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Watershed Response to Climate Change and Fire-Burns
in the Upper Umatilla River Basin
Using the Precipitation Runoff Modeling System

by

Kimberly Crystal Yazzie

A thesis submitted in partial fulfillment of the
requirements for the degree of

Master of Science
in
Environmental Science and Management

Thesis Committee:
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Portland State University
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Abstract

This study provides an analysis of watershed response to climate change and forest fire impacts, to better understand the hydrologic budget and inform water management decisions for present and future needs. The study site is 2,365 km², located in the upper Umatilla River Basin (URB) in northeastern Oregon. The Precipitation Runoff Modeling System, a distributed-parameter, physical-process watershed model, was used in this study. Model calibration yielded a Nash Sutcliffe Model Efficiency of 0.73 for both calibration (1995-2010) and validation (2010-2014) of daily streamflow. Ten Global Climate Models using Coupled Model Intercomparison Project Phase 5 experiments with Representative Concentration Pathways 4.5 and 8.5 (RCP), were used to observe hydrologic regime shifts in the 2020s, 2050s, and 2080s. Mean center timing of flow occurs earlier in the year in both pre- and post-fire conditions, where there are increased winter flows and decreased summer flows throughout the 21st century. Change in temperature and percent change in precipitation is more variable in the summer than winter increasing over time, with a slight decrease in winter precipitation in the 2080s in RCP 8.5. Temperature increases 1.6°C in RCP 4.5 and 3.3°C in RCP 8.5 by the end of the 21st century. The ratio of Snow Water Equivalent to Precipitation decreases 96% in the 2080s in RCP 8.5 before forest cover reduction, and decreases 90-99% after forest cover reduction. Potential basin recharge and the base-flow index are both sustained throughout the 21st century with slight declines before forest cover reduction, with an increase in basin recharge and increase in base-flows in the 2080s after fire-burns. However, the

simulated sustained base-flows and area-weighted basin recharge in this study, do not take into account the complex geologic structure of the Columbia River Basalt Group (CRBG). A more robust characterization and simulation of URB aquifer recharge would involve coupling the PRMS model with a groundwater model in a future study. Although groundwater recharge in the CRBG in the URB is not well understood, the long-term decline of groundwater storage presents a serious environmental challenge for the Confederated Tribes of the Umatilla Indian Reservation and communities in the URB.

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1. Introduction

Anthropogenic influences on climate change have shifted the spatial and temporal distribution of water resources worldwide, requiring changes in the management of surface- and ground-water resources (Oki and Kanae, 2006; Kundzewicz et al. 2007; Praskievicz & Chang, 2011). Quantifying recharge and streamflow response to climate change is an early step to developing long-term water resource management plans to increase understanding of the global energy balance in a hydrologic regime to improve adaptive capacity (Qi et al., 2009; Waibel et al., 2013). Identifying trends in basin runoff is important due to the strong effects on water and energy demands (Jung & Chang, 2011) that may have important ecological implications.

A marked shift in global mean surface temperature in the 20th century has been widely cited as an indicator of climate change and its direct relationship to changes in the global energy budget (Hamlet & Lettenmaier, 2007; Abatzoglou, 2014). Climate change impacts in the Pacific Northwest include the shifts in the magnitude and timing of runoff (Chang and Jung, 2010; Elsner et al., 2010, Hamlet et al., 2010; Surfleet et al. 2012), reduced proportion of precipitation falling as snow in sub-montane regions (Knowles et al., 2006; Abatzoglou 2011), decreases in snow water equivalent (Mote et al., 2005; Mastin et al., 2011), and an increase in frequency and intensity of floods and droughts (Mote et

al., 2003; Jung and Chang, 2011b; Surfleet et al. 2012). Further, changes in snowpack have affected the timing of runoff (Stewart et al., 2005; Hamlet et al. 2006; Hamlet & Lettenmaier, 2007; Barnett et al. 2008; Vano et al. 2015) and soil moisture recharge (Hamlet & Lettenmaier, 2007), increasing cool season flows and decreasing warm season flows (Hamlet, 2010).

The earth's water and energy systems, including solar radiation and temperature, can be characterized by a watershed model to guide watershed assessments and natural resource management in a changing climate. The application of watershed models can integrate biosystems, geochemistry, atmospheric sciences and coastal processes (Singh & Frevert, 2006) to analyze trends in floods and droughts (Christensen et al. 2004; Park & Markus, 2014), streamflow and water quality (Raines et al. 2002; Hutchinson et al., 2013), agriculture and sediment loading (Chang et al., 2001; Lambrechts et al., 2014), and land use change (Choi 2008; Praskievicz & Chang, 2011).

Recharge rates vary widely in space and time, and are difficult to measure directly (Battin et al. 2007; Healy 2010; Chang and Jung, 2010; Waibel et al. 2013). At the regional scale, recharge has been estimated in the Columbia River Basin (Vaccaro 1986; Bauer & Vaccaro, 1987; Hansen et al. 1994; Burns et al. 2012), while no studies have been completed at the local scale in the Umatilla River Basin (URB) nor have the effects of fire on the hydrologic regime been analyzed in the URB by a surface-model, making this study unique.

The Precipitation Runoff Modeling System (PRMS) was chosen due to its application to evaluate various combinations of climate data and land use on flow regimes; use for basin-scale analysis (Cherkauer, 2004; Surfleet et al., 2012); offers a rigorous simulation of the subsurface system; ability to quantify groundwater recharge (Hunt et al., 2001; Steuer & Hunt, 2001; Cherkauer, 2004; Vaccaro & Olsen, 2007), evaluate interdecadal variability of climate change impacts (Chang & Jung, 2010; Jung & Chang, 2011; Markstrom et al., 2012), and evaluate land cover change (Qi et al., 2009; Vynee et al., 2010; Viger et al., 2011).

Three research questions will be explored to characterize the hydrologic regime in the URB:

- (1) How does the hydrologic regime change seasonally and annually, i.e. water budget variables, in response to climate change in the 2020s, 2050s and 2080s in comparison to historical conditions?
- (2) What are the effects of land cover change after fire-burns on basin runoff and other components of the water budget in the 2050s, and 2080s in comparison to historical conditions?
- (3) Which water budget components (e.g., seasonal runoff, snow water equivalent) are sensitive to changes in climate and could potentially be considered in water resource management for climate adaptation planning?

