Glacier Impacts on Summer Streamflow in the Wind River Range, Wyoming

Jeb Bell[1]; Glenn Tootle, M.ASCE[2]; Larry Pochop[3]; Greg Kerr[4]; and Ramesh Sivanpillai[5]

Abstract: The Wind River Range (WRR) of Wyoming is host to approximately 63 glaciers. Extensive research has been conducted using remote imagery to estimate the recent area and volume changes of these glaciers with the goal of estimating the potential effects of these changes on watershed streamflow. Results show that the glaciers were mostly in recession since 1966, the beginning of the study period. The current research was performed to supplement results from the remote imagery analyses. In this paper, streamflows from glaciated and nonglaciated watersheds in the WRR for the period 1967–1992 were analyzed. The difference in July-August-September (JAS) watershed flow magnitude for the 26-year period between glaciated (Green River and Bull Lake Creek) and nonglaciated (East Fork River and Wind River) watersheds ranged between 8 and 23%. As expected, the effects of glaciers on local streamflows during JAS were shown to be much greater than that of ice melt alone. The influence of glaciers accounted for 23–54% of the late summer (JAS) flow in glaciated watersheds with approximately 2–12% because of loss of glacial mass, whereas the remainder of the increased flow was because of the glaciers decelerating the snowmelt runoff through internal storage/delayed release of liquid water. The glaciated watersheds provided a more stable source of streamflow because they displayed less year-to-year streamflow variability with coefficients of variation of 0.36 (Green River) and 0.29 (Bull Lake Creek) compared with the nonglaciated values of 0.55 (East Fork River) and 0.45 (Wind River). **DOI: 10.1061/(ASCE)HE.1943-5584.0000469.** *© 2012 American Society of Civil Engineers.*

CE Database subject headings: Glacial till; Snowmelt; Streamflow; Wyoming.

Author keywords: Glaciers meltwater streamflow.

Introduction

Glaciers of Western United States Alpine Mountains

The Wind River Range (WRR) of Wyoming is host to approximately 680 snow and ice bodies with 63 of these considered glaciers (Pochop et al. 1990). The east slope of the divide contains approximately 77% of the total glacier area, whereas the west slope contains the remaining area (Marston et al. 1991).

Hydrologists view glacier bodies as "frozen water reservoirs" that store water in the form of ice and release it at future times (Pochop et al. 1990; Fountain and Tangborn 1985; Jansson et al. 2003). Thus, watersheds with glaciers are shown to provide a more stable source of water than nonglaciated watersheds (Ferguson 1973; Fountain and Tangborn 1985; Braithwaite and Olesen 1988). Glaciers also affect the annual hydrograph by delaying seasonal runoff through internal storage of liquid water resulting from snow melt for release later in the year. Because it has been documented that most North American glaciers have been in recession in recent years (Nylen 2002; Granshaw and Fountain 2006), water planners have become more interested in their contribution to the local hydrologic system.

The presence of glaciers modifies streamflow in powerful, even unique ways important to both hydraulic and hydrologic engineering (Fernández et al. 1991; Fleming and Clarke 2005). The principal influences on streamflow are often contributions in volume during periods of recession, a delay of the maximum seasonal flow, storage of spring snowmelt in the form of liquid water for release later in the year, and a decrease in annual and monthly variation of runoff (Fountain and Tangborn 1985). The presence of glaciers results in meltwater contributions to streamflow during the late summer [(July, August, and September (JAS)] when snowmelt is decreasing, temperatures are still high, precipitation is generally low, and irrigation demand continues (Fountain and Tangborn 1985; Pochop et al. 1990). For the WRR, these are important considerations for both the west slope drainage Green River Basin and east slope drainage Wind-Big Horn River Basin (BRS 2001; States West Water Resources 2003).

Research Objectives

Mark F. Meier was the first person to study the glaciers around Gannett and Fremont Peak within the WRR in 1950. His study reported the surface area of Dinwoody Glacier to be approximately 3.47 km^2 (Meier 1951). Consequently, later studies by Marston et al. (1989) and Wolken (2000) determined the area to be 2.91 and 2.19 km^2, respectively. Along with the analysis of glacier surface ice, Marston et al. (1991) utilized aerial photography stereopairs to determine Dinwoody Glacier lost 0.064 km^3 of water equivalent between 1958 and 1983. Wolken (2000) updated the volume loss values by evaluating stereopairs from 1983 to 1994, which determined Dinwoody Glacier lost another 0.043 km^3. Finally, Thompson et al. (2011) used advancements in remote sensing analysis to evaluate glacier area change from 1966 to 2006.

[1]Graduate Research Assistant, Dept. of Civil and Architectural Engineering, Univ. of Wyoming, Laramie, WY 82071.

