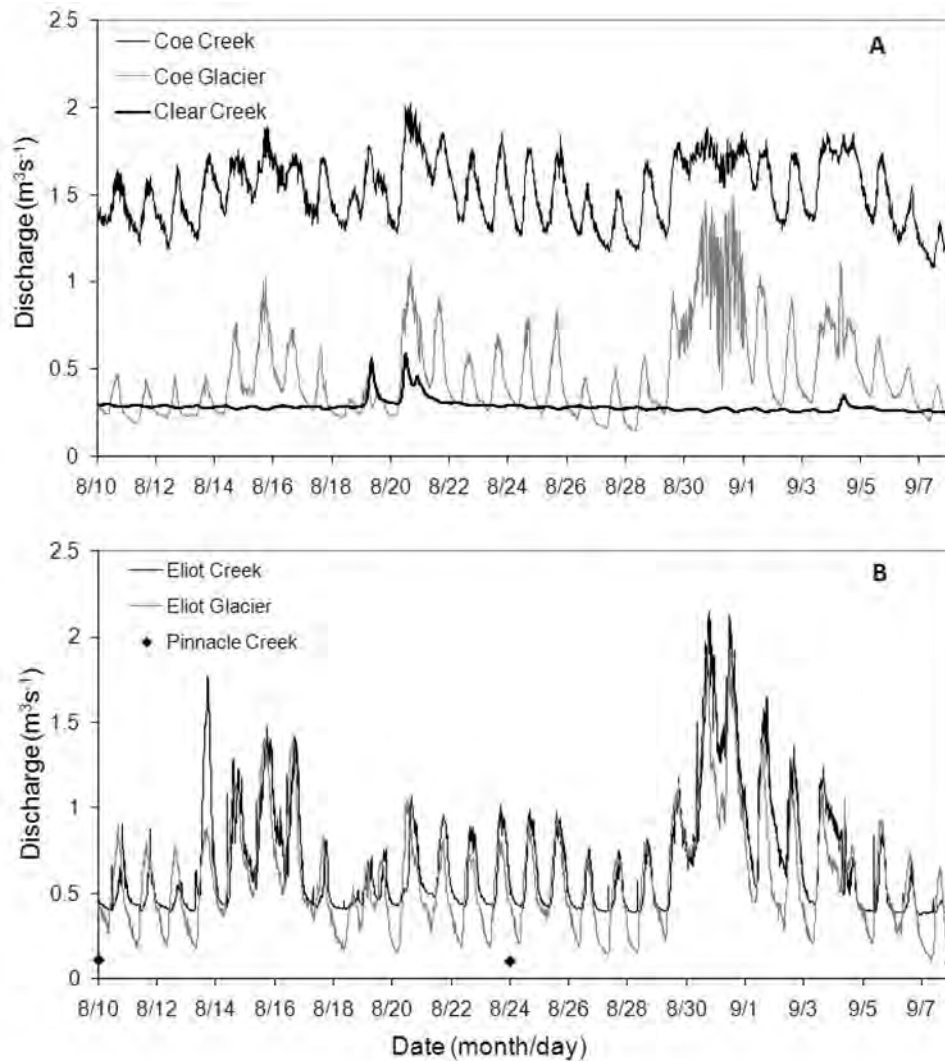


Figure 3. Discharge measurements for study period in 2007 for sites on (a) Coe and Clear creeks and (b) Eliot and Pinnacle creeks. Lines represent calculated discharge from stage–discharge relationships, black diamonds represent direct measurements of discharge at Pinnacle Creek.

glacier melt contributes between 76% and 88% of the discharge at the downstream Eliot Creek gage during the sampling periods (Table 2). For Coe Creek, the isotope tracer-derived estimates of glacier runoff contribution were similar to those of Eliot Creek with isotopically derived values ranging from 70% to 88%.

