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Assessing streamflow sensitivity to temperature increases in the Salmon River Basin, Idaho

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ABSTRACT

Increased temperatures are occurring in the Salmon River Basin (SRB) of Idaho and are anticipated to continue increasing in the future, leading to complex changes in climate and water resources. To address these concerns, the objective of this study was to evaluate streamflow changes/sensitivity when temperatures increase. A hydrological model, the Variable Infiltration Capacity (VIC) model, was applied to simulate streamflow under thirty temperature increase scenarios (i.e., rising 0.1 °C per step to 3 °C). It was found that the annual mean streamflow decreased whenever temperatures increased in the SRB. Streamflow increases in winter and decreases in spring and summer but is barely affected by temperature in autumn. On a monthly basis, streamflow responses varied in response to rising temperatures. When temperature increased, the streamflow increase occurred from November to February, and it decreased from May to July. The analysis also discovered linear relationships between rising temperatures and streamflow changes throughout the year, with the exception of June and July, which revealed logarithmic correlations. Results obtained by daily streamflow analysis showed that center time occurred 10–30 d earlier when temperatures increased 2 °C and 15–45 d earlier when temperatures increased 3 °C. Finally, the Richards–Barker Index (R–B Index), a flashiness index, also increased with rising temperatures, and a higher R–B Index causes bank erosion problems. Changes in the streamflow due to the temperature increases have a significant implication both for the water management and ecological processes.

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1. Introduction

Climate changes have occurred in the Salmon River Basin (SRB) of Idaho and are anticipated to continue in the future, leading to complex changes in the climate and water resources of the region (Intergovernmental Panel on Climate Change (IPCC), 2008). Temperature increases directly affect the hydrology in the SRB (Easterling et al., 2000; Portmann et al., 2009). The water resources in the SRB are more sensitive to temperature increases compared to other regions in the United States due to its unique topographic and climatic features (Kunkel and Pierce, 2010). Developing an improved understanding of the impacts of temperature increases on the SRB's water resources is imperative, because they may affect ecological, economical, and political conditions in the SRB. However, there is a general lack of understanding of the ways in which rising temperatures influence streamflow at the spatial and temporal scales in the SRB.

It is widely accepted that changes in climate have resulted in global warming. The most recent Intergovernmental Panel on Climate Change (Intergovernmental Panel on Climate Change (IPCC), 2009) forecasted that global temperatures will rise between1.8 and 4.0 °C by 2090 globally. Climate models consistently predict progressive temperature increases (2.2–2.8 °C) in the western United States over the next decade (Muttiah and Wurbs, 2002; Mote et al., 2003; Christensen and Lettenmaier, 2007). For example, the temperature will be up to 3 °C over the next 50 to 100 yr over the Colorado River Basin (McCabe and Wolock, 2002). The most significant effects of rising temperatures may be the alteration of hydrological cycles and streamflow regimes (Bronstert et al., 2002, 2009). Miller et al. (2011) projected decreased runoff in 2 (Gunnison and San Juan River basins) of 3 (Gunnison, San Juan, and Green River basins) headwater basins of the Colorado River through the year 2099. Annual runoff across the Washington State is projected to increase by 2–3% by 2040s driven by projected temperature increase in Mantua et al. (2010).

Many studies have investigated the impact of temperature increases on water resources for other specific regions (Nijssen et al., 2001; Luce and Holden, 2009; Ma et al., 2010). Some studies discovered that the streamflow increased along with the temperature increases. For example, Nijssen et al. (2001) studied the effect of

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temperature increase on the global large rivers; results indicated that the water resource response to the temperature increases was different for the coldest snow dominated basins than for these rainfall dominated basins. However, annual flow volumes are increased due to the temperature increases for the nine large rivers. Arnell (2003) also concluded that the streamflow increases due to the temperature increases in Europe. In contrast, some studies in different regions have found that annual streamflow has been reduced because of temperature increases. For instance, a study by Chiew et al. (2009) predicted a decrease in future runoff for Australia because of a 0.9 °C rise in global temperatures. However, some studies showed mixed annual streamflow in response to the temperature increases or temperature increases (Chen et al., 2006; Walvoord and Striegl, 2007; Beyene et al., 2010). Beyene et al. (2010) assessed the hydrological response to the temperature increases in the Nile River, and results indicated that streamflow will increase from 2010 to 2039, and decline from 2040 to 2099.

The impact of temperature increases on streamflow is given attention because of its close relation to the timing of the water resource cycle (Bronstert et al., 2002). Changes in temperature influence the magnitude and temporal pattern of streamflow, which may affect the management of water resources (Lettenmaier et al., 1994). Due to increased temperatures, streamflow now occurs 1 to 4 weeks earlier than it did in the middle of the 20th century (Stewart et al., 2005; Jefferson et al., 2008). Because the exact timing of streamflow and snowmelt in the aforementioned studies is difficult to determine, this study calculates *center time* (CT), defined as the date when half of the yearly flow has passed (Stewart, et al., 2005), to better evaluate shifts in the hydrography of the SRB.

Although much has been written outlining the effects of future temperature increases on streamflow, little previous work has been done to investigate the sensitivity of streamflow to temperature increases, and to quantify hydrologic sensitivity. Furthermore, the majority of research has focused on annual or seasonal streamflow, and to a lesser degree, on monthly and daily streamflow. This study is a comprehensive analysis of increased temperatures' impacts on streamflow at annual, seasonal, monthly, and daily scales and is important both from science and water management perspectives. The daily changes in streamflow patterns strongly affect the timing of snowmelt. The increases in temperature have immediate as well as long-term effects on river systems. At daily time scales, changes in weather can lead to changes in the incidence of floods. The streamflow analysis in a daily scale is able to capture the timing of the snowmelt, which will have potentially far reaching implications both for water management and ecological processes in the SRB.

The impacts of the temperature increases to the streamflow were certainly underestimated because reservoir storage in the rivers was not considered in these studies as mentioned above. However, the SRB poses the unique nature of an unimpaired river basin (lacking of anthropogenic characteristics). Anthropogenic modifications like large dams, reservoirs, and changes in land uses can hinder the understanding of temperature increases' impact on water resources, but the SRB has not been subjected to any major construction and is widely recognized as an unimpaired watershed.

Over the past several decades, streamflow has tended to increase with increasing precipitation and decrease with decreasing precipitation (Dai and Trenberth, 2002). However, temperature is the variable that is most often perturbed in hydrologic simulations of climate change (Christensen and Lettenmaier, 2007) and has very complicated impacts on streamflow. An overall increase in temperature has lowered the snow line and ultimately influenced the timing and magnitude of flow in the main stem of the Salmon River and its tributaries, which is of interest from both a water resource management and an ecologic sustainability perspective. However, the ways in which rising temperatures affect the streamflow in the SRB at the temporal and magnitude scales are still unclear. This paper seeks to address the aforementioned issues by posing the following questions:

(1) How can we create long-term, high-resolution streamflow models for the SRB? Modeling streamflow is important because existing observations are too sparse and cover too short of a time span, limiting meaningful associations between streamflow and temperature increases in the SRB;
(2) What are the impacts of temperature increases on annual streamflow in the SRB? A detailed investigation of shifts in annual streamflow patterns due to temperature increases will provide a sound base for ecological studies concerning streamflow, such as the lifecycle of salmon and trout and the overall health of the aquatic environment;
(3) How will rising temperatures affect seasonal streamflow in the SRB? This question is important for farmers in knowing when to retain excess water for irrigation as temperatures change;
(4) Streamflow in which month is likely to be the most sensitive to rising temperatures in the SRB? This information will become increasingly important for the design of monthly water supply and agricultural plans for both farmers and water managers;
(5) How many days will the CT of streamflow shift when temperatures increase in the SRB? Being able to pinpoint the timing of this shift is a critical aspect of flood control design given the effects of earlier snowmelt from temperature increases.

The rest of this paper is organized as follows: Section 2 describes the study area, Section 3 outlines the description of the hydrologic model used in the study, Section 4 presents research setups and results including the evaluation of the impacts of rising temperatures on annual, seasonal, monthly, and daily streamflow, and Section 5 presents final conclusions and further discussion.

2. Study area

The study area was the SRB (Fig. 1), located in north-central Idaho. The Salmon River is an unimpaired river system with forest headwaters that originate at a high elevation in a mountainous environment. It flows 684 km downstream from the snowy climate, losing 2134 m in elevation as it winds through a vast wilderness before reaching a rain-dominated region near its confluence with the Snake River.

The Salmon River is the largest tributary of the Snake River, contributing approximately 20% of the total Snake River water but containing only 12% of the Snake River drainage area. The Salmon River's most valuable contribution, however, is that it produces 45% of all the salmon and contains more than 70% of the remaining salmon habitat in the entire Columbia River Basin. It also grows through the contributions of several large tributaries including the East Fork, Pahsimeroi, Lemhi, Middle Fork, South Fork, and Little Salmon Rivers. The Salmon River's flow is extremely seasonal because snowmelt comprises the majority of its runoff (Crozier et al., 2008). Precipitation varies by season from low (June through August) to high (October through April). Average annual air temperatures range from a high of 12 °C in the lower basin to −6 °C in the upper basin.

The Salmon River provides critical water supplies for agriculture and ecosystems throughout Idaho. Cities, villages, and towns rely upon the river for drinking water. As such, a recent analysis of the region was conducted by the Intergovernmental Panel on Climate Change (IPCC) (2009) determined that it is susceptible to current projected climate warming trends. Increasing temperatures in mountainous environments have driven and will continue driving the snow line upward, ultimately influencing the timing and magnitude of the streamflow.