[2]Assistant Professor, Dept. of Civil and Environmental Engineering, Univ. of Tennessee, Knoxville, TN 37996 (corresponding author). E-mail: gtootle@utk.edu

[3]Professor, Office of Water Programs, Univ. of Wyoming, Laramie, WY 82071.

[4]Director, Office of Water Programs, Univ. of Wyoming.

[5]Research Professor, Wyoming Geographic Information Science Center (WyGISC) and Dept. of Botany, Univ. of Wyoming, Laramie, WY 82071.

Note. This manuscript was submitted on January 25, 2011; approved on June 24, 2011; published online on June 27, 2011. Discussion period open until September 1, 2012; separate discussions must be submitted for individual papers. This paper is part of the ***Journal of Hydrologic Engineering***, Vol. 17, No. 4, April 1, 2012. ©ASCE, ISSN 1084-0699/2012/4-521–527/$25.00.

The current study supplements these previous research efforts by performing a paired-watershed evaluation of glacier meltwater contributions to streamflow in the WRR. Similar to the research efforts of Fountain and Tangborn (1985) in the North Cascades of Washington and Fleming and Clarke (2005) in the southwestern Canadian subarctic, a paired-watershed analysis was completed for glaciated and nonglaciated watershed pairs in both the Green River Basin and the Wind-Bighorn River Basin. The extent to which glaciers affect the variability of local streamflow, specifically late summer values, was investigated.

Study Area and Background

Wind River Range

The WRR is a large, continuous mountain range running 200 km along the continental divide in west-central Wyoming. The barrier serves as the dividing entity for three major river basins in the American West: Missouri-Mississippi, Green-Colorado, and to a lesser extent the Snake-Columbia (Marston et al. 1991). The western portion of the divide drains water to the Green River, which is the largest tributary of the Colorado River. The Green River Basin in Wyoming utilizes approximately 0.51 km^3 (410,000 acre-ft) per year of water for agricultural crop irrigation, which is approximately 67% of the basin wide water allocation (States West Water Resources 2001). The east slope of the divide drains to the Wind-Big Horn River before its confluence with the Yellowstone River to the north and finally the Missouri River. The Wind-Bighorn River Basin in Wyoming utilizes approximately 1.44 km^3 (1,165,000 acre-ft) per year of water for crop irrigation, which is approximately 82% of the basin wide water allocation (BRS 2003).

Paired-Watershed Characteristics

Glaciated and nonglaciated watersheds are found on both the west and east slopes of the continental divide. For this study, the west slope glaciated Green River watershed is upstream of the U.S. Geological Survey (USGS) gauge No. 09188500, Warren Bridge on the Green River near Daniel, WY (Fig. 1). The watershed has an area of 1,212 km^2 and a mean elevation of 2,838 m.