[27] Comparing the discharge-derived estimates of glacier runoff contributions with those estimated using the isotopic method, we find that they agree closely for Eliot Creek but not for Coe Creek. For the 24 h period centered on the time of the sample collection on 24 August 2009, the volumetric contribution of discharge from Eliot Glacier runoff to discharge at the downstream gage was 87% compared with the isotopically derived estimate of $87 \pm 4\%$. A similar comparison for 24 August 2009 of Coe Creek gives a volumetric contribution of glacier runoff of only 31% compared with the isotopically derived glacier runoff contribution of $88 \pm 5\%$. As described earlier, a scaled area approach estimates that glaciers in the Coe and Compass subbasins contribute about 38% of the discharge in Coe Creek during the study period. We speculate that some of the difference between the isotope and discharge

measurements for Coe Creek results from undermeasurement of glacier runoff by the discharge method, as described previously. However, without more conclusive measurements we assume that the 31% value serves as a lower bound for the estimated glacier contribution in the Coe Creek drainage.

[28] Isotopically determined glacier melt contributions to runoff in September and October were lower than that of August. This is consistent with the expected decline in autumn glacier runoff with decreasing seasonal air temperatures. For the sampling period, the measured monthly average daily air temperature at the Mt. Hood SNOTEL site was for 12°C in August decreasing to 9°C in September and 4°C in October (Mt. Hood SNOTEL site, <http://www.or.nrcs.usda.gov/snow/maps/sitepages/21d08s.html>, last accessed 23 June 2010).

[29] The total discharge of the Upper Middle Fork Hood River is the combined discharge from Eliot, Coe, Pinnacle and Clear creeks, which was measured as $6.5 \times 10^6 \text{ m}^3$ for the 28 day study period. According to gage measurements, Coe and Eliot Glaciers alone discharged $2.6 \times 10^6 \text{ m}^3$, or 41% of this overall flow. Combining the isotopically derived glacier

Table 4. Change in Eliot Glacier Runoff From SRM Simulations in Response to Decreasing Glacier Area and Increasing Temperature, Relative to 2007 Conditions, With the 2007 Ratio of Debris-Covered and Debris-Free Glacier Ice Maintained

SRM Simulation Temperature Change	SRM Simulation Glacier Area Change	August		September		August–September Percent Change From 2007
		Average Daily Glacier Runoff ($\text{m}^3 \text{d}^{-1}$)	Percent Change From 2007	Average Daily Glacier Runoff ($\text{m}^3 \text{d}^{-1}$)	Percent Change From 2007	
No change	No change	5.0×10^4	-	2.7×10^4	-	-
No change	-25%	3.7×10^4	-26	2.2×10^4	-19	-23
No change	-50%	2.6×10^4	-49	1.7×10^4	-36	-44
No change	-75%	1.3×10^4	-74	1.2×10^4	-54	-67
+1°C	No change	5.6×10^4	+12	3.2×10^4	+17	+14
+2°C	No change	6.2×10^4	+25	3.6×10^4	+31	+28
+3°C	No change	6.8×10^4	+38	4.0×10^4	+50	+41
+4°C	No change	7.5×10^4	+50	4.5×10^4	+65	+55
+1°C	-50%	2.9×10^4	-42	2.0×10^4	-25	-36
+2°C	-50%	3.2×10^4	-36	2.2×10^4	-15	-29
+3°C	-50%	3.5×10^4	-30	2.4×10^4	-7	-22
+4°C	-50%	3.8×10^4	-23	2.6×10^4	-1	-15

runoff from Eliot, Coe, and all other ice bodies in their sub-basins gives a total glacier contribution to the Upper Middle Fork Hood River of $4.8 \times 10^6 \text{ m}^3$ (73%) over the study period. The gage measurements, because they do not include the discharge of isolated glacier bodies, serve as a minimal estimate of glacier contributions, while the isotopic estimates could be considered an upper bound. Thus, glacier melt contributes 41%–73% of streamflow in the Upper Middle Fork Hood River.

3.2. Projecting Future Glacier Runoff Contributions to Late Summer Streamflow

3.2.1. SRM Model Calibration and Validation

[30] SRM-simulated discharge for Eliot Creek was calibrated with Eliot Glacier discharge data from 1 August 2007 to 29 September 2007. Runoff coefficients and recession coefficients were iteratively modified to fit measured daily discharge at the glacier terminus. Model accuracy was assessed and deemed successful with a Nash–Sutcliffe [Nash and Sutcliffe, 1970] coefficient of variation (R^2) of 0.89. The total volumetric difference between measured and modeled discharge was calculated at 0.4%.