3. Model setup

3.1. Variable Infiltration Capacity (VIC) model

For most rivers, observation data—especially long-term, spatially distributed records—cannot be obtained; therefore, hydrological models

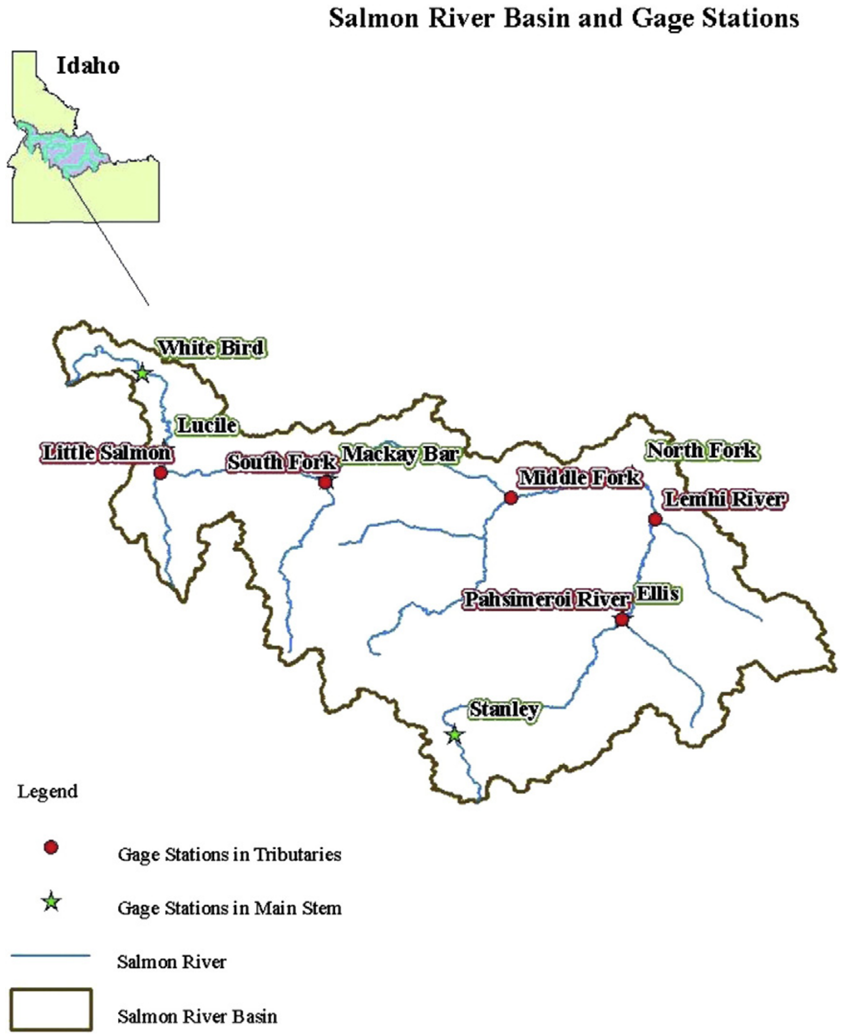

Fig. 1. Salmon River Basin and gage stations.

that simulate water cycle processes throughout basins are considered useful research tools. In this study, the Variable Infiltration Capacity (VIC) model was used to simulate streamflow for a daily time step and at 1/16° spatial resolution along with an offline routing model for the period of 1976 to 2005 (Lohmann et al., 1998). A model calibration was performed by comparing the modeled streamflow and observed streamflow in the study.

The VIC model is a macroscale, grid-based water and energy balance model, which has been successfully applied to many large river basins with practical results (Abdulla et al., 1996; Maurer et al., 2001; Lakshmi et al., 2004; Tang and Piechota, 2009). Distinguishing features of the VIC model include the subgrid variability in soil moisture, land surface vegetation, precipitation, and topography in use of the elevation band. In this study, the VIC model consisted of three soil layers: 10 cm deep for Layer 1, 30 cm deep for Layer 2, and 100 cm deep for Layer 3. Surface runoff was generated in the upper two layers by a variable infiltration curve, and baseflow was produced from the bottom layer (Todini, 1996).

The VIC model input data sets include meteorological, soil, and vegetation data. The meteorological data include temperature, precipitation, and wind. The 1/16° spatial resolution meteorological data sets were developed from the 1/8° resolution data by Maurer et al. (2002) (http://www.engr.scu.edu/~emaurer/gridded_obs/index_gridded_obs.html). The Symap algorithm (Shepard, 1984) was used to interpolate the 1/8 resolution data to the 1/16° data sets in the current study. The 1/8° datasets were developed from observation data following methods described in Maurer et al. (2002). The observation temperature and precipitation were obtained from the National Climatic Data Center's Cooperative Observer station data (Maurer et al., 2002). Daily surface wind speeds were obtained from the National Centers for Environmental Prediction/National Centers for Atmospheric Research (NCEP/NCAR) reanalysis project (Kalnay et al., 1996).

The 1/16° soil characteristics used in the current research were interpolated from the 1/8° data sets in the Land Data Assimilation System (LDAS) (Maurer et al., 2002) project. The 1/8° resolution soil datasets are aggregated from the 1-km resolution dataset produced by White and Miller (1998) maintained by the Earth System Science Center. Soil characteristics data sets include field capacity, wilting point, saturated hydraulic conductivity, soil types, and so on.

The VIC model allows for different types of vegetation and land cover (Liang et al., 1994). The land cover characterization was obtained from the Land Data Assimilation Scheme based on the University of Maryland's global vegetation classifications (Hansen and Reed, 2000). The 1/16° land cover data sets were developed based on the 1/8° resolution land cover data (Maurer et al., 2002) by identifying the land cover types present in each 1/16° grid cell and the proportion of the grid cell occupied by each in the current studies. Vegetation parameters such as architectural resistance, minimum stomatal resistance albedo, roughness length, zero-plane displacement, rooting depth, and fraction were specified for each individual vegetation class.

Once the cells' characteristics were defined in the model and meteorological forcing was applied, each cell produced water as both runoff and base flow. These contributions were summed up downstream using a routing model (Route 1.0) that generated streamflow and transported grid cell surface runoff and baseflow produced by the VIC model to the outlet of that grid cell, and then into the river system (Wood et al., 1997). A more thorough description of the VIC model is given in Liang et al. (1994) and Lohmann et al. (1998).

3.2. Model calibration

The model calibration was performed by comparing observed streamflow to simulated streamflow. Streamflow stations were identified by National Water Information System web data retrieval from the United States Geological Survey. The monthly streamflows over a 30-yr period (1976–2005) at four stations in the SRB—the Stanley, Salmon, Little Salmon, and White Bird stations—were used for calibration (Fig. 1).

The VIC curve and baseflow curve are the two governing curves of the VIC model. The infiltration parameter (b_{inf}) and the maximum infiltration capacity define the shape of the VIC curve. Four parameters define the shape of the base flow curve: (1) the maximum baseflow that can occur from the third soil layer in mm/d (D_{smax}); (2) the fraction of D_{smax} wherein nonlinear (i.e., rapidly increasing) baseflow begins; (3) the fraction of the lowest soil layer's maximum soil moisture wherein nonlinear baseflow occurs; and (4) the depth of the second soil layer (d_2) (Liang et al., 1994; Wood et al., 1997). Previous studies indicated that b_{inf} and d_2 were the key parameters for model calibration (Wood et al., 1997, Demaria et al., 2007; Tang and Piechota, 2009).

A 15-yr period (1976–1990) and a separate 15-yr period (1990–2005) of the 30-yr record were utilized for the calibration and validation process. Both 15-yr periods encompassed a range of wet, dry, and normal years for testing the VIC model's performance. Two standard statistical techniques, the Nash–Sutcliffe coefficient (*E*; Nash and Sutcliffe, 1970) and the widely used coefficient of determination (*R*) value, were used to test and evaluate the accuracy of the model simulations. The finalized values of the calibration parameters were: infiltration parameter (b_{inf}) = 0.1; the maximum baseflow (D_{smax}) = 10 mm/d; the fraction of the lowest soil layer's maximum soil moisture wherein nonlinear baseflow occurs (D_s) = 0.05, fraction of maximum soil moisture (W_s) = 0.9 mm/d and the depth of the second soil layer (d_2) = 30 cm.

3.3. Calibration results

Fig. 2 shows the calibration results for three stations: the Salmon, Little Salmon, and White Bird stations. The variability of streamflow and CT were captured well. The *R* values (Fig. 2, right side) were fairly high with 0.85 at Salmon, 0.90 at White Bird, and 0.91 at Little Salmon for the entire study period. The *E* values were 0.75 at Salmon, 0.81 at White Bird, and 0.82 at Little Salmon. However, major discrepancies occurred between the modeled and observed streamflow during the peak months of May, June, and July. The VIC model underestimated the peak flow. Because few precipitation observation stations exist in the mountainous regions of the SRB, the forced precipitation data may not have accurately represented snow accumulation at high elevations. Most of the precipitation observation stations are located in valleys and towns rather than in the mountains, where more precipitation falls with snow.

The emphasis of this study was on examining the hydrological variances of different temperature change scenarios rather than on streamflow predictions, which avoided the effects of model bias. To this extent, we can assume that the simulated streamflow was adequately judged. Nonetheless, the high *R* and *E* values calculated during calibration suggest that the calibrated model can be applied to the further study of streamflow variation.

4. Analysis and results

This section examines the streamflow changes in the SRB at annual, seasonal, monthly, and daily scales under 30 different temperature increase scenarios.