The Columbia River Basin has been researched extensively (Drost et al., 1986; Vaccaro 1986; Tolan et al., 1987; Bauer & Vaccaro, 1990; Chang et al., 2013), including the lower URB due to Critical Groundwater Areas and Ground Water Limited Areas with long-term decline in groundwater (Sceva, 1966; Oberlander & Miller, 1981; Burns et al., 2012). The upper URB in particular, is largely understudied. The aquifer system covers 113,959 km², including southeastern Washington, northeastern Oregon, and western Idaho. The United States Geological Survey (USGS) in a study of the Columbia Plateau Regional Aquifer System (CPRAS) found significant groundwater-level declines (Burns et al., 2012) and in the upper URB, declines have been estimated to be 107 cm/yr (Wallulis, 1995). Groundwater recharge in the URB has been estimated to range from 10,000 to ~64,000 ac-ft/yr (OWRD 1988; Ely, 2001). Previous studies estimated recharge for predevelopment conditions in the region to be 6.90 cm/yr, and in the 1980s, 10.80 cm/yr, an increase due to irrigation (Hansen et al., 1994). Studies have found significant recharge to occur with direct infiltration of interflow zones in the Columbia River Basalt Group (CRBG) at or near the surface or when directly connected to surface water (Newcomb, 1959, 1969; Davies-Smith et al., 1988; US DOE, 1988; Lindberg, 1989; Hansen et al., 1994; Tolan et al., 2009).

As for fire-burns, studies indicate that after a wildfire watershed systems often have increased runoff due to decreased canopy interception, reduced vegetation cover, and changed chemical and physical properties of soil (Moody & Martin, 2001; Viera et al, 2015). The infiltration rate after a wildfire has been

observed to decrease two to seven-fold (Cerdeira, 1998; Martin & Moody, 2001; Moody & Martin, 2001) and cause erosion from overland flow (Robichaud, 2005). Changes in peak discharges are more apparent than changes in annual runoff (Moody & Martin, 2001), where in some places, peak runoff increases by two orders of magnitude (Rulli et al, 2006, 2013; Terranova et al, 2009). Analysis of the effects of forest cover reduction can provide information on post-fire rehabilitation treatments on hill slopes to address increased runoff, decreased infiltration rates, and erosion (Robichaud, 2005), and to predict the magnitude of floods after a fire where the frequency and magnitude of peak discharge events are affected (Moody & Martin, 2001).

2. Study Area

2.1.1 Problem Statement

The upper Umatilla River Basin (URB) is largely understudied with a lack of understanding the surface and groundwater interactions and the effects of climate change on the hydrologic regime. Groundwater has been the primary supply of water throughout the 20th century for domestic, municipal, and agricultural needs with over 1,100-1,200 wells in the upper URB, and 700-800 wells on the Umatilla Indian Reservation (UIR) (Ely, 2001). In the 1920s, the Umatilla Reclamation Project blocked the return of anadromous fish resulting in a steep decline of salmon return. There are four Oregon Water Resources Department Groundwater Restricted Areas in the lower URB with long-term groundwater declines (Burns et al., 2012; OWRD, 2012). In 1988, the Umatilla Basin Project Act resulted in a bucket-for-bucket exchange of Umatilla River water for Columbia River water, which improved flows to restore salmonid and steelhead populations (The Umatilla Local Advisory Comm., 2012). Burns et al. (2012) indicate water levels have declined 30-91 m since 1970 in some of the deeper CRBG aquifers where the physical characteristics of basalt, depositional environment, folding and faulting impede groundwater flow.

Understanding groundwater and surface-water interactions is important to sustainable water resources management in the study area and region. Disproportionate impacts of climate change in tribal communities requires climate adaptation planning where some tribes are developing adaptation plans and few have implemented climate change into forest management plans (IFMAT, 2013).

The cultural value of water cannot be understated, making it more important to protect natural resources vital to the Confederated Tribes of the Umatilla Indian Reservation (CTUIR). The tribes ceded more than 25,899 km² to the United States in the Treaty of 1855, reserving their rights to fish, hunt and gather traditional foods on ceded lands (CUJ, 2012). CTUIR implemented the protection of first foods, including water, salmon, deer, cous, and huckleberry into natural resource management as a form of self-determination for environmental equity and tribal resilience, (Ely, 2001; Jones et al, 2008). First Foods are the “minimum ecological products necessary to sustain CTUIR culture” (Jones et al., 2008). This study can help support protection of cultural and ecosystem services in the study area by providing runoff trends throughout the 21st century for climate adaptation planning.

2.1.2. Geography

The study site is on the Columbia Plateau in the Yakima Fold Belt in the URB in northeastern Oregon (Burns et al., 2012). It is home to four informal physiographic subprovinces, each with its unique groundwater characteristics, including the Yakima Fold Belt, Blue Mountains, Palouse Slope subprovinces, and the Clearwater Embayment (Burns et al., 2012). The study site is 2,365 km² in area, bounded on the east by the Blue Mountains and Umatilla National Forest within the Upper Grand Ronde Subbasin, on the South by the North Fork John Day Watershed Subbasin, on the west by the Willow River Watershed subbasin, on the North by the Middle-Columbia-Lake Wallula subbasin and to the

northeast, the Walla Walla subbasin (Fig. 1). A significant portion of the UIR, 647 km², is within the study site boundaries (Fig. 1).

Originating in the Blue Mountains, the Umatilla River is a gravel-bed channel system with a multi-channel pattern (Hughes, 2008). It is largely groundwater fed, according to baseflow estimates based on a seepage run that determined 70-80% of groundwater contributions to surface water, providing flow in the summer months (Ely et al, 2001, 2012). The upper basin is approximately 14% in drainage area but supplies 40-50% of the average flow to the Umatilla River (US Forest Service, 2001; Forest Management Plan, 2010). The Umatilla River is a 145 km reach that enters the Columbia River, which further extends 2,000 km before it exits into the Pacific Ocean. The Umatilla River has eight major tributaries, including the North and South Forks, and Meacham Creek, which are in the study area. In the mid-basin, Wildhorse, Tutuilla, McKay, and Birch creeks join the main stem, and in the lower basin, Butter Creek joins the Umatilla River (CTUIR, 2010).

The basin is mostly semiarid, located east of the Cascades in the rain shadow. It receives 12.7 cm in annual precipitation in the lowlands and 127 cm in the highlands in the Blue Mountains (Ely, 2001). The mean annual precipitation for the Columbia Plateau is 43 cm (Kahle et al, 2011; Burns et al, 2012). Elevation in the URB ranges from 82-1,676 meters in the lowlands to the highlands (Ely, 2001). The study site encompasses 55.1% coniferous tree cover, 0.1% deciduous and mixed tree cover, 21.7% shrub cover, 21.8% grass, and

1.3% bare soil (Homer et al, 2015). The most common types of forest type are grand fir and Douglas fir, followed by ponderosa pine and lodgepole pine, at 26%, 25%, 17% and 6% cover, respectively (Christensen et al, 2007). U.S. Forest Service owns 13% of the land in the subbasin, while CTUIR manages 12%, urban areas comprise 6.0% and the rest of the land is shared by agriculture and rangeland uses (Vynee et al, 2010). Contrary to coniferous tree cover in the Blue Mountains, mid-elevation areas include grasslands and rolling hills, and desert vegetation downstream (Fig. 2). Annual water use on the UIR in 2005 includes 52% for irrigation, 30% for municipal needs, 18% for domestic needs, and 0% for pollution abatement (Technical Report, CTUIR, 2007).