[31] Because discharge records in these glacierized watersheds are limited to just one season, model validation was performed using the adjacent Coe Creek. Using Coe Glacier discharge from 10 August 2007 to 27 September 2007, the simulated versus measured discharge yielded a Nash–Sutcliffe coefficient of 0.81 and a volumetric difference of 5.4%. Thus, for the study period the model calibration parameters (derived from Eliot Creek) appear to be sufficiently transferable such that the model can explain over 80% of the variance in glacier discharge in Coe Creek.

3.2.2. Simulated Estimates of Discharge for Projected Changes in Glacier Covered Area and Temperature

[32] In order to assess potential effects of climate change, we explored the model output for changes in glacier-covered area and increased temperature. In our simulations of the effects of changes in glacier-covered area on glacier runoff, we initially ran SRM with present-day climatological mean temperature and precipitation values and compared discharge from the Eliot, Coe, and Compass subbasins (Figure 2). As expected, these modeled discharge data show that the Eliot subbasin, with the highest fraction of glacier cover, has the

highest mean discharge. The Compass subbasin, with only small ice bodies and snowfields, has the lowest discharge. These discharge differences become smaller later in the melt season as temperatures decline and precipitation becomes a major contributor to discharge.

[33] To isolate the effect of glacier recession on late summer glacier runoff component, SRM simulations were run under summer drought conditions by setting precipitation to zero. Using Eliot Glacier, we ran discharge simulations for glacier area reductions of 25%, 50%, and 75%, for no change in temperature. The size and thickness of the debris-covered portions of the glaciers were scaled according to the 2007 ratio of debris-covered area to nondebris-covered area. Reductions in total glacier area of 25%, 50%, and 75% result in decreases in glacier-contributed discharge of 23%, 44%, and 67% respectively (Table 4). There was an average decrease of 0.9% total glacier runoff for each 1% reduction in glacier area. The effects of glacier loss are smallest in the latter part of September, since glacier melt contributes a smaller portion of streamflow during that period.

[34] We assessed the sensitivity of the model to increased temperatures for the Eliot Glacier using the present-day glacier area and a 50% decrease in glacier area. The climatological daily mean precipitation was unchanged but the climatological mean temperatures were incrementally increased for each model run. For the present-day glacier area, a 1°C warming would increase the 2 month combined runoff from Eliot glacier by $3.6 \times 10^5 \text{ m}^3$, an overall increase of 14% from 2007 (Table 4). The effect of increased temperature is more pronounced for August than for September, yielding a mean increase in discharge of $6000 \text{ m}^3 \text{d}^{-1}$, while the percent increase is greater for September than for August.

3.2.3. Sensitivity of Model Results to Changes in Degree Day Factor, Glacier Debris Cover, and Accumulation Area Ratio

[35] As a means of bounding our estimates of potential future glacier melt contributions, we conducted a series of model sensitivity analyses. Using Eliot Glacier as our test case, we examined the effects of degree day factor, glacier debris cover, and accumulation area ratio on glacier melt-water discharge. We also examined the sensitivity of the model to the number of elevation zones defined in the model and details of that analysis are given by Phillippe [2008]. Climatologic daily mean values for temperature (25 years)

Table 5. Sensitivity of Modeled Eliot Glacier Runoff to Degree Day Factor^a

DDF Case	Accumulation Area DDF (mm °C ⁻¹ d ⁻¹)	Ablation Area Ice DDF (mm °C ⁻¹ d ⁻¹)	August–September Discharge (m ³)	Percent Change From Model Glacier Runoff
Eliot Glacier 30 year mean	4.4	7.1	2.3×10^6	-
DDF +1 SD	5.7	9.1	2.7×10^6	+15
DDF +2 SD	6.9	11.0	3.0×10^6	+30
DDF -1 SD	3.2	5.2	2.0×10^6	-15
DDF -2 SD	2.0	3.3	1.6×10^6	-30
Maximum DDF	6.1 [Singh <i>et al.</i> , 2000]	13.8 [Schytt, 1964]	3.1×10^6	+35
Minimum DDF	2.5 [Braithwaite and Olesen, 1988]	5.5 [Laumann and Reeh, 1993]	1.9×10^6	-19

^aThe standard deviation and mean DDF values were taken from previously published studies measuring the DDF of snow and ice on glaciers [Singh *et al.*, 2000; Zhang *et al.*, 2006a]. SD, standard deviation.

and precipitation (27 years) were used as baseline meteorological input to SRM.