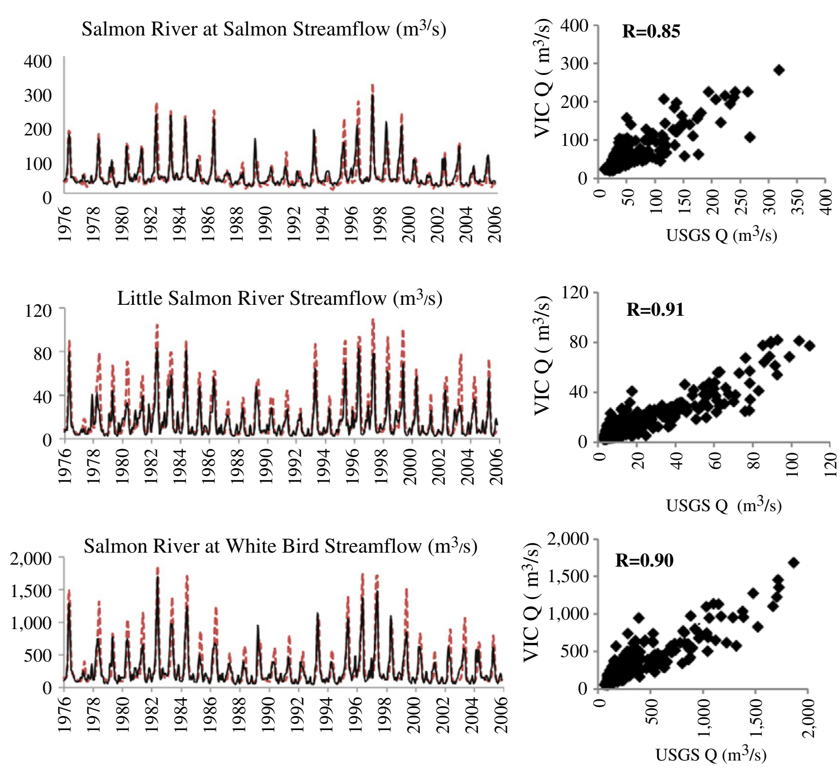

Fig. 2. VIC model calibration results in three stations.

4.1. Annual scale

4.1.1. Analysis

The objective of this section is to share the analysis of changes in annual streamflow patterns due to temperature increases in the SRB. In this study, it has been hypothesized that an increase in temperature results in a decrease of annual streamflow based on previous studies in the western United States (Christensen et al., 2004). In an attempt to prove this hypothesis, we compared the VIC-modeled streamflow for 30 scenarios developed independently by increasing 0.1 °C per step up to 3 °C. To quantify the streamflow's response to increased temperatures, mean annual streamflow changes were documented as magnitudes in addition to differences in percentages between the base condition (i.e., historical data over the 30-yr period, 1976–2005) and the temperature increases at eleven stations (Fig. 1). A moving, 5-yr window of average annual streamflows was also applied across the 30-yr sample to account for short-term fluctuations and highlight long-term variables.

4.1.2. Results

Figs. 3 and 4 illustrate the impact of temperature increases on the 30-yr annual mean streamflow at three gage stations in terms of absolute magnitude and relative percentage, respectively. Eight more stations were evaluated, although the results are not discussed in this study. The average annual streamflow decline generally range from 1% to 4%, with the largest decrease in White Bird station, when the temperature increased 1 °C. The average annual streamflow is expected to decrease in range from 2% to 6%, with the least decrease in Pahsimeroi River Station, when the temperature rose 2 °C. The average annual streamflow experienced a decrease from 3% to 8% when temperature increases 3 °C. The highest annual streamflow decrease is in the White Bird station. It should be noted that the temperature increases are associated with increased evaporation, which imply decrease runoff.

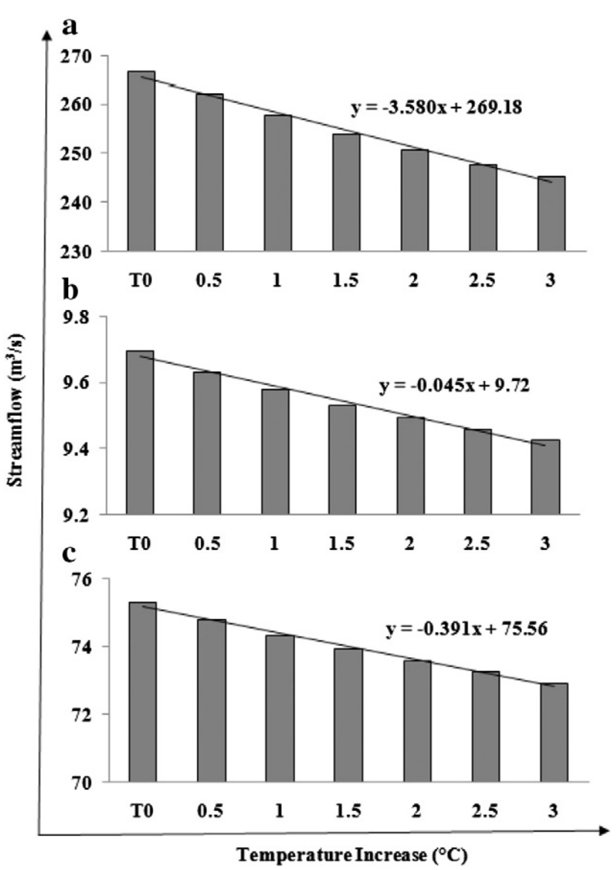

Fig. 3. Annually mean streamflow changes in magnitude scale under seven temperature change scenarios in three stations: a) White Bird, b) outlet of Pahsimeroi River, and c) outlet of Middle Fork. The seven scenarios are: T0 = historical temperature, Δt = 0.5 °C, 1 °C, 1.5 °C, 2 °C, 2.5 °C, and 3 °C. X means the base temperature (T0), and the temperature increases based on the base temperature. Y means the average annual streamflow changes in magnitude based on the base condition.

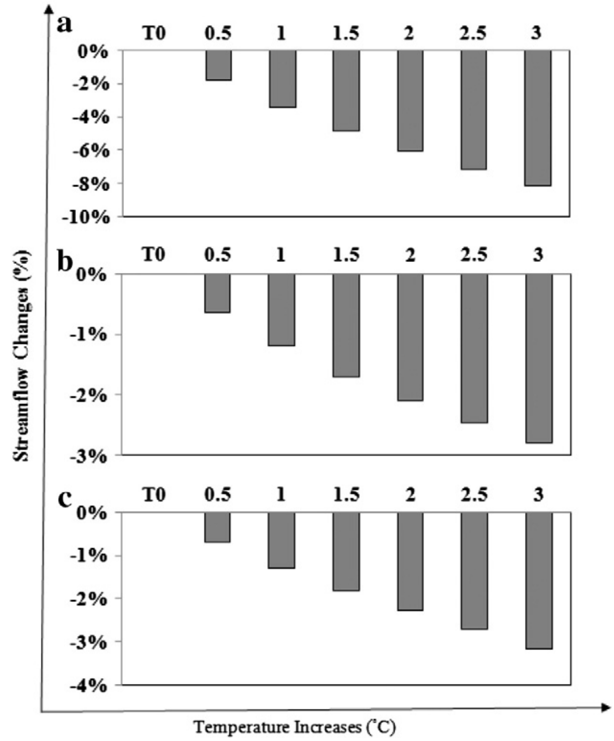

Fig. 4. Annually mean streamflow changes in percentage under seven temperature change scenarios in three stations: a) White Bird, b) outlet of Pahsimeroi River, and c) outlet of Middle Fork. The seven scenarios are: T0 = historical temperature, Δt = 0.5 °C, 1 °C, 1.5 °C, 2 °C, 2.5 °C, and 3 °C. X means the base temperature (T0), and the temperature increases based on the base temperature. Y means the average annual streamflow changes in percentage based on the base condition.

Overall, increases in temperature led to decreases in average annual streamflow at the three stations (Figs. 3 and 4). Decreases in the average annual streamflow were also recorded at the other eight observation stations. These decreases are relatively modest with a range of 1% to 8% change.

The relationship between temperature increases (x) and the mean annual streamflow (y) are displayed as equations in Fig. 3, allowing estimation of the mean annual streamflow under different scenarios of temperature increase. For instance, the mean annual streamflow would decrease 3.58 m^3/s, 0.045 m^3/s, and 0.391 m^3/s when temperatures increased 1 °C at the White Bird, Pahsimeroi River, and Middle Fork stations, respectively. These results suggest that mean annual streamflow would continue to decrease as temperatures continued to increase. A similar result was found for the Colorado River Basin in which an increase in temperature was proportional to decreases in both snowpack and total runoff (Christensen et al., 2004).

The impact of temperature on streamflow was also relative to the spatial distribution of the stations. For example, average annual streamflow increases in percentages were roughly equal for the Pahsimeroi River (2.8%) and Middle Fork (3.2%) stations (Fig. 4), with the Pahsimeroi River station exhibiting the weaker signal in terms of streamflow magnitude (1.0 m^3/s/2.4 m^3/s) when temperature increased 3 °C. The Middle Fork station is located in a warmer basin, whereas the Pahsimeroi River station is in a snow-dominated basin situated in the upper SRB with the coldest average temperature, running as low as −8 °C, among the five sub-basins (i.e., the Pahsimeroi

River, Lemhi, Middle Fork, South Fork, and Little Salmon basins). Even the highest temperature increases (3 °C) remained quite cold and well below the snow melting threshold of 0 °C at the Pahsimeroi River station. Smaller reductions in streamflow were also found for most stations located along the upper Salmon River, which were less susceptible to the temperature increases. The impacts of increased temperatures on streamflow at White Bird were significant in terms of magnitude and percentage scales. For example, the most reduction in streamflow occurred at White Bird when temperatures increased 3 °C, and the least reduction was seen at the Pahsimeroi River station when the temperature increased 1 °C.

The temperature increases also affected the moving 5-yr average streamflow. Fig. 5 shows the 5-yr mean streamflow changes when temperature increases at White Bird, Pahsimeroi River, and Middle Fork for three climate scenarios: the base condition, a temperature increase of 2 °C, and a temperature increase of 3 °C. Although there were a few differences in magnitude, the general patterns of the 5-yr moving average of streamflow were the same at all three stations.

There are four time periods over the 30 yr studied that warrant special attention (Fig. 5). The first is the low flow period from 1983 to 1991 in which the flow dipped as low as 67 m^3/s; 1987 and 1988 were especially low. According to Sheffield et al. (2009), one of the most severe droughts in United States history occurred from 1987 to 1990. Another low flow period took place from 2001 to 2004 during a nationwide drought (Piechota, et al., 2004). Conversely, a high streamflow period occurred between 1979 and 1982, especially from 1980 to 1982, with flows as high as 131 m^3/s. That may have been a direct result of flooding conditions caused by the 1982 El Niño. The years from 1993 to 1997 comprise another high flow period during the Great Flood of 1993. Significant rainfall combined with already saturated soil conditions was the cause of the severe flooding in the SRB and over the Midwest.