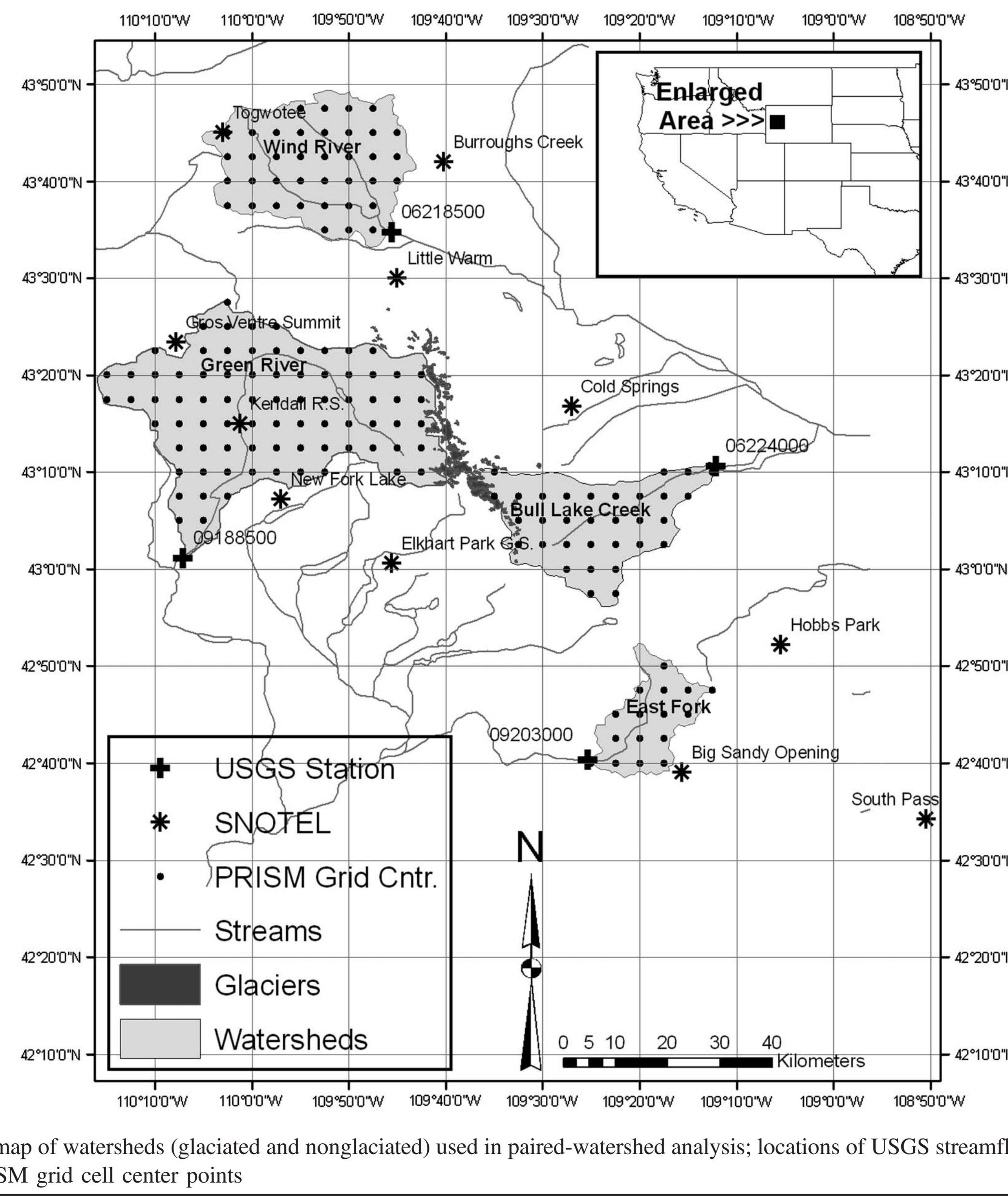

Fig. 1. Location map of watersheds (glaciated and nonglaciated) used in paired-watershed analysis; locations of USGS streamflow gages, SNOTEL stations, and PRISM grid cell center points

Thompson et al. (2011) estimated glacier area in the watershed to be 8.9 km^2 (0.7% of the watershed area) in 1966 and 7.4 km^2 (0.6% of the watershed area) in 1989. This glaciated watershed contains the following glacier or ice bodies: Mammoth, Stroud, Connie, Continental, Sourdough, and 12 unnamed glaciers or ice bodies. The nonglaciated East Fork watershed on the west slope is upstream of USGS gauge No. 09203000, East Fork River near Big Sandy, WY (Fig. 1). The watershed has an area of 205 km^2 and a mean elevation of 2,973 m, and is located south of the glaciated west slope Green River watershed.

For this study, the east slope glaciated Bull Lake Creek watershed is above USGS gauge No. 06224000, Bull Lake Creek above Bull Lake (Fig. 1). The watershed has an area of 484 km^2 and a mean elevation of 3,140 m. Thompson et al. (2011) estimated the glacier area in the watershed to be 12.5 km^2 (2.6% of the watershed area) in 1966 and 10.6 km^2 (2.2% of the watershed area) in 1989. This glaciated watershed contains the following glacier or ice bodies: Knife Point, Bull Lake, Upper Fremont, Sacagawea, Helen, and eight other unnamed glaciers or ice bodies. The nonglaciated Wind River watershed on the east slope is upstream of USGS gauge No. 06218500, Wind River above Dubois, WY (Fig. 1). The watershed has an area of 601 km^2 and a mean elevation of 2,697 m, and is located north of the glaciated east slope Bull Lake Creek watershed.

The paired watersheds have approximately the same mean elevation, are composed of the same soil types and strata, have the same land type characteristics, and have similar climate regimes. As stated in Fountain and Tanghorn (1985), when comparing glaciated and nonglaciated watersheds, it is critical that the precipitation characteristics of the watersheds are similar. Thus, snowpack accumulation must be similar, and, given the assumption that glacier meltwater will occur during the JAS season, the watersheds must display similar precipitation patterns during that season. Weather stations for comparison of climate conditions between glaciated and nonglaciated watersheds are limited. The Wyoming Climate Atlases provide a basis for further comparison of the climate conditions (Martner 1986; Curtis and Grimes 2004). Martner shows isolines for the period 1951–1980, which run parallel to the continental divide for mean annual temperature, mean annual precipitation, and mean annual snowfall. Curtis, who used PRISM to extrapolate data spatially, presents similar colored plots on the basis of 1961–1990 data. Spacing between Martner's temperature isolines appear to be slightly greater for the northern most Wind River watershed than for the other three watersheds. Otherwise, the atlases show that the annual temperature and precipitation and annual snowfall are similar for the glaciated and nonglaciated watersheds on both the east and west slopes of the WRR. Curtis and Crimes (2004) also plot the average number of days with thunderstorms (1901–1995). Generally, the majority of thunderstorms (convectional storms) occur during the summer months, and the isoline shows all four watersheds experience a similar number (approximately 35) of days with thunderstorms.