[36] There are a number of site-specific glacier characteristics that affect DDF values such as slope, aspect, elevation, and albedo. These characteristics may change over time and therefore the associated DDF values for the glacier change as well. In our DDF sensitivity analysis, we assessed the effect of varying DDF on our model estimates of glacier runoff by simulating the mean DDF for ice (7.1 mm °C⁻¹ d⁻¹) and snow (4.4 mm °C⁻¹ d⁻¹), the mean DDF perturbed by 1 and 2 standard deviations, and the maximum and minimum DDF values reported in the literature (Table 5).

[37] Using the Eliot Glacier to test model sensitivity, we find that with each 1 standard deviation change from the mean DDF for the accumulation and ablation zones, the simulated discharge changes by approximately 16% (Table 5). The standard deviation and mean DDF values were taken from 21 previously published studies measuring the DDF of snow and ice on glaciers [see Singh *et al.*, 2000; Zhang *et al.*, 2006a]. Using the minimum and maximum reported DDF values results in glacier discharge changes of -19% and +35% respectively. The modeled discharge is linear with respect to DDF but changes in glacier characteristics such as the hypsometric distribution of the glacier, AAR, and extent of debris cover can combine to cause a potentially nonlinear response.

[38] Table 6 summarizes the sensitivity analysis of varying debris cover on the Eliot Glacier. The debris-covered area was scaled up and down in areal increments of 10%. As debris cover envelops greater areas of the ablation zone, glacier runoff decreases by 2.3% of its original value for every 10% increase in debris-covered area. When the debris-covered area decreases in 10% increments, the model shows ~2.9% increases in discharge. As shown in Table 6, failing to

Table 6. Sensitivity of Modeled Eliot Glacier Runoff to Changes in the Extent of Debris Cover

Change in Extent of Debris Cover Glacier	August–September Discharge (m ³)	Percent Change From Model Glacier Runoff
No change	2.3×10^6	-
10% increase	2.3×10^6	-2.3
20% increase	2.2×10^6	-4.6
30% increase	2.2×10^6	-6.9
10% decrease	2.4×10^6	2.8
20% decrease	2.5×10^6	5.7
30% decrease	2.5×10^5	8.6
Debris removed, glacier surface	3.3×10^6	41
Debris removed nonglacier	1.7×10^6	-27

incorporate the debris-covered section of the glacier into SRM as in the “debris removed nonglacier” case (e.g., assuming that there is no ice beneath the debris mantle) would result in a serious underestimation of glacier discharge. This case provides an estimate of the omission error that would result from inaccurately mapping debris-covered glacier as nonglacier. Conversely, using an ablation zone DDF (e.g., bare ice) rather than a debris cover DDF would significantly increase modeled discharge, as shown by the ‘Debris Removed Glacier Surface’ value.

[39] The Accumulation Area Ratio (AAR) is the ratio of the accumulation area of a glacier (at the end of the melt season) to the entire glacier area. AAR has been found to correlate with glacier mass balance [Dyrgerov, 1996; Hock *et al.*, 2007] and so is an important glacier hydrologic variable. Although SRM does not explicitly parameterize the AAR, the area above the snowline (roughly corresponding with the accumulation area) can be assigned a different DDF than the ablation area. Our sensitivity analysis of ±10% and ±20% in AAR showed that decreasing the AAR from its present value of 0.52 increases runoff, while increasing the AAR decreases runoff (Table 7). Reducing the AAR means that the accumulation area (which has a lower DDF value than the ablation area) occupies a smaller proportion of the glacier. As AAR decreases, a proportionately larger area of the glacier is assigned a higher DDF, thereby increasing runoff. A 10% decrease in the AAR increases runoff by 2.6%, while a 10% increase in AAR decreases runoff by 3.4%. These changes are not symmetric because of the area-elevation distribution of the glacier. The potential effect of the climate change on glacier melt production depends in part on glacier hypsometry and the relative rates of retreat of the ELA and glacier terminus (i.e., whether the AAR increases or decreases).