The black line in Fig. 5 represents the average streamflow at the base condition, and the dashed black line denotes a temperature increase of 3 °C. Note that the differences at the three stations all decreased in the same direction. Plots indicated an 8.0 m^3/s difference in average streamflow at White Bird, a 0.3 m^3/s streamflow difference in average streamflow at Pahsimeroi River, and a 2.5 m^3/s difference at Middle Fork.

4.2. Seasonal scale

4.2.1. Analysis

Seasonal changes in streamflow patterns strongly affect the timing of snowmelt and runoff, which suggests that studies of hydrological sensitivity to temperature increase should take seasonal hydrologic changes into account. As expected, the amplitude of streamflow had a strong seasonal signal in the SRB. At the seasonal scale, the hypothesis was that increases in temperature would cause a cyclical variation of streamflow in the Salmon River from increases in winter to decreases in summer. The four seasons were defined as winter (December–February), spring (March–May), summer (June–August), and autumn (September–November). To analyze seasonal streamflow trends, six stations along the main stem were selected along with five tributary stations. Regression equations for seasonal

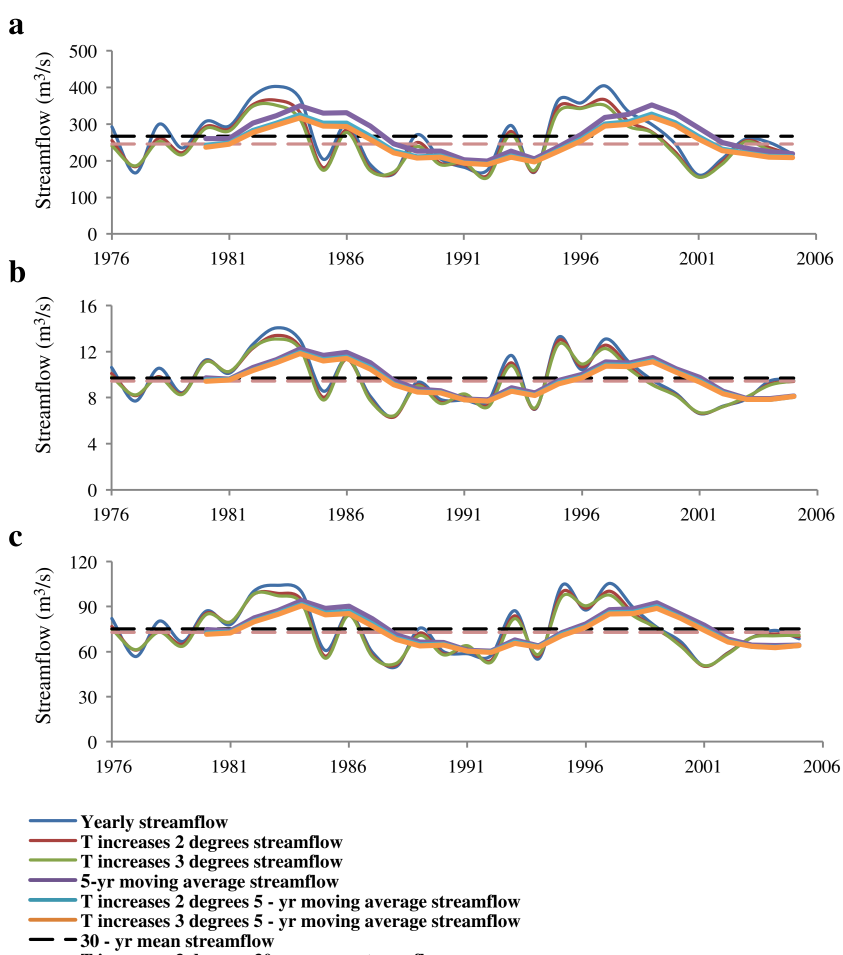

Fig. 5. Yearly streamflow and 5-yr moving average streamflow with temperature increase in three stations: a) White Bird, b) outlet of Pahsimeroi River, and c) outlet of Middle Fork.

streamflow changes and temperature increases are also evaluated in this section.

4.2.2. Results

The effects of seasonal temperature increases on streamflow coincided with decreases on the annual scale (Section 4.1); the impacts on the seasonal scale, however, were more varied.

Streamflow decreased in the spring and summer and increased during autumn and winter (Fig. 6). The increased temperature has two potential impacts on water resource. First, the increased temperatures are associated with increased evaporation, and decreased runoff. Second, the increased temperature lead to extreme precipitation, which is associated with increased runoff. Spring streamflow decreased at an average of 15 to 20% when temperatures increased 3 °C. The decreases in spring runoff translate to a significant drop in mean annual streamflow. Summer streamflow decreased an average of 10 to 40% when temperatures increased 3 °C. This would significantly impact the summer water resources system such as the urban water supply in the SRB and instream flow for fish and irrigation. Furthermore, the significant streamflow decrease in summer would increase summer drought risks. This provides significant information for farmers on the need to store extra water for summer droughts.

The biggest change occurred at the White Bird station, which experienced a 42% decrease in streamflow. Autumn saw a slight increase in streamflow (3–4%) at the three stations, and winter streamflow rose an average of 50% to 60% when temperatures increased 3 °C. The significant streamflow increase in winter would intensify winter flood risks, which is significant information needed for the design and operation of flood control infrastructure. However, the total annual runoff decreased (Fig. 3), indicating that wetter winters were not sufficient for offsetting dryer springs and summers. Unsurprisingly, the results that are not pictured from the other gage stations reflected the same trends as these three stations.

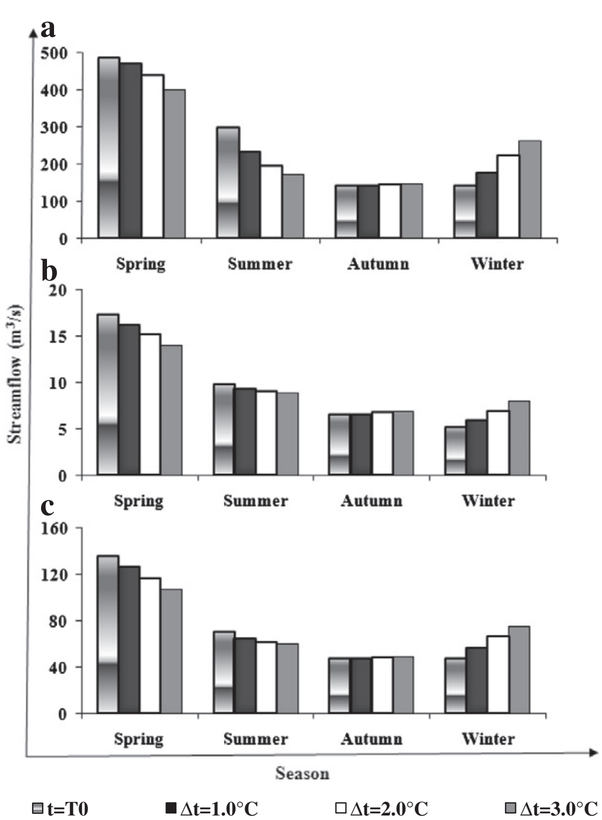

Fig. 6. Seasonal mean streamflow changes under four temperature scenarios in three stations: a) White Bird, b) outlet of Pahsimeroi River, and c) outlet of Middle Fork.

These results demonstrate that as temperatures increase in the SRB, more precipitation will fall as rain rather than snow in winter, increasing the winter streamflow and decreasing the summer streamflow. These changes will affect irrigation, water supply, flood control, and other water resource objectives.

At the base condition, streamflow was typically highest in spring, closely followed by summer. When temperatures increased, however, winter streamflow was greater than summer streamflow, indicating that changes in temperature alter the seasonal distribution of streamflow due to an earlier snowmelt. The earlier snowmelt resulted in increased streamflow in winter, decreased spring and summer streamflow, and possibly even prolonged summer droughts.

Fig. 7 further illustrates the relationship between increased temperatures and streamflow changes at the White Bird station for the first 6 mo (Fig. 7a, January–June), the last 6 mo (Fig. 7b, July–December), and the four seasons (Fig. 7c). There was a linear relationship between monthly streamflow and rising temperatures for every month except June and July, which exhibited logarithmic correlations. Positive linear relationships occurred for January, February, November, and December, and negative linear relationships were observed for April, May, August, September, and October.

The most similar phenomena were observed in August and September. For instance, when temperatures increased 1 °C, monthly streamflow increased 0.05 m^3/s and 0.04 m^3/s in August and September, respectively. When temperature increased, The maximum increased streamflow occurred in January, in which a 1 °C increase in temperature resulted in a 0.35 m^3/s increase in monthly streamflow. Conversely, the minimum increased streamflow was observed in September, when a 1 °C shift in temperature only resulted in a 0.04 m^3/s increase in monthly streamflow.

Streamflow changes in spring, autumn, and winter all exhibited linear relationships with the temperature increases, and the linear regression parameters range from 0.01 (autumn) to 0.03 (winter). Summer, with its logarithmic relationship, was the sole exception. Fig. 7c also confirms that autumnal flow was the least susceptible to rising temperatures, because the streamflow only increased 0.01 m^3/s when the temperature increased 1 °C. These results could be extrapolated to estimate monthly and seasonal streamflow given various scenarios of temporal change.

The key aspects of these changes in seasonal streamflow are considerations for water resource managers in managing winter flood and summer drought risk increases when temperatures increase.

4.3. Monthly scale

4.3.1. Analysis

Monthly streamflow changes are even more relevant in the SRB from a water management perspective. This study hypothesized that streamflow would increase from November to March, decrease from May to August, and vary in other months due to rising temperatures in the SRB. To test this hypothesis, monthly changes in streamflow from the six main-stem stations were calculated as percentages. To further explore the hypothesis, a similar analysis was performed for the five tributaries.