Tootle et al. (2006) evaluated Pacific Ocean climate variability and Wind River Range hydrology. They identified seven Natural Resources Conservation Service (NRCS 2011) snowpack telemetry (SNOTEL) stations: four on the west slope (Gros Ventre Summit, Kendall R.S., Elkhart Park G.S., and Big Sandy Opening) and three on the east slope (Little Warm, Hobbs Park, and South Pass) (Fig. 1). April 1st Snow Water Equivalent (SWE) values (inches) were converted to centimeters for a period of record of 1961–2000. The 5-year filter analysis resulted in stations (both west and east slope) having similar temporal relationships. Tootle et al. (2006) reported for the 40-year period of record (1961–2000), the average April 1st SWE for the four west slope SNOTEL stations was 36.6 cm (14.4 in.) with a standard deviation of 11.9 cm (4.7 in.), whereas the average April 1st SWE for the three east slope SNOTEL stations was 36.6 cm (14.4 in.) with a standard deviation of 10.7 cm (4.2 in.). The period of record (1967–1992) for the current research resulted in the average April 1st SWE for the four west slope SNOTEL stations was 35.9 cm (14.1 in.) with a standard deviation of 12.3 cm (4.9 in.), whereas the average April 1st SWE for the three east slope SNOTEL stations was 36.4 cm (14.3 in.) with a standard deviation of 10.7 cm (4.2 in.). Therefore, April 1st snowpack amounts were virtually identical for both the east slope and west slope.

Finally, a detailed analysis of JAS precipitation from 1967 to 1992 for the four watersheds was performed using the PRISM data set (PRISM Climate Group 2004). For each watershed, gridded PRISM data were accessed (center points for 4-km by 4-km grids are displayed in Fig. 1). For each grid center point in a watershed, the JAS precipitation (millimeters times 100, which was converted to centimeters) was determined for an individual year. For that individual year, all the grid center points in the watershed were then averaged to provide the average JAS precipitation. This process was repeated for each year for each watershed. This resulted in four vectors (one for each watershed) of yearly (1967–1992) average JAS precipitation. The average and standard deviation of JAS precipitation for the period of record for each watershed were very similar (Table 1), and box plots displaying median values [Green River (11.6 cm), East Fork River (11.7 cm), Bull Lake Creek (11.5 cm), and Wind River (11.7 cm)] were also very similar (Fig. 2). The nonparametric rank sum test was applied to the JAS precipitation to compare the four watersheds (Tootle et al. 2005). The method compares two independent data sets and determines if one data set has significantly larger values than the other data set. The p values ($p \gg 0.1$) showed that for every combination of the four watersheds, the JAS precipitation is statistically similar (Table 1). The temporal variability of the JAS precipitation in all four watersheds was very similar with intercorrelations ranging from 0.78 to 0.94 ($p < 0.01$) (Table 1). On the basis of the data collected from the Wyoming Climate Atlases, April 1st snowpack,

Table 1. JAS Precipitation for glaciated and Nonglaciated Watersheds

	1967–1992 JAS									
	Precipitation (cm)		Rank-sum test p value				Intercorrelation r value			
	Average	Standard deviation	GR	EFR	BLC	WR	GR	EFR	BLC	WR
Green river (GR)	12.6	5.1		0.72	0.96	0.60		0.82	0.78	0.94
East fork river (EFR)	11.8	5.0			0.52	0.88			0.91	0.83
Bull lake creek (BLC)	12.9	5.4				0.58				0.78
Wind river (WR)	11.7	4.1								

Note: Results (p values) from ran-sum test and inter correlation (r) values.

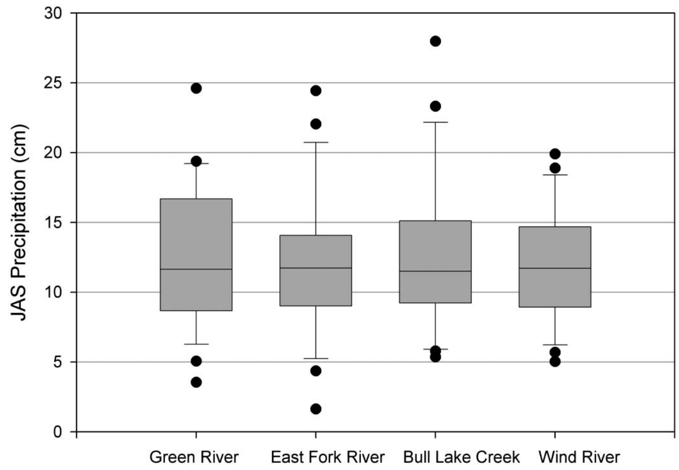

Fig. 2. Boxplots of median and quartile values of JAS precipitation (cm) for glaciated and nonglaciated watersheds for the period 1967–1992 on the basis of the PRISM data set