3.2.4. Effect of Combined Glacier Recession and Temperature Increase on Glacier-Contributed Runoff

[40] The final set of model runs combined the effects of glacier recession and temperature increases. To estimate the

Table 7. Sensitivity of Modeled Eliot Glacier Runoff to Changes in Accumulation Area Ratio

Change in Accumulation Area Ratio	AAR Value	August–September Discharge (m ³)	Percent Change From Model Glacier Runoff
No change	0.52	2.3×10^6	-
10% increase	0.57	2.2×10^6	-3.4
20% increase	0.62	2.2×10^6	-6.3
10% decrease	0.47	2.4×10^6	+2.6
20% decrease	0.42	2.5×10^6	+5.6

Table 8. Change in Combined Eliot, Coe, and Compass Glacier Runoff From SRM Simulations in Response to Decreasing Glacier Area and Increasing Temperature, Relative to 2007 Conditions, With the 2007 Ratio of Debris-Covered and Debris-Free Glacier Ice Maintained

SRM Simulation Temperature Change	SRM Simulation Glacier Area Change	August		September		August–September Percent Change From 2007
		Average Glacier Runoff ($\text{m}^3 \text{d}^{-1}$)	Percent Change From 2007	Average Glacier Runoff ($\text{m}^3 \text{d}^{-1}$)	Percent Change From 2007	
No change	No change	9.2×10^4	-	5.1×10^4	-	-
No change	-50%	5.9×10^4	-36	4.6×10^4	-10	-27
+1°C	-50%	5.4×10^4	-42	4.1×10^4	-19	-34
+2°C	-50%	5.9×10^4	-36	4.6×10^4	-11	-27
+3°C	-50%	6.5×10^4	-29	5.0×10^4	-3	-20
+4°C	-50%	7.0×10^4	-24	5.4×10^4	+5	-14

timing of future glacier recession for the entire watershed, recent rates of recession based on known locations of past glacier termini were extrapolated to future scenarios. Based on our 2007 GPS coordinates and a 1989 measurement of the glacier terminus position [Lundstrom *et al.*, 1993], the Eliot Glacier terminus has receded 284 m in the past 18 years, at an average rate of 15.8 m yr^{-1} . Based on this recession rate, Eliot Glacier will reach 50% of its 2007 extent in approximately 50 years. However, extrapolating the historical recession rate into the future likely produces a conservative estimate because recession has been accelerating over the last 50 years and changes in Eliot Glacier area have been shown to lag precipitation and temperature changes by 10–15 years [Jackson and Fountain, 2007; Lillquist and Walker, 2006]. These findings suggest that Eliot Glacier does not yet reflect the record high temperatures of the past 15 years [Lemke *et al.*, 2007], and its recession rate may accelerate in the coming years. On a regional basis, temperatures are projected to increase by a range of 1.1 – 6.4°C in the next 100 years [Lemke *et al.*, 2007]. Here, we coupled a 50% reduction in glacier area with temperature increases of 1 – 4°C , scenarios that are plausible within the next 50 years. We modeled these effects for the Eliot Glacier subbasin (Table 4) and also for all of the ice bodies in the Eliot, Coe, and Compass subbasins (Table 8).

[41] Results of our temperature sensitivity simulations for Eliot Glacier show a mean increase in total discharge of 14% per 1°C increase in temperature (computed by averaging the four temperature scenarios). However, as shown in Table 4, the effect of a 50% reduction in glacier area outweighs the effect of a temperature increase thereby leading to an overall decrease in glacier runoff. Looking at the combined effects of glacier recession and temperature increase runoff from glaciers in the Eliot, Coe, and Compass subbasins, our model results show that for a 2°C temperature increase and a 50% reduction in glacier area, the glaciers will produce only about $3.2 \times 10^6 \text{ m}^3$ of runoff for the combined months of August and September, a reduction of 27% compared with present-day conditions (Table 8). August would see the greatest declines in glacier-contributed discharge, with a 36% total decrease. In contrast, modeled glacier runoff in September shows only an 11% decline.