Figure 1: The study site is 2,365 km² in the upper Umatilla River Basin in northeastern Oregon.

Figure 2: Photos of the URB upstream in the Blue Mountains (L), hilly grasslands at mid-elevation (M), and downstream near the City of Pendleton (R).

2.1.3. Geology

The geologic structure of the basin has an effect on groundwater flow and availability. The URB includes two geologic features, the Deschutes-Umatilla Plateau and the Blue Mountains. The study site is underlain by the Miocene CRBG that erupted 17-6 million years ago. It spans eastern Washington, northeastern Oregon and west central Idaho (Hooper, 1982; Tolan, 2009). From oldest to youngest in geologic age, the Grande Ronde basalt, Wanapum Basalt, and Saddle Mountain Basalt, are formations within the CRBG, each with unique physical, chemical, and paleomagnetic properties (Reidel et al, 1989b). The Grande Ronde Basalt and Wanapum Basalt formations are the primary CRBG formations on the UIR. The Grand Ronde Basalt is thicker and is found to have more permeable interflow zones and has a greater extent than other basalt formations (CTUIR, 2007; Burns et al, 2012).

2.1.4. Hydrogeology

CRBG aquifers are an important regional aquifer system providing the primary water supply for the URB. They are generally classified to be semi-confined to confined, with water bearing units and transmitting zones occurring in the Grande Ronde, Wanapum, and Saddle Mountains basalt groups, the interstratified Ellensburg Formation, sedimentary interbeds, interflow zones and in basalt flow tops (Gephart et al, 1979; USDOE, 1988; Hansen et al, 1994; Packard et al, 1996; Sabol & Downey, 1997; Bauer & Hansen, 2000; Tolan, 2009). Reidel and Hooper (1989) provide a comprehensive summary on the Columbia River flood province. Characterization of the CRBG aquifer system (Drost & Whiteman, 1986; Vaccaro, 1999; Tolan, 2009), groundwater recharge estimates (Bauer & Vaccaro, 1990), and groundwater behavior (Davies-Smith et al, 1988) offer a window into the complexity of the CRBGs not represented by PRMS, but when paired with the Modular Groundwater Flow Model (MODFLOW), surface and groundwater interactions can be quantified in a coupled Groundwater and Surface-Water Flow Model (GSFLOW) (Markstrom, 2008).

3. Data and Methods

3.1.1 Historic Climate Data

Climate data consisting of precipitation, minimum and maximum temperature, and solar radiation are the primary inputs in a PRMS simulation (Leavesley et al., 1983; Markstrom et al., 2008; Fig. 4). For this study, daily time-series of precipitation, minimum and maximum temperature were obtained for model input using the USGS' Geo Data Portal (GDP). Solar radiation data was not obtained, in which case PRMS internally estimates daily shortwave solar radiation (Markstrom et al., 2015). Discretized HRUs were uploaded to the GDP as a shapefile to obtain a daily time series of precipitation and minimum and maximum temperature (Table 1).

University of Idaho Gridded Surface Meteorological data (METDATA) for the continental U.S. were used for climate forcings at 4-km (1/24-degree) resolution for model calibration (1995-2010). High resolution gridded dataset was derived from observations and regional reanalysis using a hybrid method, combining spatially rich data from the Parameter-elevation Regressions on Independent Slopes Model (PRISM) and temporally rich data from the North America Land Data Assimilation System Phase 2 (NLDAS-2), (Mitchel et al., 2004; Daly et al., 2008; Abatzoglou, 2013).

3.2.1 Future Climate Data

Downscaled future climate data with a resolution of 4-km (1/24-degree) was obtained for the 1980s (1970-1999), 2020s (2010-2039), 2050s (2040-

2069), and 2080s (2070-2099). Future climate data were derived using the Multivariate Adaptive Constructed Analogs statistical downscaling method (MACA), a non-interpolated-based approach (Abatzoglou & Brown, 2012). Data from the Coupled Model Inter-Comparison Project 5 (CMIP5) use Representative Concentration Pathways (RCPs), instead of emission scenarios that define concentrations of greenhouse gases and aerosols. RCP4.5 and RCP8.5 scenarios were used in this analysis, where RCP4.5 is a medium stabilization scenario where an additional 4.5 W/m^2 of radiative forcing energy is trapped in the atmosphere by year 2100 (Van Vuuren et al., 2011). RCP8.5 has a very high baseline emission scenario at 90th percentile, where an additional 8.5 W/m^2 is trapped by 2100, where no climate action is executed (Van Vuuren et al., 2011).

Ten Global Climate Models (GCMs) were used in this study (Table 3), based on a model evaluation in the Pacific Northwest, that used eighteen performance metrics to rank model performance (Rupp et al., 2013). The GCMs chosen in this study had a low normalized error score compared to other models and were chosen based on availability (Rupp et al., 2013). These models take into account annual and decadal climate variability associated with the El Niño Southern Oscillation and the Pacific Decadal Oscillation.

Model performance for each GCM was compared to observations in the baseline period where temperature and precipitation predictions were analyzed for any discrepancies. There are minor differences in precipitation projections as differences are minimal ($<.02$), with greater differences in temperature ($<2.64^\circ\text{C}$)

but not large enough to exclude any GCMs (See Appendix C). Further, bias corrections were not made due to taking the ensemble mean of 10 GCMs, a consensus of climactic change, and because studies have found little to no difference in selecting or weighting GCM output (Mote et al., 2011).

Table 1: Datasets, models and tools used for analysis, including data derived.