4.3.2. Results

As expected, the impact of rising temperatures was more significant at the monthly scale than at the seasonal and annual scales. Fig. 8 illustrates the average monthly streamflow variations throughout the 30 climate warming scenarios for the three gage stations along the main stem: Stanley, North Fork, and White Bird.

As shown in the plots for the Stanley station, the general patterns of monthly streamflow changes were fairly consistent among the 30 temperature increase scenarios. For instance, average monthly streamflow rose from late October to early April, with the greatest change occurring in February when the increase was over 150%.

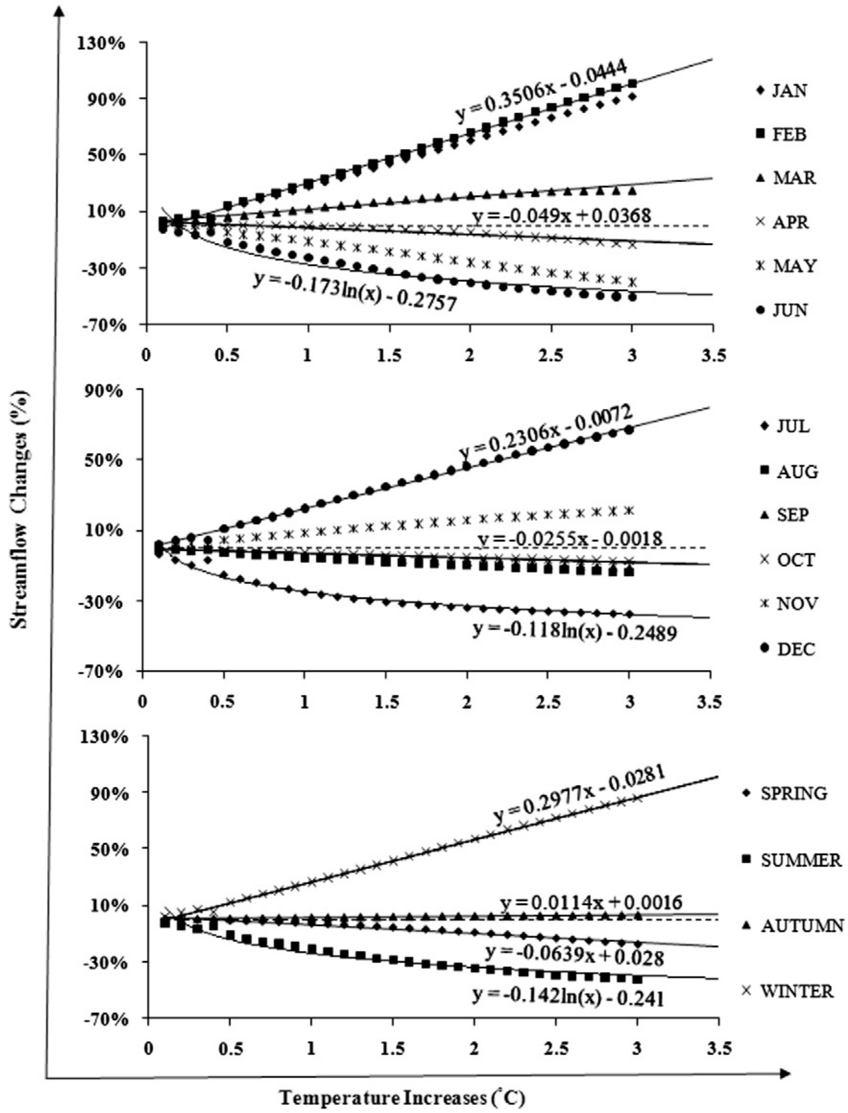

Fig. 7. Relationship between streamflow changes in percentage with temperature increase in White Bird for: a) from January to June, b) from July to December, and c) Four seasons.

Average monthly flow decreases were observed from late April to early October, with the largest drops recorded in May and June, which may result in a summer drought. Streamflows for July, August, and September did not appear to be as sensitive to rising temperatures. The month with the largest percentage increase in streamflow for the North Fork station was January, with more than a 60% increase. The largest increase was seen in February for the White Bird station, with a maximum change of over 100%. Streamflows decreased from March to October at North Fork, with the sharpest drop of nearly 30% occurring in May. The reduction started later at White Bird, and April to October saw the largest decrease, representing a greater than 50% change. These results illustrate the significant effect that increasing temperatures will have on the monthly water resources in the SRB.

Fig. 9 illustrates the monthly streamflow changes at the tributary observation stations. The pronounced monthly cycles recorded at the main-stem stations were reflected at the tributaries as well. According to the plot, in four stations, monthly streamflow appeared to peak in May, and minimum streamflow occurred in August. At the South Fork and Little Salmon stations, the largest decrease in streamflow occurred in June. For the North Fork station, it was May, and it was April for the Pahsimeroi River station. The month with the largest increase at all four stations was February.

Streamflow showed the strongest increases, up to 48%, from November to March at the North Fork and Little Salmon stations. From November to February at the Pahsimeroi River station, it increased up to 36%. From November to April at the South Fork station, it increased up to 90%. The increased streamflow during the cold, winter months may have been due to increased winter precipitation as rainfall rather than snow due to increased temperatures. Streamflows at all four stations were not as sensitive to rising temperatures from August through October. The most significant streamflow changes only occurred during of the first half of the year, from January to June.

Notably, the magnitude of the increasing streamflow during the months in which it increased at three separate stations was not strong enough to compensate for the decreasing streamflow during the months in which it lessened. Thus, the decrease in mean annual streamflow in relation to rising temperatures was unsurprising (Figs. 3 and 4).

4.4. Daily scale

4.4.1. Analysis

Snowmelts taking place earlier in the year, as caused by temperature increases, are a challenge to snow-dominated basins like the SRB. Although the exact day when the snowmelt pulse starts cannot be accurately determined, the CT was defined as the date when 50% of the

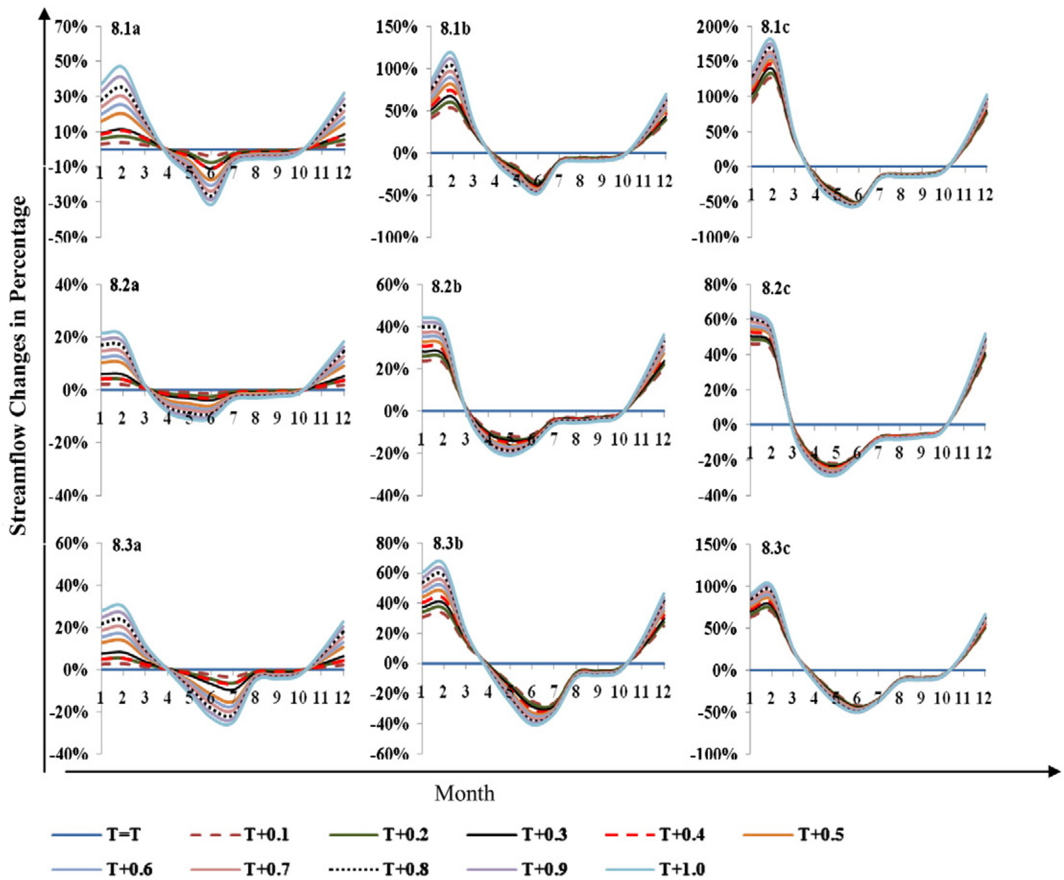

Fig. 8. Monthly streamflow changes (%) under 30 temperature scenarios in three stations, 8.1) Stanley; 8.2) North Fork, and 8.3) White Bird. 30 temperature change scenarios are: temperature increases 0.1 °C per step, a) Δt = 0–1.0 °C, b) Δt = 1.1–2.0 °C, c) Δt = 2.1–3.0 °C.

water-year flow had passed (Stewart et al., 2005). This definition provided a time-integrated perspective on the snowmelt and the overall streamflow distribution for each year (Stewart et al., 2005). This study hypothesized that the CT would occur earlier in the spring than in years past with the increase in temperatures (Stewart et al., 2005; Hidalgo et al., 2009).