[42] Figure 4 shows modeled daily glacier runoff from both Eliot and Coe Glaciers for present-day glacier and climate conditions and for reduced glacier area and warmer temperature. For the case with reduced glacier area and increased temperature, we also show the sensitivities of glacier runoff to degree day factor, accumulation area ratio, and debris cover

extent. These daily results show that projected decreases in glacier runoff are not uniform over time with the largest simulated declines occurring in August. As discussed earlier, the model results are most sensitive to choice of degree day factor although compared with DDFs for other temperate glaciers, a 1 standard deviation error is more likely than a 2 standard deviation error. The 10% and 20% variations in accumulation area ratio result in modest changes in modeled glacier runoff, with the largest differences in August. The debris cover effects are rather small, even in August and thus uncertainty in this parameter is unlikely to lead to substantial error. Under conditions of shrinking glacier area, it is plausible that AAR would decrease and debris-covered area would increase, thus moderating their individual effects on glacier runoff. However, if both AAR and debris-covered area were to decrease, their combined effects would magnify glacier runoff.

4. Discussion

[43] There are several sources of uncertainty in our estimates of present and future glacier melt rates. Discharge estimates are subject to error caused by difficulties in accurately recording stage and measuring discharge in turbulent, sediment-laden streams. Isotopic discharge estimates required assumptions about equal isotopic values of meltwater from large glaciers and smaller ice bodies and isotopic uniformity of groundwater. For our model simulations with SRM it was necessary to generate runoff and recession coefficients in the calibration process. Because we had only one season of stream discharge data, these coefficients may not be representative for other years. Runoff coefficients for both rain and snow vary throughout the water year [Guo *et al.*, 2007], and these parameter changes were applied to match the decreasing runoff volumes over the 2 month period. In the model, the adjustment of the runoff coefficient provided for steep declines in discharge in late September. Runoff may rapidly decrease in September because of decreased temperatures and insolation slowing melt rate or depletion of englacial, subglacial, and firn storage late in the ablation season [Seaberg *et al.*, 1988; Hock and Hooke, 1993]. Because values for the degree day factors in each glacier zone were held constant, changes in the runoff coefficients may have been overemphasized. The temporal variability of the degree day factor may be better captured if direct measurements of ablation each day in both the ablation and accumulation zones were available. Another potentially useful approach is that of Carenzo *et al.* [2009]