Data	Resolution	Source
Historic Climate Data	4-km	Abatzaglou (2012)
Future Climate Data	4-km	Abatzaglou (2012)
Streamflow - US Geological Survey Stream Gage 14020850		USGS (2013)
Soils - NRCS State Soils Geographic	30-m	STATSGO (2013)
Land Use & Land Cover – Nat'l Land Cover Data	30-m	USGS (2013)
DEM - National Elevation Dataset	30-m	USGS (2013)
Point data and acres burned in the URB		USFS, Umatilla Natl. Forest
Models & Tools	Version	
Precipitation Runoff Modeling System (PRMS)	3.0.5	USGS (2013)
Geo Data Portal (GDP)		USGS (2013)
Let Us Calibrate (LUCA)		USGS (2013)
Web-based Hydrograph Analysis Tool		Purdue University (2015)

3.3 Methods

3.3.1 PRMS

The Precipitation Runoff Modeling System, version 3.0.5, was used in this study. It is a distributed-parameter, physical-process based watershed model developed by the USGS (Leavesley et al., 1983; Marskstrom et al., 2008, 2015). PRMS workflow includes preparation of input files including historical data, and

basin characteristics for each HRU, followed by calibration and validation, then running future climate data under different land cover scenarios (Fig. 4). PRMS estimates water-balance relations, stream flow regimes, soil-water relations and groundwater recharge (Markstrom et al., 2008). It simulates hydrologic processes for evaluation of the distribution of water among runoff, evapotranspiration, and infiltration (Cherkauer, 2004; Markstrom et al., 2015). Precipitation, air temperature, and solar radiation are the primary inputs into PRMS to compute evaporation, transpiration, sublimation, snowmelt, surface runoff, and infiltration (Fig. 3). Water that surpasses field capacity is routed to the subsurface reservoir or groundwater reservoirs. From these two reservoirs, water becomes interflow or groundwater discharge to a stream or lake, if it does not enter the ground-water sink where recharge occurs (Fig. 3; Markstrom, 2008, 2015). However, in this study the ground-water sink option was not used.

Figure 3: Schematic diagram of a watershed and its meteorological inputs simulated by PRMS (Markstrom et al., 2008)

Figure 4: Workflow of data input, PRMS calibration, and data output.

3.3.2 Indicators for detecting climate change impact

Water and energy balance variables will be evaluated under two different climate scenarios with low and high radiative forcings, RCP4.5 and RCP8.5, and comparing these scenarios with historical conditions in a multi-model ensemble approach. The following climate change impact detection indices were used: (1) Change in temperature and percent change in precipitation. (2) Ensemble mean changes in annual and seasonal runoff. (3) Center timing of streamflow (CT), indicating what date half of the water for that water year has passed. (4) The ratio

of snow water equivalent to precipitation (SWE/P). (5) Potential recharge, to analyze ground and surface water interactions. (6) The baseflow index (BFI), to indicate groundwater flow contributions to surface water flows. (7) Lastly, the Coefficient of Variation (COV) was calculated for seasonal flows to understand the variability and uncertainty in GCM projections within the winter and summer seasons. Four of the seven indices were used to analyze watershed response to land cover change, including CT, Seasonal flows, SWE/P, and basin recharge.

Climate change detection indices were compared between the means of historical and future time periods and deemed statistically significant at the 95% confidence interval. Parametric analysis was completed with one-way analysis of variance (ANOVA), and non-parametric testing was completed with the Kruskal-Wallis test. After detecting an overall significant difference, multiple comparisons were analyzed with Tukey's honest significance test or Kruskal Wallis' multiple comparisons test.

3.3.3 Generation of Hydrologic Response Units (HRU)

The discretization of HRUs is based on hydrologic and physical characteristics such as land use, vegetation, soil type, elevation, drainage boundaries, geology, and more (Markstrom et al., 2015). Partitioning the watershed to generate HRUs in the Umatilla River Subbasin was completed using watershed boundaries, soils, and land cover to derive a total of 107 HRUs (Table 2, Fig. 5). Each HRU unit is homogenous with unique physical and hydrologic facets, yielding a homogenous hydrologic response to climatic inputs. For each HRU, a daily water and energy balance is computed and the sum of the

responses of all HRUs, weighted on a unit-area basis, calculates the daily system response of the basin (Leavesley et al., 1983, Markstrom et al., 2015).

The number and size of HRUs is left to the discretion of the modeler, although a higher number of HRUs may result in diminished returns as the study site is broken down into smaller units while the real system is unchanged, where the sum and average of water and energy balances for each HRU will have no significant change. Partitioning HRUs smaller than 4-5% was avoided for daily-flow computations as suggested by Leavesley et al. (1983).

Figure 5: Discretization of 107 HRUs included using watershed subbasins (T), land use land cover (M), and soil type (L).

3.3.4 Post Fire Analysis

To analyze the watershed response to forest fire, variables associated with land cover change were modified. Assumptions were made that the following variables, when increased or decreased, would mimic post-fire conditions (Table 2). Similarly, forest-cover reduction was analyzed by Konrad (2004), where model parameters pertaining to seasonal vegetation cover, shortwave radiation transmitted through the canopy, upper soil storage capacity, and total soil storage capacity were modified. Shapefile data on fire history in the URB was obtained from the U.S. Forest Service at Umatilla National Forest. Areas of fire occurrence in the 1980s, was used as a baseline map from which it was increased three (188,994 ac.) and nine times (190,927 ac.) for the 2050s, and 2080s, respectively. Turner et al. (2015) observed the average area burned per year increased three to nine fold in the Willamette River Basin, Oregon. Two models, HadGEM2 and MIROC5, were run under the 8.5 scenario with different parameters. An extremely warm model, HadGEM2, has acute summer drying and is generally warm year-round and was chosen to represent the RCP 8.5 scenario in the 2080s with high radiative forcings, while MIROC5 is representative of an intermediate, warm model and was chosen to simulate runoff in the 2050s. These parameters were closely related to characteristics of a post-fire response, including loss of vegetation cover, increase in soil hydrophobicity (Soto & Diaz-Fierros, 1998; Neary & Ffolliott, 2005; Versini et al., 2013), reducing soil infiltration capacity and increasing runoff (Letey, 2001; Versini et al., 2013, Springer 2015). Other watershed-scale effects of fire are

differential and depends on burn severity, intensity, extent and the regional hydro-climatic regime (Gresswell, 1999; Sillins, 2014).

Table 2: Model parameters changed for forest cover reduction.

PRMS Parameter	Description	Initial Value	Assigned Value	% Δ
COVDEN_SUM	Summer vegetation cover density	0.5	0.1	-80%
COVDEN_WIN	Winter vegetation cover density	0.5	0.1	-80%
RAD_TRNCF	Solar radiation transmission coeff.	0.3	0.5	40%
SOIL_RECHR_MAX	Max. storage for soil recharge zone	1.64	0.55	-67%
SOIL_MOIST_MAX	Max. value of water for soil zone	2.14-12.54	1.08	-50%

Table 3: Global Climate Models used in this study.

Model Name	Model Agency	Country
CNRM-CM5	Natl. Centre of Meteorological Research	France
HadGEM2-ES	Met Office Hadley Center	UK
CanESM2	Canadian Centre for Climate Modeling & Analysis	Canada
MIROC5	Atmosphere & Ocean Research Inst, Japan & Natl. Inst. for Environmental Studies, Japan	Japan
NorESM1-M	Norwegian Climate Center	Norway
CSIRO-Mk3.6.0	Commonwealth Sci. & Industrial Res. Organizational/Queensland Climate Change Centre of Excellence	Australia
MRI-CGCM3	Meteorological Research Institute	Japan
INM-CM4	Institute for Numerical Mathematics	Russia
BCC-CSM1.1	Beijing Climate Ctr., China Meteorological Admin.	China
GFDL-ESM2M	NOAA Geophysical Fluid Dynamics Laboratory	USA

3.4 Calibration & Verification

An important step in hydrologic model development is model calibration. There are two main approaches to estimate model parameters, an a priori approach using theoretical or empirical relationships in comparison to observed data that

capture the physical characteristics of a watershed, and the second, automatic model calibration (Duan & Gupta et al, 2003). The second approach was used to calibrate the PRMS model by adjusting parameter values to best match simulated runoff values to observed streamflow conditions. The hydrograph peak, volume, and time, were analyzed and compared between observed and simulated flow.