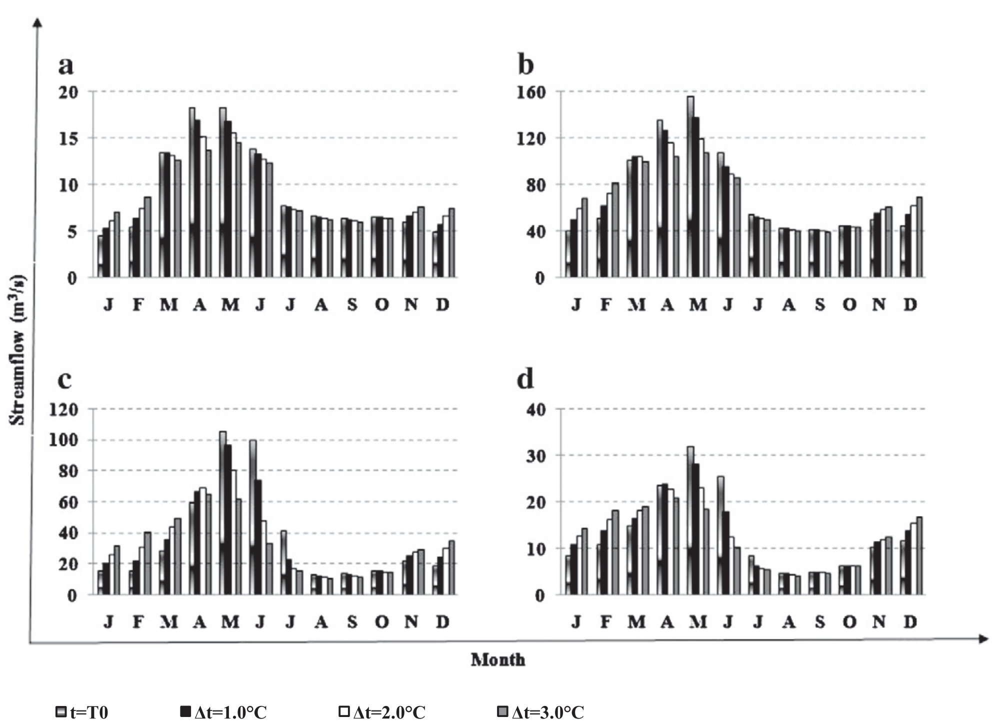

Fig. 9. Monthly average streamflow changes under four temperature scenarios in four tributaries outlets: a) Pahsimeroi, b) North Fork, c) South Fork, and d) Little Salmon.

First, the CTs were calculated for the 30 temperature increases scenarios at 11 gage stations during the study period (Fig. 1). Second, daily streamflow variations were evaluated for the 30 temperature increase scenarios using the Richards–Barker Index (R–B Index; Baker et al., 2004), which is an important component of the hydrologic regime. The R–B Index reflects the frequency and rapidity of changes in daily streamflow, which is the sum of the absolute value of day-to-day changes in daily streamflow volume (i.e., the path length), divided by the total annual flow (Eq. (1)). This calculation method measures the path length of flow oscillation. Longer paths relate to flashier streamflow, and shorter paths indicate more constant flows. Results from the R–B Index show the complete range of hydrological responses over an annual time step under increased temperatures. This is Eq. (1), where Q_t is the daily streamflow for Day t, and n is the number of days in a year:

$$R-B\ Index = \frac{\sum_{t=1}^{n}|Q_{t-1}-Q_t|}{\sum_{t}^{n}Q_t} \quad (1)$$

Values for the R–B Index could theoretically range from zero to 2. The R–B Index is equal to zero if the stream flow were absolutely constant. Its value increases as the path length and flashiness increase. Temperature increases may lead to increased or decreased flashiness. Our hypothesis stated that increased temperatures in the SRB would increase the R–B Index as well.

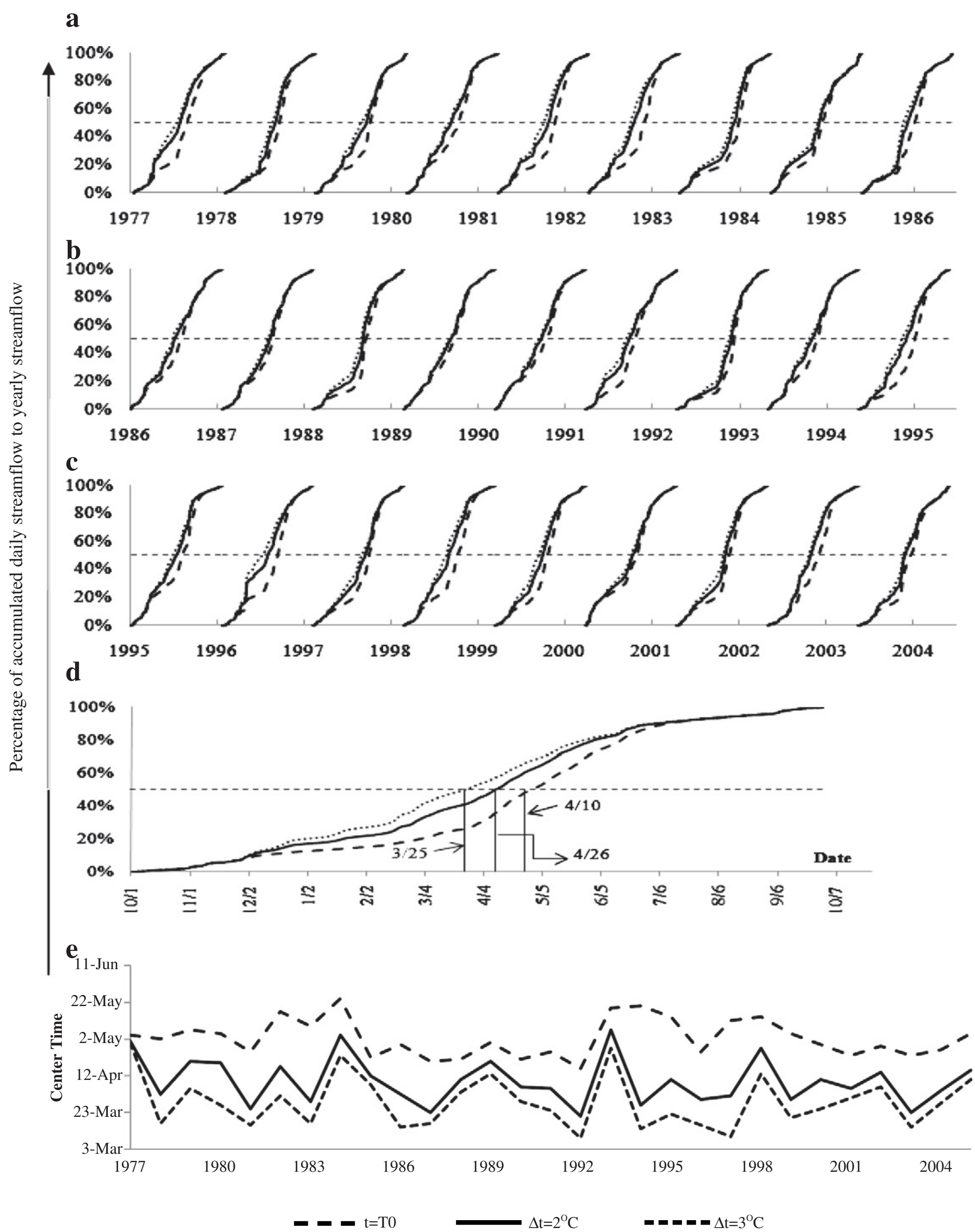

Fig. 10. CT curves under three climate scenarios of a) from 1977 to 2005, b) in year 2000, and c) CT date curve from 1977 to 2005.

4.4.2. Results

To illustrate how the CT changed during the 30-yr study period, Fig. 10a shows the CT curves under three scenarios: the base condition, a 2 °C increase in temperature, and a 3 °C increase in temperature. Fig. 10b illustrates the CT curve for the year 2000 at the White Bird station. Additionally, Table 1 and Fig. 10c show the exact CT dates for the study period.

As expected, widespread trends toward earlier CTs were observed. The most prevalent trends in CT dates included a 10 to 30 d shift wherein temperatures increased 2 °C and a 15 to 45 d shift wherein temperatures increased 3 °C. Earlier snowmelts representing a CT shift of more than 30 d were observed in 1981, 1982, 1983, 1995, 1997, 1999, and 2004. Temperatures increased 2 °C in each of those years. The earlier snowmelts resulted in a decrease in summer streamflow in addition to a prolonged drought period (Stewart et al., 2005). The drought from 1981 to 1983 was regarded as the worst drought of the 20th century in terms of overall impact, which is not surprising given the earlier CT dates. The drought period that began in 1991 reached its maximum intensity and coverage from 1994 to 1995.

Consequently, during the 2 yr in which temperatures increased by 3 °C, CTs shifted more than 45 d. Conversely, the smallest CT shifts, 12 and 22 d, occurred in 1993 when temperatures increased by 2 °C and 3 °C, respectively. The smaller shift in CTs may have been a contributing factor to the great flood of 1993. An important piece of accompanying evidence is the 30 yr of historical data for the White Bird that show that the CT typically began in late April or early May. Compared with the annual streamflow decreases (relatively modest with a range of 1% to 8% change (Section 4.1.2)), the CT changes are significant. That indicates that changes in climate primarily affected the timing rather than the amount of average annual streamflow.

Fig. 11 shows the R–B Indices at three stations—the Pahsimeroi River, Middle Fork, and White Bird—under different temperatures. The R–B flashiness index reflects the frequency and rapidity of short-term changes in streamflow, especially for the daily time step. Fluctuations in the R–B Index, which are expected from year to year under the same scenarios due to natural variations in weather, were observed at the three stations. Increases in the R–B Index due to higher peak flows accompanied increases in temperature (Fig. 11). The median R–B Indices increased 0.02 when temperatures increased 2 °C for all three stations. The median R–B Indices increased 0.03 when temperatures increased 3 °C for both White Birds and Pahsimeroi, with the exception of the Middle Fork, where R–B Index increased 0.04. Contrary to the historical records, stronger fluctuations were observed when temperatures increased, especially at the Pahsimeroi River station. The highest R–B Indices at all three stations were observed in 1993, which may due to the great Midwestern flood that caused over $20.1 billion in damages at that time.