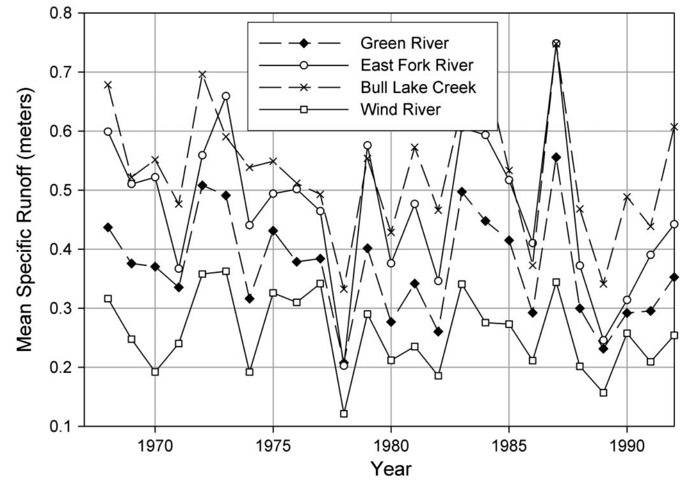

Fig. 3. Mean specific runoff (meters) for glaciated and nonglaciated watersheds for the period 1968–1992; runoff year is from April of previous year to March of current year

and JAS precipitation (PRISM), the writers are extremely confident that the watersheds experience similar climatic influences, and thus were suitable for the paired-watershed analysis.

Data, Methods, and Results

Streamflow Data

An unimpaired stream gauge station is defined as a station with minimal effects of anthropogenic uses including storage, diversion, and consumptive use. Unimpaired stations were identified using the Hydro-Climatic Data Network (HCDN) (Slack et al. 1993; Wallis et al. 1991). U.S. Geological Survey stream gauge information was obtained for all stream gages in this study from the National Water Information System (NWIS 2011).

Although there are many gages recording streamflow from both slopes of the WRR, it was difficult to find unimpaired stations with continuous data that overlapped from station to station (Barnett et al. 2010; Watson et al. 2009). The gauge stations were evaluated to meet the previously stated criteria, and four stations were selected: Green River above Warren Bridge near Daniel, WY; East Fork River near Big Sandy, WY; Bull Lake Creek above Bull Lake, WY; and Wind River near Dubois, WY, with a period of record of 1967–1992. For each gauge, the average monthly discharge data in cubic feet per second (cfs) was converted to volume (m^3). This was accomplished by taking the monthly runoff in cfs and multiplying $60\ \mathrm{s/min} \times 60\ \mathrm{min/h} \times 24\ \mathrm{h/day} \times 30\ \mathrm{days} \times 0.0283\ \mathrm{m^3/ft^3}$. A "runoff year" summation (April of current year to March of the following year) was computed for the annual hydrograph analysis. Given the time period of 1967–1992, the first runoff year would consist of April 1967 to March 1968, and the last runoff year would consist of April 1991 to March 1992. This time frame was selected to capture the spring freshet along with the late summer and early fall months into the same 12-month period. The mean specific runoff (Fountain and Tangborn 1985) was computed by taking the yearly volumes in cubic meters and then dividing by the watershed area in square kilometers—km^2 and then multiplying by $1\ \mathrm{km^2}/1{,}000{,}000\ \mathrm{m^2}$ to get meters, which was plotted for each watershed (Fig. 3).

Paired-Watershed Analysis (Glaciated versus Non-Glaciated)

The methods of Fountain and Tangborn (1985) in the North Cascades of Washington and Fleming and Clarke (2005) in the Canadian subarctic were the basis of the current research. Many of the same techniques used in these previous research efforts were used to quantify and qualify the effect of glaciers on WRR streamflow. The mean specific runoff, which is the runoff year annual runoff divided by the watershed area (Fountain and Tangborn 1985), was plotted and visually inspected for each watershed (Fig. 3). The mean specific runoff was implemented to remove differences resulting from scaling differences between the watersheds. This inspection verified that the watersheds follow the same temporal trends in streamflow on a year-to-year basis.

Annual Hydrograph Implications

The monthly volumes for each watershed gauge station were converted to a percentage of annual streamflow (Table 2, Fig. 4). This procedure allowed for two separate but important analyses: the qualitative observation of the hydrograph to determine how glaciated watersheds compare to the nonglaciated watersheds, i.e., higher late summer (JAS) percentages in the glaciated watershed, and also the quantitative comparison between the different watersheds to determine the glacier streamflow contribution.