Model calibration was completed using LUCA, version 2.0.0, developed by the USGS and written in the JAVA programming language. LUCA is a multiple-objective, stepwise, automated procedure for model calibration with a graphical user interface, and uses the Shuffled Complex Evolution global search algorithm to calibrate PRMS and models developed with the USGS's Modular Modeling System (MMS) (Duan, 1991; Hay & Umemoto, 2006). MMS is a framework for modeling and is an integrated system of computer software to combine user-selected algorithms to simulate physical processes (Leavesley et al. 1996). Multiple objective calibration allows for multiple parameters such as saturation threshold, Potential Evapotranspiration (PET) sublimation of snow surface to be calibrated with observed streamflow data (Hay & Umemoto, 2006).

LUCA was run at four steps and six rounds using fifteen years of historic streamflow for calibration (1995-2010) and four years for verification (2010-2014). The first step calibrated the water balance, the second step, daily flow timing of all flows, the third step, daily flow timing of high flows, and the fourth

step, daily flow timing of low flows (Table 4). Parameters were adjusted depending on whether the model was over- and under-predicting volume in low and high precipitation years. Final parameter values fall within range of probable parameter ranges (Table 4). The timing of runoff peaks were also analyzed in addition to comparing the monthly annual mean of observed and simulated streamflow (Fig. 6)

Table 4: Tabulated summary of final parameters used for calibration.

Step	Calibration Data Set	Parameters	Final Value	Range
1	Water Balance	rain_cbh_adj_mo	1.128	0.6-1.4
		snow_cbh_adj_mo	1.4	0.6-1.4
2	Daily Flow Timing (all flows)	adjmix_rain_hru_mo	0.4-1.4	0.6-1.4
		cecn_coef	2.12	2.0-10.0
		emis_noppt	0.975	0.76-1.0
		freeh2o_cap	0.019	0.01-0.2
		K_coef	23.859	1-24.0
		potet_sublim	0.541	0.1-0.75
		slowcoef_lin	0.004	0.001-0.5
		soil_moist_max	2.14-12.537	2-10
		soil_rechr_max	1.643	1.5-5
		tmax_allrain_hru_mo	22-52	34-45
		tmax_allsnow_hru	37	30-40
3	Daily Flow Timing (high flows)	fastcoef_lin	0.005	0.001-0.8
		pref_flow_den	0.1	0-0.1
		sat_threshold	3.031-13.955	1.0-15.0
		smidx_coef	0.001	0.001-0.06
4	Daily Flow Timing (low flows)	gwflow_coef	0.024	0.001-0.1
		soil2gw_max	0.103	0-0.5
		ssr2gw_rate	0.582	0.05-0.8
1	Daily Flow Timing (low flows)	gwflow_coef	0.024	0.001-0.5
		gwsink_coef	0.02	0.0-0.05
		soil2gw_max	0.103	0-0.5
		ssr2gw_rate	0.582	0.05-0.8
		soil_moist_max	2.14-12.537	2-10
		slowcoef_sq	0.161	0.05-0.3

LUCA uses streamflow data obtained for a selected basin to calibrate simulated runoff. PRMS requires using unregulated streamflow to calibrate the model parameters in order to determine hydrologic conditions that are representative of the watershed. USGS Stream Gage 14020850, Umatilla River West Reservation Boundary Near Pendleton, OR, was used (Fig. 1). Although the stream gage is regulated and diverted for municipal uses near the City of Pendleton, it was determined through communication with the Department of Natural Resources, CTUIR, that consumptive use is miniscule to the total volume of streamflow and was determined negligible. Otherwise, consumptive use would be calculated and added to the hydrograph to obtain normal streamflow conditions. Consumptive use includes irrigation, municipal, and all other uses, such as domestic needs (Cooper, 2002).

3.4.1 Model Performance Evaluation

To analyze PRMS model performance, four statistical analyses were used: Nash-Sutcliffe efficiency (NSE), percent bias (PBIAS), Kling-Gupta efficiency (KGE), and the Normalized Root Mean Square Error (NRMSE) (Table 5). The Nash-Sutcliffe coefficient R^2 indicates accuracy with a range of negative infinity to one, where one is a perfect fit. A negative coefficient would indicate the mean value of observed data would be a better predictor than the model, and a value of zero would indicate the model would not yield better results than using the average of the flows (Nash and Sutcliffe, 1970; Bathurst, 1997; Krause et. al, 2005).

$$NSE = \frac{\sum_{i=1}^n (y_i^{obs} - y_i^{sim})^2}{\sum_{i=1}^n (y_i^{obs} - y^{mean})^2} \quad (1)$$

The PBIAS determines under or over-prediction of simulated data in comparison to observed data with an optimal value of 0 indicating optimal simulation. A negative value indicates underestimation, while a positive value indicates overestimation of model performance (Moriassi, 2007).

$$PBIAS = \frac{\sum_{i=1}^n (y_i^{sim} - y_i^{obs})}{\sum_{i=1}^n (y_i^{obs})} \times 100 \quad (2)$$

KGE is an alternative to NSE used here, where the closer KGE is to 1 on a scale from negative infinity to one, the more accurate the model is. Different components of the model area are evaluated, such as correlation, bias, and variability (Gupta et al. 2009).

$$KGE = 1 - \sqrt{(r-1)^2 + (\alpha-1)^2 + (\beta-1)^2} \quad (3)$$

NRMSE was also calculated where the lower the value, the lower the variance. RMSE is first calculated to indicate the differences, known as residuals between

simulated and observed values. The lower the percentage, the better the accuracy (Hyndman & Koehler, 2006).

$$NRMSE = \frac{RMSE}{X_{obs,max} - X_{obs,min}}$$

$$RMSE = \sqrt{\frac{\sum_{i=1}^n (X_{obs,i} - X_{model,i})^2}{n}}$$

(4)

Table 5: Tabulated summary of statistics used for calibration and validation.

	NSE	% Bias	KGE	NRMSE
Initial Model Results	0.04	4	0.57	97.7
After Calibration (1995-2010)	0.73	3.5	0.81	52.2
Validation (2010-2014)	0.73	3.5	0.83	52.1

Figure 6: Historical and simulated mean monthly streamflow (1995-2010).