The R–B Index is a useful tool for diagnosing the scale of the channel problem. A higher R–B Index causes channel erosion problems, resulting in changes to channel shape, width, depth, and slope, especially in a river such as the Salmon, which is steep and composed of non-cohesive materials. Changes in the Salmon River shape, in turn, will also have significant impacts on stream ecology.

Table 1
Center time in three scenarios: 1) temperature is the historical record (T_0); 2) temperature increases 2 °C, and 3) temperature increases 3 °C.

Year	Center time		
	$T=T_0$	$T=T_0+2.0$	$T=T_0+3.0$
1977	5/4/1977	5/1/1977	4/30/1977
1978	5/2/1978	4/2/1978	3/17/1978
1979	5/7/1979	4/20/1979	4/5/1979
1980	5/5/1980	4/19/1980	3/27/1980
1981	4/25/1981	3/25/1981	3/16/1981
1982	5/17/1982	4/17/1982	4/1/1982
1983	5/9/1983	3/29/1983	3/17/1983
1984	5/24/1984	5/4/1984	4/23/1984
1985	4/22/1985	4/12/1985	4/7/1985
1986	4/29/1986	4/2/1986	3/15/1986
1987	4/20/1987	3/23/1987	3/17/1987
1988	4/21/1988	4/10/1988	4/3/1988
1989	4/30/1989	4/20/1989	4/13/1989
1990	4/21/1990	4/6/1990	3/29/1990
1991	4/25/1991	4/5/1991	3/24/1991
1992	4/16/1992	3/21/1992	3/9/1992
1993	5/19/1993	5/7/1993	4/27/1993
1994	4/20/1994	3/27/1994	3/14/1994
1995	5/14/1995	4/10/1995	3/22/1995
1996	4/25/1996	3/30/1996	3/16/1996
1997	5/12/1997	4/1/1997	3/10/1997
1998	5/14/1998	4/27/1998	4/13/1998
1999	5/5/1999	3/30/1999	3/20/1999
2000	4/29/2000	4/10/2000	3/25/2000
2001	4/23/2001	4/5/2001	3/31/2001
2002	4/28/2002	4/14/2002	4/6/2002
2003	4/23/2003	3/23/2003	3/15/2003
2004	4/26/2004	4/4/2004	3/28/2004
2005	5/5/2005	4/15/2005	4/10/2005

5. Conclusions and discussions

Changes in temperatures will change water resources in the SRB and evaluating what type of changes are likely to occur is a major challenge faced by water resource planners. This study reaffirmed previous results and improves on previous studies with the following contributions: (1) the development of large-scale, long-term, and high-resolution streamflow data in the SRB; (2) a first analysis of streamflow changes at annual, seasonal, monthly, and daily time scales for temperature increases in the SRB; (3) the identification of the exact CT dates for the study period; and (4) a comprehensive analysis of the R–B Index for temperature changes in the SRB.

The analysis of the impacts of increased temperatures on streamflow at the annual scale gave interesting results. The streamflows consistently decreased with rising temperatures, although station-to-station variability was seen due to spatial distribution. For instance, mean annual streamflow declined an average of 2 to 6% when temperatures increased 2 °C, and they declined 3 to 8% when temperatures increased 3 °C. Decreases in streamflow were also observed during the 5-yr moving average. This analysis can be used for better management and development of the water resources in the SRB.

Seasonally, streamflow decreased in spring and summer, increased in winter, and remained mostly stable in autumn. When temperatures increased 3 °C, streamflow decreased an average of 15 to 20% in spring and 10 to 40% in summer. Given the significant decrease in summer streamflow in the SRB, additional water storage may be required to provide irrigation during late summer. However, given the significant increase in winter streamflow, flood control infrastructure may need to be built up for those months. It was also observed that linear relationships existed between streamflow changes and increased temperatures for autumn, winter, and spring. Summer exhibited logarithmic correlations.

Evaluating monthly results from increased temperatures was slightly more complex than assessing seasonal streamflow. The majority of the streamflow increases occurred from November to February, when temperatures decreased, and decreased from May to July, when temperatures increased. February displayed the most sensitivity to temperature changes. These results indicated a drier summer whenever temperatures increased. Fortunately, increased streamflow in the colder months helped compensate for the drier conditions found prior to the following Mays. These results indicated a linear relationship between monthly streamflow changes and increased temperatures for all months with the exceptions of June and July in which logarithmic relationships were observed.

The impacts of rising temperatures on the daily streamflow resulted in two new observations. First, earlier snowmelt flows throughout the SRB occurred during the study period. The average CT shifted 10 to 30 d earlier when temperatures increased 2 °C, and it shifted 15 to 45 d earlier when temperatures increased 3 °C. These shifts are

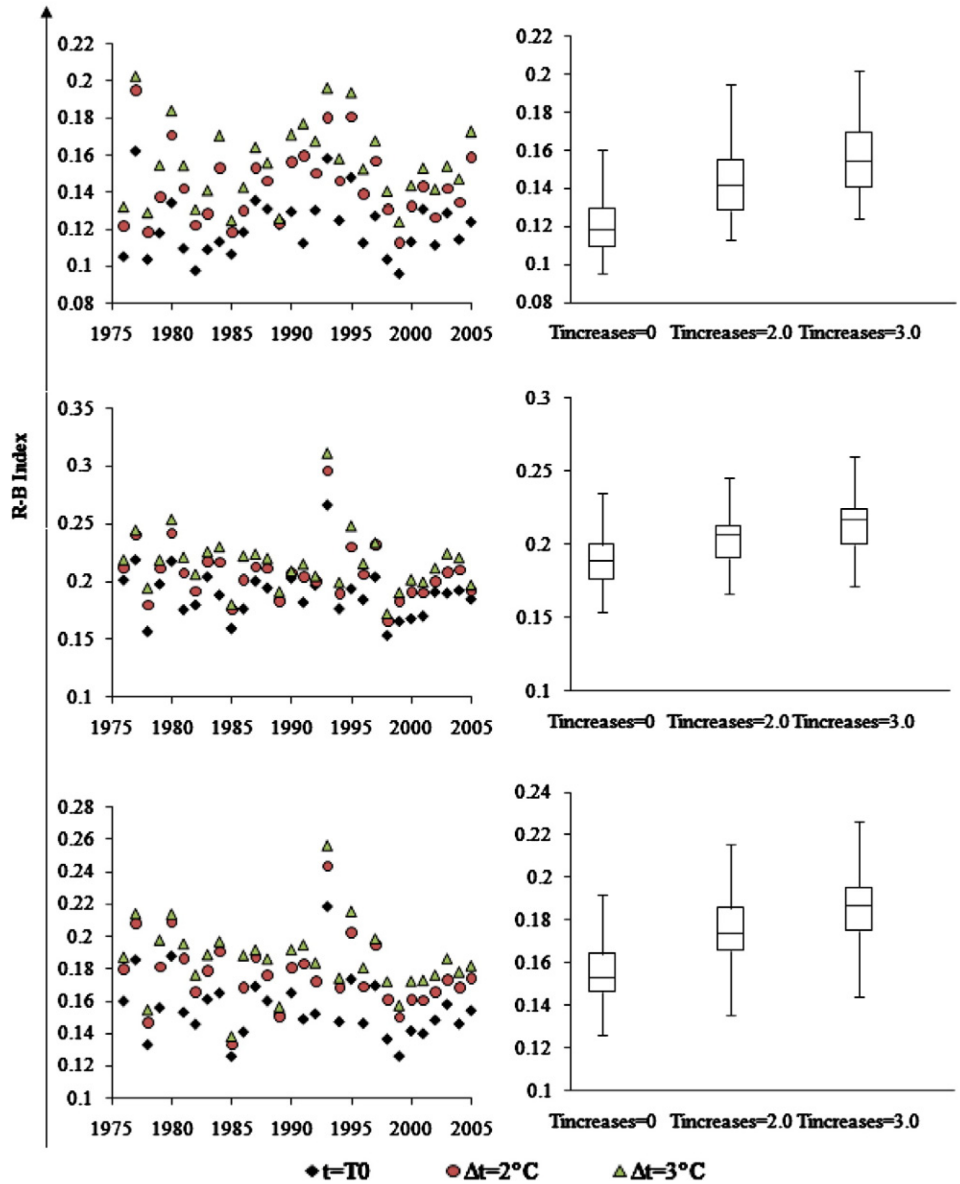

Fig. 11. R–B Index values and Box and Whiskers Plots of the R–B Index under three temperature scenarios in three stations: a) White Bird, b) outlet of Pahsimeroi River, and c) outlet of Middle Fork.

expected to become more pronounced as temperatures continue to rise. The amplitude of the shifts in CT increased during the drought years. Second, the R–B Index is well suited for determining gradual changes in daily streamflow and flow regimes associated with temperature changes. Increases in the R–B Index coincided with increases in temperature; ergo, rising temperatures led to higher R–B Indices when compared to the base conditions without temperature changes. These findings suggest that special attention should be paid to potential bank erosion problems due to increasing temperatures.

The overall results of this study indicate that temperature increases have significant impacts on the streamflow in the SRB. These varied impacts have many implications for water resources and management. Furthermore, funding for this study also provided significant information for scientists who work on the Salmon River. Although this particular study focused upon improving the understanding of the correlations between streamflow changes and rising temperatures in the SRB, it can nonetheless serve as the framework for other basins, like the Snake River Basin, because the SRB shares similar characteristics with other river basins in the northwestern United States.

Changes in streamflow are likely affected by both temperature and precipitation in the SRB, but it is not immediately clear from these results which of these factors hold the greater influence. This presents several questions for future consideration:

(1) How does precipitation in the SRB change influent streamflow in terms of both magnitude and timing?
(2) How do temperature increases and precipitation changes dominate streamflow changes in the SRB?
(3) What will the streamflow in the SRB be during the next century (2010–2099)?

These are the questions that will drive further explorations of this study in the future.

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References

Abdulla, F.A., Lettenmaier, D.P., Wood, E.F., Smith, J.A., 1996. Application of a macroscale hydrologic model to estimate the water balance of the Arkansas Red River basin. Journal of Geophysical Research-Atmospheres 101, 7449–7459.