The effect of glaciers on downstream flow was determined through the paired-watershed analysis with minor assumptions. The first was that a majority of glacial melt occurs during the late summer months (JAS), which was affirmed by Meier (1969), Fountain and Tangborn (1985), and Thompson (2009). Also, the comparison model was assumed to be macroscopic, meaning losses from groundwater recharge and evaporation were considered negligible. On each side of the continental divide, the percentage of JAS streamflow (JAS%) for the glaciated and nonglaciated watersheds ranged from 20 to 43% (Table 2).

To determine the percentage of late summer (JAS) streamflow attributed to the presence of the glaciers, the nonglaciated JAS% is subtracted from the glaciated JAS% and this value is then divided by the glaciated JAS%. For example, the nonglaciated Wind River Rane JAS% (27.1%) is subtracted from the glaciated Bull Lake Creek JAS% (43.4%). This value (16.3%) is then divided by the glaciated Bull Lake Creek JAS% (43.4%), which resulted in 37.6% (Table 3). The approximate percentage of late summer (JAS)

Table 2. Monthly Percentages of Annual Streamflow by Watershed

Watershed	Glaciers	Monthly streamflows expressed as a percentage of total annual runoff												JAS (%)
		Jan.	Feb.	Mar.	Apr.	May.	June	July	Aug.	Sept.	Oct.	Nov.	Dec.	
Green River	Yes	1.9	1.9	2.1	4.6	16.5	30.0	20.6	9.2	5.2	3.5	2.5	2.0	35.0
East Fork River	No	0.8	0.8	0.9	2.8	24.4	46.2	14.9	3.3	1.9	1.8	1.3	0.9	20.1
Bull Lake Creek	Yes	0.8	0.7	0.7	1.8	13.1	33.9	25.4	11.9	6.1	2.9	1.6	1.0	43.4
Wind River	No	2.7	2.7	3.1	5.0	16.6	32.1	15.4	6.7	4.9	4.4	3.5	2.9	27.1

Note: Estimated monthly percentages and summation of JAS percentages.

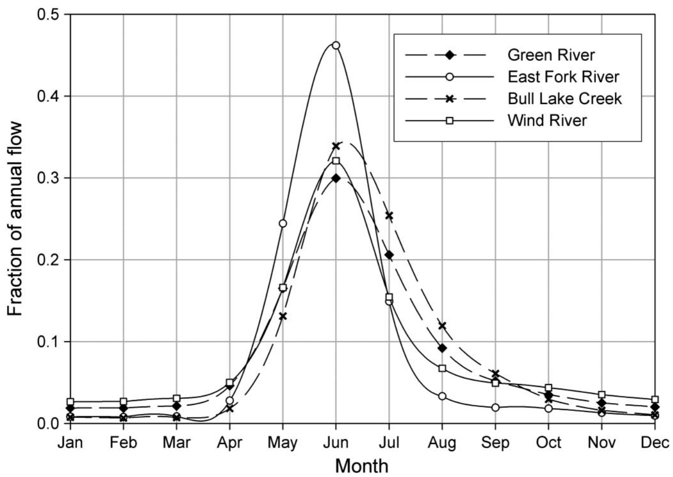

Fig. 4. Monthly streamflow percentages (line graph) of the annual hydrograph for glaciated and nonglaciated watersheds for the period 1967–1992

Table 3. Effect of Glaciers on Summer Season (JAS) Streamflows

	Green River %JAS	Bull Lake Creek %JAS
East Fork River	42.6%	53.7%
Wind River	22.6%	37.6%

Note: Estimated percentages of JAS streamflow contributions by the glaciers.

streamflow attributed to the presence of the glaciers ranged from 23 to 54% (Table 3). The JAS glacier contribution was not strictly icemelt but included delayed snowmelt runoff through internal liquid water storage, with the latter being considerably greater than the former.

Streamflow Variability

The year-to-year runoff variability of different watersheds for the late summer months of JAS was determined through a dimensionless coefficient of variation (CV). It was calculated by summing the monthly average discharge volumes of the 3 months (JAS), then finding the average and standard deviation across the 26-year period of record and finally taking the ratio of standard deviation and the average (Fountain and Tangborn 1985). The coefficient of variation equation is given as Eq. (1)

$$\mathrm{CV} = \frac{\sigma}{\bar{X}} \quad (1)$$

The hypothesis was that greater glacierization results in lower coefficient of variation values. The JAS CV for the Green River and Bull Lake Creek watersheds were 0.36 and 0.29, respectively. The JAS CV for the East Fork River watershed was 0.55, whereas the JAS CV for the Wind River watershed was 0.45, thus supporting the research hypothesis.