4. Results

4.1.1 Mean Annual Temperature

Model uncertainty is observed when analyzing mean annual temperature for all GCMs for RCP 4.5 (Fig. 7). The range in annual average temperature increases from 8.6°C in the 2020s to 9.5°C in the 2050s, to 10.2°C in the 2080s in RCP 4.5. In RCP 8.5, the average annual temperature increases from 8.8°C in the 2020s, to 10.2°C in the 2050s to 12.1°C in the 2080s (Fig. 8). By the end of the 21st century, there is a 3.3°C increase in mean temperature in RCP8.5, and an increase in 1.6°C in the RCP 4.5 scenario.

Figure 7: Annual mean temperature for all GCMs for RCP 4.5 with the ensemble mean in black.

Figure 8: Annual mean temperature for all GCMs for RCP 8.5 with ensemble mean in black.

4.1.2 Change in Temperature and Percent Change in Precipitation

Air temperature is projected to steadily increase in the future with increased model variability toward the end of the 21st century (Fig. 9). In the 2020s, there is less variation in change in temperature for both scenarios (Fig. 9). In RCP 4.5 in the 2020s, a change in temperature ranging from 0.7 °C to 2.2 °C is observed, and 0.9 °C to 2.2 °C in the RCP 8.5 scenario (Table 6). A percent change in precipitation from -3.4% to 4.5% for RCP 4.5 and -5.7% to 6.3% for RCP 8.5 is observed in the 2020s (Table 6). In the 2050s, change in temperature ranges from 1.2 °C to 3.2 °C for RCP 4.5, and from 1.7 °C to 4.1 °C for RCP 8.5. This includes a range in percent change in precipitation from -2.9% to 7.3% for

RCP 4.5 and -5.3% to 12.3% for RCP 8.5 (Table 6). In the 2080s, change in temperature ranges from 1.6°C to 4.1°C in RCP 4.5, and 3.1°C to 6.6°C in RCP 8.5, where percent change in precipitation ranges from -4.8% to 13.6% in RCP 4.5 and -3.6% to 11.3% in RCP 8.5 (Table 6). With each successive time period, there is a steady increase in the range of change in precipitation and temperature for all GCMs with the exception of a slight drop in percent change in precipitation in the 2080s in RCP 8.5 (Fig. 9).

Figure 9: Change in temperature and annual percent change in precipitation in comparison to historical conditions for all GCMs.

Table 6: Change in temperature (T) and percent change in precipitation (P)

RCP 4.5		% Δ in T	% Δ in P
2020s	min	0.65	-3.40
	max	2.20	4.54
2050s	min	1.16	-2.88
	max	3.20	7.34
2080s	min	1.58	-4.79
	max	4.09	13.57
RCP 8.5			
2020s	min	0.94	-5.69
	max	2.21	6.28
2050s	min	1.73	-5.28
	max	4.10	12.32
2080s	min	3.13	-3.58
	max	6.58	11.30

4.1.3 Seasonal Change in Temperature and Percent Change in Precipitation

Seasonal change in temperature is apparent in both winter and summer, in comparison to the historical period (Fig. 10). There is less model uncertainty in the winter with more variation in the summer season throughout the century in both scenarios for both temperature and precipitation. In the summer, there is substantial variability in model predictions in change in temperature in the 2050s and 2080s. Percent change in precipitation increases slightly and consistently with less GCM uncertainty in the winter in comparison to summer for both scenarios in the 2020s, 2050s, and 2080s (Table 7). In the summer, there is a significant increase in percent change in precipitation with high model uncertainty in both scenarios particularly in the 2050s and 2080s (Table 7).

Figure 10: Change in seasonal temperature (T) and percent change in precipitation (P) for all GCMs in comparison to historical records.

Table 7: Minimum and maximum values of change in seasonal temperature (T) and percent change in precipitation (P) in comparison to historical records.

			Winter		Summer	
RCP 4.5			% Δ in T	% Δ in P	% Δ in T	% Δ P
2020s	min		0.8	-4.1	-4.1	-26.1
	max		2.1	7.1	7.1	4.5
2050s	min		1.1	1.5	1.5	-30.8
	max		3.2	13.8	13.8	3.8
2080s	min		1.6	-4.0	-4.0	-23.1
	max		3.9	19.6	19.6	28.6
RCP 8.5			% Δ in T	% Δ in P	% Δ in T	% Δ P
2020s	min		0.5	-1.1	0.9	-24.2
	max		2.7	15.5	3.0	1.1
2050s	min		1.6	0.4	1.9	-42.0
	max		4.2	20.4	5.8	14.3
2080s	min		3.1	3.5	3.7	-45.6
	max		6.0	17.5	8.9	20.6

Comparing individual GCM performance to observed historical data from 1980-1999 offered insight into the variability of GCM performance. The difference between observed and expected mean annual temperature is substantial (2.63-1.46°C), with a median of 2.32°C ($\mu = 2.30$; $\sigma = 0.40$). The difference in mean annual precipitation between observed and expected was low (0.008-0.016cm), with a median of 0.012cm ($\mu = 0.012$; $\sigma = 0.003$), indicating that GCM simulations of precipitation are similar to observed precipitation.

4.1.4 Center Timing of runoff

CT was observed to occur earlier in the year with each time period in both the RCP 4.5 and RCP 8.5 scenarios (Fig. 11). In the 2020s in RCP 4.5, CT occurs 16 days earlier than the reference period (Table 8). In the 2050s, it occurs 22 days earlier and 26 days earlier in the 2080s, than the reference period in the 1980s (Table 8). For RCP 8.5 in the 2020s, CT occurs 16 days earlier than the reference period, and 24 days earlier in the 2050s, and occurs 33 days earlier in the 2080s (Table 8). Between the two scenarios, CT occurred earlier in the year by 7 days in the RCP 8.5 scenario at the end of the century (Table 8). Mean CT is not significantly different between the historical and future time periods in the RCP 4.5 scenario (p-value=0.25, $\alpha=0.05$) and RCP 8.5 scenario (p-value=0.98) (See Appendix A).

Figure 11: Ensemble mean basin CT for RCP 4.5 and RCP 8.5 scenarios.

Table 8: Mean CT before forest cover reduction.