Arnell, N.W., 2003. Relative effects of multi-decadal climatic variability and changes in the mean and variability of climate due to global warming: future streamflows in Britain. Journal of Hydrology 270, 195–213.

Baker, D.B., Richards, R.P., Loftus, T.T., Kramer, J.W., 2004. A new flashiness index: characteristics and applications to midwestern rivers and streams. Journal of the American Water Resources Association 40, 503–522.

Beyene, T., Lettenmaier, D.P., Kabat, P., 2010. Hydrologic impacts of climate change on the Nile River Basin: implications of the 2007 IPCC scenarios. Climatic Change 100 (3–4), 433–461.

Bronstert, A., Niehoff, D., Burger, G., 2002. Effects of climate and land-use change on storm runoff generation: present knowledge and modelling capabilities. Hydrological Processes 16, 509–529.

Bronstert, A., Kneis, D., Bogena, H., 2009. Interactions and feedbacks in hydrological change: relevance and possibilities of modelling. Hydrologie und Wasserbewirtschaftung 53, 289–304.

Chen, Y.N., Takeuchi, K., Xu, C.C., Chen, Y.P., Xu, Z.X., 2006. Regional Climate Change and Its Effects on River Runoff in the Tarim Basin, China. John Wiley & Sons Ltd., pp. 2207–2216.

Chiew, F.H.S., Teng, J., Vaze, J., Post, D.A., Perraud, J.M., Kirono, D.G.C., Viney, N.R., 2009. Estimating climate change impact on runoff across southeast Australia: method, results, and implications of the modeling method. Water Resources Research 45, 17.

Christensen, N.S., Lettenmaier, D.P., 2007. A multimodel ensemble approach to assessment of climate change impacts on the hydrology and water resources of the Colorado River Basin. Hydrology and Earth System Sciences 11, 1417–1434.

Christensen, N.S., Wood, A.W., Voisin, N., Lettenmaier, D.P., Palmer, R.N., 2004. Effects of climate change on the hydrology and water resources of the Colorado River Basin. Climatic Change 62 (1–3), 337–363.

Crozier, L.G., Zabel, R.W., Hamlett, A.F., 2008. Predicting differential effects of climate change at the population level with life-cycle models of spring Chinook salmon. Global Change Biology 14, 236–249.

Dai, A., Trenberth, K.E., 2002. Estimates of freshwater discharge from continents: latitudinal and seasonal variations. Journal of Hydrometeorology 3 (6), 660–687.

Demaria, E.M., Nijssen, B., Wagener, T., 2007. Monte Carlo sensitivity analysis of land surface parameters using the Variable Infiltration Capacity model. Journal of Geophysical Research-Atmospheres 112 (D11).

Easterling, D.R., Meehl, G.A., Parmesan, C., Changnon, S.A., Karl, T.R., Mearns, L.O., 2000. Climate extremes: observations, modeling, and impacts. Science 289 (5487), 2068–2074.

Hansen, M.C., Reed, B., 2000. A comparison of the IGBP DISCover and University of Maryland 1 km global land cover products. International Journal of Remote Sensing 21, 1365–1373.

Hidalgo, H.G., Das, T., Dettinger, M.D., Cayan, D.R., Pierce, D.W., Barnett, T.P., Bala, G., Mirin, A., Wood, A.W., Bonfils, C., Santer, B.D., Nozawa, T., 2009. Detection and attribution of streamflow timing changes to climate change in the western United States. Journal of Climate 22, 3838–3855.

Intergovernmental Panel on Climate Change (IPCC), 2008, 2009: climate change and water. Intergovernmental Panel on Climate Change Technical Paper VI. IPCC Secretariat, Geneva.

Jefferson, A., Nolin, A., Lewis, S., Tague, C., 2008. Hydrogeologic controls on streamflow sensitivity to climate variation. Hydrological Processes 22, 4371–4385.

Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J., Zhu, Y., Chelliah, M., Ebisuzaki, W., Higgins, W., Janowiak, J., Mo, K.C., Ropelewski, C., Wang, J., Leetmaa, A., Reynolds, R., Jenne, R., Joseph, D., 1996. The NCEP/NCAR 40-year reanalysis project. Bulletin of the American Meteorological Society 77, 437–471.

Kunkel, M.L., Pierce, J.L., 2010. Reconstructing snowmelt in Idaho's watershed using historic streamflow records. Climatic Change 98 (1–2), 155–176.

Lakshmi, V., Piechota, T., Narayan, U., Tang, C.L., 2004. Soil moisture as an indicator of weather extremes. Geophysical Research Letters 31.

Lettenmaier, D.P., Wood, E.F., Wallis, J.R., 1994. Hydro-climatological trends in the continental United States, 1948–88. Journal of Climate 7, 586–607.

Liang, X., Lettenmaier, D.P., Wood, E.F., Burges, S.J., 1994. A simple hydrologically based model of land-surface water and energy fluxes for general-circulation models. Journal of Geophysical Research-Atmospheres 99, 14415–14428.

Lohmann, D., Lettenmaier, D.P., Liang, X., Wood, E.F., Boone, A., Chang, S., Chen, F., Dai, Y.J., Desborough, C., Dickinson, R.E., Duan, Q.Y., Ek, M., Gusev, Y.M., Habets, F., Irannejad, P., Koster, R., Mitchell, K.E., Nasonova, O.N., Noilhan, J., Schaake, J., Schlosser, A., Shao, Y.P., Shmakin, A.B., Verseghy, D., Warrach, K., Wetzel, P., Xue, Y.K., Yang, Z.L., Zeng, Q.C., 1998. The Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) phase 2(c) Red-Arkansas River basin experiment: 3. Spatial and temporal analysis of water fluxes. Global and Planetary Change 19, 161–179.

Luce, C.H., Holden, Z.A., 2009. Declining annual streamflow distributions in the Pacific Northwest United States, 1948–2006. Geophysical Research Letters 36.

Ma, H., Yang, D., Tan, S.K., Gao, B., Hu, Q., 2010. Impact of climate variability and human activity on streamflow decrease in the Miyun Reservoir catchment. Journal of Hydrology 389, 317–324.

Mantua, N., Tohver, I., Hamlet, A., 2010. Climate change impacts on streamflow extremes and summertime stream temperature and their possible consequences for freshwater salmon habitat in Washington State. Climatic Change 102 (1–2), 187–223.

Maurer, E.P., O'Donnell, G.M., Lettenmaier, D.P., Roads, J.O., 2001. Evaluation of the land surface water budget in NCEP/NCAR and NCEP/DOE reanalyses using an off-line hydrologic model. Journal of Geophysical Research-Atmospheres 106, 17841–17862.

Maurer, E.P., Wood, A.W., Adam, J.C., Lettenmaier, D.P., Nijssen, B., 2002. A long-term hydrologically based dataset of land surface fluxes and states for the conterminous United States. Journal of Climate 15, 3237–3251.

McCabe, G.J., Wolock, D.M., 2002. A step increase in streamflow in the conterminous United States. Geophysical Research Letters 29.

Miller, W.P., Piechota, T.C., Gangopadhyay, S., Pruitt, T., 2011. Development of streamflow projections under changing climate conditions over Colorado River basin headwaters. Hydrology and Earth System Sciences 15 (7), 2145–2164.

Mote, P.W., Parson, E., Hamlet, A.F., Keeton, W.S., Lettenmaier, D., Mantua, N., Miles, E.L., Peterson, D., Peterson, D.L., Slaughter, R., Snover, A.K., 2003. Preparing for climatic change: the water, salmon, and forests of the Pacific Northwest. Climatic Change 61, 45–88.

Muttiah, R.S., Wurbs, R.A., 2002. Modeling the impacts of climate change on water supply reliabilities. Water International 27, 407–419.

Nash, J.E., Sutcliffe, J.V., 1970. River flow forecasting through conceptual models part I — a discussion of principles. Journal of Hydrology 10 (3), 282–290.

Nijssen, B., Schnur, R., Lettenmaier, D.P., 2001. Global retrospective estimation of soil moisture using the variable infiltration capacity land surface model, 1980–93. Journal of Climate 14, 1790–1808.

Piechota, T.C., Hidalgo, H., Timilsena, J., Tootle, G.A., 2004. Western U.S. drought: how bad is it? Eos 85 (32), 301–308.

Portmann, R.W., Solomon, S., Hegerl, G.C., 2009. Spatial and seasonal patterns in climate change, temperatures, and precipitation across the United States. Proceedings of the National Academy of Sciences of the United States of America 106 (18), 7324–7329.

Sheffield, J., Andreadis, K.M., Wood, E.F., Lettenmaier, D.P., 2009. Global and continental drought in the second half of the twentieth century: severity–area–duration analysis and temporal variability of large-scale events. Journal of Climate 22, 1962–1981.

Shepard, D.S., 1984. Computer mapping: the SYMAP interpolation algorithm. In: Willmott, G.L., Reidel, C.J. (Eds.), Spatial Statistics and Models Gaille, pp. 133–145.

Stewart, I.T., Cayan, D.R., Dettinger, M.D., 2005. Changes toward earlier streamflow timing across western North America. Journal of Climate 18, 1136–1155.

Tang, C.L., Piechota, T.C., 2009. Spatial and temporal soil moisture and drought variability in the Upper Colorado River Basin. Journal of Hydrology 379, 122–135.

Todini, E., 1996. The ARNO rainfall–runoff model. Journal of Hydrology 175, 339–382.

Walvoord, M.A., Striegl, R.G., 2007. Increased groundwater to stream discharge from permafrost thawing in the Yukon River basin: potential impacts on lateral export of carbon and nitrogen. Geophysical Research Letters 34, 6.

White, R.A., Miller, D.A., 1998. Processing soil survey information with Perl. Computers and Geosciences 24 (6), 591–593.

Wood, E.F., Lettenmaier, D., Liang, X., Nijssen, B., Wetzel, S.W., 1997. Hydrological modeling of continental-scale basins. Annual Review of Earth and Planetary Sciences 25, 279–294.