Along with the coefficient of variation for the 26-year period, the yearly deviation from the period mean was calculated using Eq. (2), where $\bar{y}_i$ = year average of JAS flow, and y_i = JAS flow of year i

$$\mathrm{dev} = \frac{\sqrt{(y_i - \bar{y}_i)^2}}{\bar{y}_i} \quad (2)$$

The decomposition of the variability to a yearly basis allowed the determination of which specific years accounted for the difference between glaciated and nonglaciated watersheds. In addition, the deviation from the mean values could be compared with standardized snow water equivalent values from historic April 1st snow water equivalent data (Aziz et al. 2010; Hunter et al. 2007), obtained from the NRCS for SNOTEL stations in the WRR that were spatially near the watersheds (Fig. 1). For the Green River watershed (Kendall R.S. and New Fork Lakes) and Wind River watershed (Togwotee Pass and Burroughs Creek), two stations for each watershed were identified to be in or adjacent to the watershed. One station each was identified for the East Fork River watershed (Big Sandy Opening) and Bull Lake Creek watershed (Cold Springs). The hypothesis was that extremely high values in the JAS yearly deviation (drought years) would correlate to large negative values in the snow water equivalent values. However, when calculating Eqation. (2), one must identify low JAS streamflow years from high JAS streamflow years given both extreme events will result in high yearly JAS deviation values. Therefore, for each of the four watersheds, the lowest quartile (i.e., the five lowest JAS streamflow years for the 26-year period of record) of streamflows and the corresponding years were determined. For the Green River, East Fork River, and Bull Lake Creek watersheds, the low flow years were identical (1977, 1981, 1985, 1988, and 1992) (Table 4). For the Wind River watershed, four of the five years (1977, 1981, 1988, and 1992) were the same as the other three watersheds with 1969 being added to the Wind River data set (Table 4). The identification of droughts in similar years supports that the four watersheds are climatically similar. Next, April 1st Snow Water Equivalent data for each of the six stations, for the same period of record (1967–1992) were standardized to a mean of 0 and standard deviation of 1 (Table 4). As displayed in Table 4, the low JAS streamflow years consistently corresponded to low April 1st snow water equivalent, with average snowpack values being approximately 1 standard deviation below the mean.

Summary and Discussion

The paired-watershed analysis determined glaciers affect the hydrology of the WRR range in multiple ways. As previously presented, historical climate data resulted in the four watersheds

J. Hydrol. Eng. 2012.17:521-527.

Table 4. Identification of Lowest Quartile Streamflow Years for each Watershed with the Yearly Deviation of the Mean and Corresponding April 1st Snow Water Equivalent Value

	Green river			East fork river		Bull lake creek		Wind river		
Year	Dev	SWE[a]	SWE[b]	Dev	SWE[c]	Dev	SWE[d]	Dev	SWE[e]	SWE[f]
1969								0.44	0.0	−0.6
1977	0.43	−1.6	−1.6	0.65	−1.8	0.31	−1.7	0.68	−1.8	−2.1
1981	0.38	−1.5	−1.5	0.61	−1.3	0.29	−0.5	0.44	−1.1	−1.1
1985	0.37	−0.6	−1.2	0.49	−0.4	0.55	−1.0			
1988	0.63	−0.5	−0.9	0.85	−1.2	0.49	−1.1	0.59	−0.7	−0.1
1992	0.52	−1.4	−1.0	0.59	−1.3	0.46	−0.7	0.55	−1.7	−1.0
Average	0.47	−1.1	−1.2	0.64	−1.2	0.42	−1.0	0.54	−1.1	−1.0

Note: Listings of SNOTEL stations; dev = yearly deviation of the mean; SWE = April 1st snow water equivalent standardized to a mean of 1 and standard deviation of.
[a]Kendall R.S.
[b]New Fork Lake.
[c]Big Sandy.
[d]Cold Springs.
[e]Togwatee.
[f]Burroughs Creek.