	RCP 4.5		RCP 8.5	
	μ CT	Date	μ CT	Date
1980s	174	6/23	174	6/23
2020s	158	6/7 (16d)	158	6/7 (16d)
2050s	152	6/1 (22d)	150	5/30 (24d)
2080s	148	5/25 (26d)	141	5/21 (33d)

4.1.5 Seasonal Flows

Precipitation can be held in the snowpack depending on winter temperature, a key factor of streamflow timing and snow melt processes (Hamlet, 2007; Mote et al, 2005, Tohver et al, 2014). Analysis of seasonal flows provides insight into future streamflow behavior. In the URB, winter flows increased from

historic to future conditions by 41% in the 2020s, 64% in the 2050s, and 71% in the 2080s in RCP 4.5 (Table 9, Fig. 12). In RCP 8.5, winter flow increased by 41% in the 2020s, 73% in the 2050s, and 99% in the 2080s. Winter flows ranged from 24.98 m³/s to 30.33 m³/s in RCP 4.5, and from 24.98 m³/s to 35.36 m³/s in RCP 8.5, compared to 17.69 m³/s in historic conditions (Table 9). Winter flows are significantly different between historical and future time periods in both scenarios, except between the 2050s and 2080s in the RCP 4.5 scenario (See Appendix A). There is less variability and a decrease in runoff in the summer (Fig. 12). Summer flow decreased by 39% in the 2020s, 49% in the 2050s, and 53% in the 2080s, in RCP 4.5 (Table 9). In RCP 8.5, summer flow decreased by 38% in the 2020s, 51% in the 2050s, and 66% in the 2080s (Table 9). Mean runoff in the summer decreased 3.64 cm by the end of the century in RCP 4.5. (Table 9). In RCP 8.5, flow decreased 4.55 cm by the 2080s in comparison to the historical period in the 2080s (Table 9). Summer flows are significantly different in the RCP 4.5 scenario, except between the 2050s and 2080s, and between the 2020s and 2050s in the RCP 8.5 scenario (See Appendix A).

Figure 12: Change in basin mean runoff for summer and winter flow

Table 9: Percent change of seasonal flows in comparison to historical conditions.

		Winter (m ³ /s)				Summer (m ³ /s)			
		1980s	2020s	2050s	2080s	1980s	2020s	2050s	2080s
Historical	μ Runoff	17.69				6.916			
	Min	10.3				4.746			
	Max	25.26				10.092			
RCP 4.5	μ Runoff		24.98	29.26	30.33		4.25	3.52	3.28
	Min		15.48	21.74	24.3		2.67	1.99	2.10
	Max		31.79	36.28	39.02		6.26	4.70	5.14
	% Δ		41.21	63.76	71.45		-38.56	-49.13	-52.65
RCP 8.5	μ Runoff		24.98	30.63	35.36		4.27	3.36	2.37
	Min		15.48	22.4	27.99		2.01	2.20	1.54
	Max		31.79	44.5	44.74		6.44	5.13	3.48
	% Δ		41.21	73.15	98.89		-38.30	-51.42	-65.73

4.1.6 Snow Water Equivalent/Precipitation

Winter temperature is a predictor of the timing of spring snowmelt and streamflow in snow dominated and rain-snow transient river basins in the PNW (Hamlet et al., 2007; Mote et al., 2005; Tohver et al., 2014). An important component of the snowpack is snow water equivalent (SWE), the amount of water that will infiltrate with potential for groundwater recharge that may sustain summer flows. The ratio of SWE over P decreases with each time period in the 21st century for both RCP 4.5 and RCP 8.5 scenarios (Fig. 13). In RCP 4.5, SWE/P is observed to decrease 50% in the 2020s in comparison to baseline conditions, -75% in the 2050s, and -75% in the 2080s (Table 10). In RCP 8.5, there is a 50% decrease, -75% in the 2050s, and -96% in the 2080s (Table 10). April 1st SWE decreases with each time period while mean precipitation is steady with slight increases with each time period (Table 10). Precipitation may increase but retention in the snowpack decreases throughout the century indicative of a hydrologic regime shift. All time periods are significantly different from each other at the 0.05 level of significance in the RCP 4.5 (p-value=2.2E-16) and RCP 8.5 (p-value=2.2E-16) scenarios (See Appendix A).

Figure 13: Ensemble mean of the ratio of snow water equivalent to precipitation.

Table 10: Percent change of SWE/P before forest cover reduction.

	Hist.	RCP4.5			RCP8.5		
	1980s	2020s	2050s	2080s	2020s	2050s	2080s
April 1 st μ							
SWE(cm)	6.6	3.2	2.0	1.4	3.2	1.4	0.0
μ P (cm)	75.7	77.4	79.2	78.2	76.7	79.5	80.9
μ SWE/P	0.08	0.04	0.02	0.02	0.04	0.02	0.0036
Min	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Max	0.29	0.21	0.16	0.17	0.24	0.19	0.062
% Δ		-50.0	-75.0	-75.0	-50.0	-75.0	-96.0

4.1.7 Potential Basin Recharge

Quantifying groundwater is difficult, due to the spatial and temporal variability of water below the subsurface. The estimation of aquifer recharge and groundwater availability is critical to water management to meet domestic,

municipal and ecological needs. In the URB, there is a consistent decrease in cumulative potential mean annual recharge in both future scenarios in comparison to historic conditions (Fig. 14). Potential basin recharge decreases 4.3% in the 2020s, -4.8% in the 2050s and -10.3% in the 2080s, in comparison to the 1980s in RCP 4.5 (Table 11). Mean basin recharge is significantly different only between the 2080s and the 1980s, 2020s, and 2050s in RCP 4.5 (See Appendix A). In RCP 8.5, there is a 4.3% decrease in the 2020s, -7.4% in the 2050s, and a 14.8% decrease in the 2080s (Table 11). At the end of the 21st century, mean recharge value decreases 4.4 cm in comparison to the 1980s in RCP 4.5 and decreases 6.3 cm in RCP 8.5 (Table 11). Similar to RCP 4.5, in RCP 8.5 there is a significant difference between the 2080s and the 1980s, 2020s, and 2050s, in addition to a significant difference between the 1980s and 2050s (See Appendix A).

There is a marked shift in peak recharge in both RCP 4.5 and RCP 8.5 scenarios, where peak recharge occurs in March for historical conditions and the 2020s, and shifts to January for both the 2050s and 2080s in RCP 4.5 (Fig. 15). Peak mean recharge by the end of the 21st century decreases by 12.4% but with minimal change from 9.98 cm in the 1980s to 8.74 cm in the 2080s in RCP 4.5 (Table 12). In RCP 8.5, peak mean recharge occurs in March in the 2020s, and shifts to February in the 2050s and to January in the 2080s (Fig. 16). Mean recharge decreases 6.6% in the 2020s and decreases 10.8% in the 2050s, but increases to historical levels with a -0.3% change in the 2080s (Table 12). Peak mean recharge is 9.98 cm in the baseline period and 9.95 cm at the end of the

century, a 0.03 cm difference (Table 12). An increase in the 2080s in RCP 8.5 may be due to decreased canopy interception and less evaporation occurring at the watershed surface with an increased potential for infiltration to occur, contributing to basin recharge. This shift in basin recharge to earlier months is similar to a shift in center timing of flow, occurring earlier in the year as observed by Waibel et al. (2013) in the Deschutes Basin, Oregon.