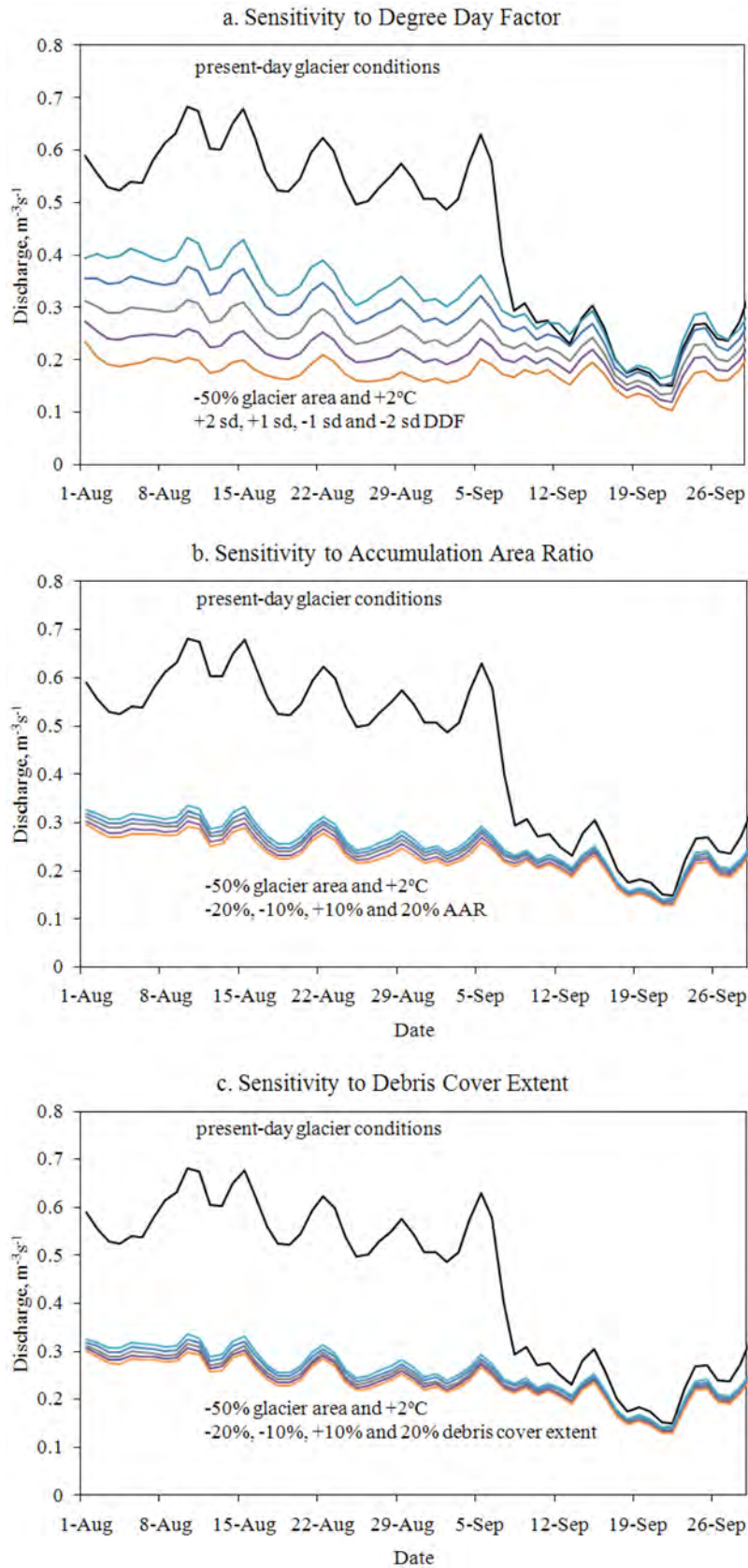


Figure 4. Modeled daily glacier discharge for the combined glaciers for present-day glacier conditions (solid black line) and for a -50% glacial extent with a 2°C temperature increase (gray line). Colored lines show modeled discharge for modified values of (a) DDF, (b) AAR, and (c) debris cover extent for -50% glacial extent and +2°C temperature (with modifications as described in Tables 5–7).

in which they optimized the parameters of a temperature-index model for transferability to other time periods and even to other locations.

[44] We note that SRM assumes an 18 h time lag between the daily temperature cycle and discharge. Because the glaciers in this study are small they are likely to have a shorter time lag. Thus, the longer assumed time lag may shift the modeled runoff by a day. However, this difference would not affect the seasonal results presented in our discussion.

[45] Properly calculating spatial variability is important in a study such as this because as glaciers recede they often become more shaded, and overall melting decreases. The spatial variability in melt due to local differences in slope, aspect, and shading is difficult to accurately model with a temperature-index model such as the SRM [Hock, 2003]. This is a shortcoming of this study's approach, and as a result the effect of glacier recession on glacier melt might be underestimated. This could be improved by incorporating solar radiation reception into SRM as performed by Brubaker *et al.* [1996], but these data were not available for Mount Hood. Future modeling efforts would likely benefit from incorporating radiation, albedo and shading effects such as the enhanced temperature-index model, developed by Pellicciotti *et al.* [2005] for Haut Glacier d'Arolla, Switzerland, which uses the hill-shading algorithm of Corripio [2003].