displaying similar climatic characteristics including JAS precipitation (Table 1, Fig. 2). Additionally, it was determined that all four watersheds displayed the same temporal trends (Fig. 3). On the basis of a visual observation of the annual hydrograph percentages by month (Table 2, Fig. 4), the higher percentage of annual streamflow occurs in the late summer months (JAS) in glaciated watersheds when compared with nonglaciated watersheds. The higher the percentage of glaciated area in the watershed, the higher the percentage of late summer flow. The monthly distribution (percentage) of streamflow (Table 2) revealed that glaciated watersheds contributed more JAS streamflow than nonglaciated watersheds. The Bull Lake Creek watershed had the highest percentage of glacier area (±2.4%) and 43% of the annual streamflow occurred during JAS, whereas the Green River watershed (±0.7% glacier area) had approximately 35% of the annual streamflow occurring during JAS. Conversely, the nonglaciated Wind River and East Fork River watersheds resulted in approximately 27 and 20%, respectively, of the annual flow during JAS. Thus, the difference in JAS flow percentage for the 26-year period between glaciated and nonglaciated watersheds ranged from 8 to 23% (Table 2). The percentages are a reflection of the mean specific runoff and thus should not reflect area difference between the watersheds. However, the monthly delay in peak discharge from glaciated watersheds was not observed in this study as it was in the Fleming and Clarke (2005) study (Fig. 4). The peak discharge in both glaciated and nonglaciated watersheds occurred during the month of June, which indicates the main driving factor in all watersheds in the WRR was spring snowmelt. A smaller delay, on the order of weeks, appears to be present in the glaciated watersheds, especially the Bull Lake Creek watershed (Fig. 4). Also, the annual hydrograph shows the falling limb was delayed later into the year for the glaciated watersheds as a result of glaciers temporarily storing liquid water internally for release later in the year (Tangborn et al. 1975). The combination of this phenomena and actual loss of glacier mass increased late summer flow in the glaciated watersheds as compared with the nonglaciated watersheds.

The contribution of the glaciers to summer (JAS) seasonal flows ranged from 23 to 54% (Table 3). The streamflow contribution of the glaciers was more likely attributed to delayed snowmelt runoff through internal liquid water storage than actual ice melt (mass loss) of the glaciers. Williams et al. (2009) performed a stable isotope study of the Dinwoody Creek watershed, a glaciated watershed (5.5% glacier area as of 1989) adjacent to the Bull Lake Creek watershed, and determined glacier meltwater accounts for approximately 53–59% of late summer flows. Therefore, the estimates presented by the Williams et al. (2009) study were similar to the results of the paired-watershed analysis. The Dinwoody Creek watershed was not included in the present study because of the lack of streamflow data during a portion of the study period.

The JAS CV for the glaciated (Green River = 0.36; Bull Lake Creek = 0.29) and nonglaciated (East Fork River = 0.55; Wind River = 0.45) watersheds supports there was less variability in late summer flows in glaciated watersheds than nonglaciated watersheds. Table 4 indicates that strong negative trends in the snow water equivalent values correspond to high values in the yearly deviation from the mean (dev) for drought years. Also, the nonglaciated watersheds show the largest range of yearly deviations (dev) independent of aspect, which reinforces the idea that glaciated watersheds can be considered more stable sources of runoff. For example, the range for the nonglaciated East Fork River was 0.02–1.15 and nonglaciated Wind River was 0.00–0.91, whereas the range for the glaciated Green River was 0.09–0.69 and the glaciated Bull Lake Creek was 0.01–0.55. A similar pattern in variations was discovered by Fountain and Tangborn (1985) in the North Cascades of Washington state. Compared to the North Cascades study, the results also support the idea that variation within nonglaciated watersheds is much larger in arid climates. The signal discovered by the coefficient of variation for the JAS period does not hold true for an annual analysis. The "drowning out" of this signal is assumed to be associated with the small value of glacier cover and similar streamflow characteristics during winter base flow months and the spring freshet. The Fleming and Clarke (2005) study also determined that year-to-year variability in the hydrologic shifts is very subtle, and thus the coefficient of variation was dominated by short-term fluctuations.

Conclusions

Glaciers have the ability to dramatically affect the hydrologic characteristics of a watershed by storing ice and snowmelt for release at a future time. The effects on streamflow are especially important during the late summer months of July, August, and September. The current paired-watershed analysis shows the glaciated watersheds in the Wind River Range have a higher percentage of yearly streamflow occurring during JAS and less year-to-year late season runoff variability than nonglaciated watersheds. During the 1967–1992 period, a period of mostly glacier recession, the watersheds studied the glaciers provided 23–54% of JAS streamflow through a combination of delayed release of snowmelt and icemelt. Thompson (2009) showed that the percentage of streamflow contributed by glacial icemelt alone was 2–12% for the WRR glaciated watersheds. Thus, as long as the glaciers exist, the contributions of glaciers to late season watershed streamflows are significant whether the glaciers are retreating or assimilating. Because of the relative magnitudes of the watershed versus basin streamflows, the most significant effects of the glaciers are on local watershed streamflow magnitudes and variability and not on basin streamflows.

Acknowledgments

This research was supported by the University of Wyoming Water Research Program funded jointly by the United States Geological Survey (USGS), the Wyoming Water Development Commission, and the University of Wyoming. Additional support was provided by the University of Tennessee, the USGS 104B program, and the National Science Foundation Paleo Perspectives for Climate Change program award AGS-1003393. The writers wish to thank the three anonymous reviewers as their comments and suggestions greatly improved the quality of the manuscript.

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