[46] Like many midlatitude glaciers, Eliot and Coe Glaciers have extensive zones of debris cover, which are underestimated on USGS quadrangles. This study emphasizes the importance of debris cover in determining the glacier's overall degree day factor and its consequent runoff. Eliot Glacier, with 42% of its area covered in sediment, is hydrologically equivalent to a clean glacier that is only 79% of its size. It is therefore critical that glacier melt studies use field observations, aerial photography, and/or advanced remote sensing techniques to accurately identify debris-covered ice. Our model sensitivity study shows that a 10–20% uncertainty in debris-covered area produces changes in simulated discharge of about 5–6%. While there have a number of studies documenting the recession rates of midlatitude glaciers [e.g., Driedger and Kennard, 1986; Dodge, 1987; Oerlemans, 2005; Lillquist and Walker, 2006], the limited number of studies on debris-covered zones show that they are less sensitive to climate change than debris-free glaciers [Mayer *et al.*, 2006; Mihalcea *et al.*, 2008].

5. Conclusions

[47] In data sparse glacierized watersheds, the combination of discharge measurements and isotopic tracers is a valuable approach for making quantitative estimates of present-day glacier melt contributions to streamflow and the temperature-index modeling approach, when accompanied by sensitivity analyses, can provide useful qualitative estimates of future glacier melt contributions to streams. As modeling techniques and data acquisition systems have become more sophisticated, modelers are arguing to focus more on input accuracy and less on input calibration. Striving for such a goal becomes a problem in basins with few meteorological and hydrological records. Even our study area on Oregon's signature mountain peak draining to a prosperous agricultural region lacked the extensive meteorological and discharge data necessary for robust understanding of parameter accuracy. Thus, studies

like ours that combine multiple methods with model sensitivity analyses allow us to make cautious, qualitative projections for potential future impacts. This represents a constructive approach for applied water resources problems in all but the most data-rich regions.

[48] The application of stable isotope tracers and discharge measurements show that late summer flows in the Eliot and Coe creeks are dominated by glacier runoff. Because the Upper Middle Fork Hood River is highly dependent on glacier runoff, the overall flow of the watershed is very sensitive to changes in temperature during the dry season, and less dependent on summer precipitation events. The decline of Mount Hood's glaciers is likely to cause considerable decreases in flow in the Upper Middle Fork Hood River. Although this study relied on only a single season of data, the temperature and precipitation were representative of the climatological mean values. Our model sensitivity simulations show that model uncertainties are highest for degree day factor but relatively modest for accumulation area ratio and debris-covered area. Daily runoff simulations suggest that, even with the uncertainties of the calibrated model, future glacier runoff would significantly decrease in August and slightly decrease in September.

[49] Decreased glacier-covered area and increased temperature influence glacier runoff in opposite directions. A decrease in glacier-covered area yields lower glacier runoff whereas an increase in temperature during the ablation season promotes greater runoff. The relative importance of each factor determines whether or not late summer glacier runoff increases or decreases in the future. Our SRM simulations showed that Eliot Glacier discharge increases 13% for every 1°C increase, but decreases 9% for every 10% decrease in glacier area. Thus, glacier runoff will be stable or increase if the glacier covered area decreases <15% for every 1°C increase in temperature. The recession of the Eliot Glacier in the last century already exceeds this rate; therefore its discharge has likely been decreasing over time and will continue to decrease.

[50] While glacier runoff contributions to discharge at the mouth of the Hood River are small relative to other sources, they are significant upstream where water is diverted for irrigation and hydropower purposes. Glacier runoff contributions to first- and second-order watersheds are ecologically important, affecting the timing, magnitude and variability of discharge and downstream hydroecological responses [Milner *et al.*, 2009]. Our case study is a good example of the important influence of glacier runoff on small to moderate watersheds, and it can serve as an example for similar areas such as the Upper Columbia River basin in British Columbia where glaciers have been diminishing and empirical trend analyses show declines in streamflow [Stahl and Moore, 2006]. The implications of continued changes in late summer streamflow are significant for water resources policy and management in the Pacific Northwest, especially considering the potential renegotiation process of the Columbia River treaty, which could begin as early as 2014.

[51] The experimental approach used here shows that stable isotope tracers can be effectively used to determine the glacier runoff contribution to streamflow. Such an approach has potential for wider scale applications in determining glacier runoff contributions, particularly as high throughput methods of isotopic analyses are becoming available. Determining glacier runoff contributions to streamflow is

a key part of an integrated hydroclimatologic monitoring approach in glacierized watersheds.

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