Figure 14: Cumulative ensemble mean basin recharge.

Table 11: Cumulative mean and percent change in basin recharge.

	Hist.	RCP4.5			RCP8.5		
	1980s	2020s	2050s	2080s	2020s	2050s	2080s
μ Recharge (cm)	42.77	40.93	40.71	38.36	40.93	39.6	36.46
Min	15.23	11.83	8.29	6.64	9.27	6.04	6.31
Max	72.79	75.05	75.62	77.61	80.51	70.65	75.09
% Δ		-4.30	-4.82	-10.31	-4.30	-7.41	-14.75

Figure 15: Ensemble mean for monthly basin recharge in the RCP 4.5 scenario.

Figure 16: Ensemble mean for monthly basin recharge in the RCP 8.5 scenario

Table 12: Mean monthly and percent change in basin recharge.

	Hist.	RCP4.5			RCP8.5		
	1980s	2020s	2050s	2080s	2020s	2050s	2080s
μ Peak Recharge (cm)	9.98	9.79	8.87	8.74	9.32	8.9	9.95
Month	Mar	Mar	Jan	Jan	Mar	Feb	Jan
% Δ		-1.90	-11.12	-12.42	-6.61	-10.82	-0.30

5. Post-Fire Results

5.1.1 Center Time

Land cover change analysis was completed using two GCMs, MIROC5 and HadGEM2-ES, before and after forest cover reduction (pre- and post-fire) in the 2050s and 2080s in the RCP 8.5 scenario. Mean CT, seasonal flows, SWE/P, basin recharge, mean monthly recharge, BFI, and COV of seasonal flows were determined. Mean CT in the 2050s before forest cover reduction occurred 38 days earlier than baseline conditions, and occurred 44 days earlier after forest cover reduction, a difference of six days in the 2050s before and after land cover change (Fig. 17; Table 13). In the 2080s, mean CT occurred 31 days before forest cover reduction, and occurred 35 days earlier than baseline conditions after forest cover reduction, a difference of 4 days between before and after forest cover reduction in the 2080s (Table 13). Mean CT is significantly different between historical conditions and before and after forest cover reduction in both the 2050s and 2080s, but is not significantly different between the two land cover conditions within the 2050s and 2080s (See Appendix B).

Figure 17: Mean Center Timing of flow for MIROC5 and HadGEM2-ES in the RCP 8.5 scenario before and after forest cover reduction.

Table 13: Mean CT before and after forest cover reduction in the RCP 8.5 scenario.

	MIROC5			HadGEM2-ES		
	Historical 1980s	Prefire 2050s	Postfire 2080s	Historical 1980s	Prefire 2020s	Postfire 2080s
Δ in days in μ CT	(6/30)	38d (5/23)	44d (5/17)	(6/22)	31d (5/22)	35d (5/18)

5.1.2 Seasonal Flows

Winter runoff in the 2050s showed a 133% increase before forest cover reduction and a 151% increase after forest cover reduction (Fig. 18; Table 14). In the 2080s, winter runoff increases 80% before land cover change and increases

91% after land cover change (Table 14). Winter flows are significantly different between historical and both land cover conditions in the 2050s and 2080s, but is not significantly different between before and after forest cover reduction within the 2050s and 2080s (See Appendix B). A decrease in summer flows is observed for both before and after forest cover reduction. A 63% decrease in both land cover conditions is observed in the 2050s, and a 73% and 68% decrease before and after forest cover reduction in the 2080s (Fig. 18; Table 14). In the 2050s, summer runoff between historical and before and after forest cover reduction are significantly different, while there is no significant difference in the 2080s, between historical and before and after forest cover reduction ($p\text{-value}=0.48$) (See Appendix B).

Figure 18: Seasonal flows for MIROC5 and HadGEM2-ES in the RCP 8.5 scenario before and after forest cover reduction.

Table 14: Percent change of seasonal flows before and after forest cover reduction in the RCP 8.5 scenario (WTR=Winter; SMR=Summer).

MIROC5						
	1980s Hist.		2080s Prefire		2080s Postfire	
μ Runoff (cm ³ /s)	WTR	SMR	WTR	SMR	WTR	SMR
	15.93	8.85	37.14	3.30	39.95	3.26
Min	1.54	2.34	18.07	1.08	21.48	1.27
Max	38.36	19.59	67.74	7.88	69.41	9.54
% Δ			133.15	-62.7	150.8	-63.2

HadGEM2-ES						
	1980s Hist.		2080s Prefire		2080s Postfire	
μ Runoff (cm ³ /s)	WTR	SMR	WTR	SMR	WTR	SMR
	18.94	7.12	33.92	1.93	36.13	2.31
Min	1.79	1.01	6.34	0.32	9.57	0.39
Max	42.67	20.31	80.30	9.18	80.86	10.24
% Δ			79.1	-72.9	90.8	-67.6

5.1.3 Snow Water Equivalent/Precipitation

The ratio of SWE to P is substantially lower in the 2050s in comparison to the 1980s with an 85% decrease before forest cover reduction and 90% decrease after forest cover reduction (Fig. 19; Table 15). The absolute difference is 0.088 and 0.094, before and after forest cover reduction, in the 2050s in comparison to the 1980s (Table 15). Between the two land cover conditions in the 2050s, there is a 38% decrease in SWE/P (Table 15). In the 2080s, the ratio is significantly lower with a 99% decrease in both before and after forest cover reduction and an absolute difference in 0.073 in the 2080s in comparison to historical conditions (See Appendix B). This is indicative of earlier snowmelt, winter runoff and decreased summer baseflow. This does not consider expected variability at different elevations and aspect. There is significant difference between both land cover conditions in comparison to historical conditions in the

2050s and 2080s, and no significant difference between the 2050s and 2080s before and after forest cover reduction (See Appendix B).

Figure 19: Ratio of snow water equivalent to precipitation before and after forest cover reduction in the RCP 8.5 scenario.

Table 15: Ratio of snow water equivalent to precipitation before and after forest cover reduction in the RCP 8.5 scenario.

	MIROC5			HadGEM2-ES		
	Historical	Before LCC	After LCC	Historical	Before LCC	After LCC
	1980s	2050s	2050s	1980s	2080s	2080s
μ SWE (cm)	8.4	1.69	1.15	6.38	8.7E-02	4.4E-02
μ P (cm)	77.2	88.61	88.61	76.96	79.89	79.89
μ SWE/P	1.04E-01	1.6E-02	0.01	7.4E-02	1E-03	5E-04
% Δ		-84.6	-90.4		-98.6	-99.3

5.1.4 Potential Basin Recharge

Cumulative potential basin recharge decreases in the 2050s and 2080s by 2% and 27% in comparison to historical conditions before forest cover reduction (Fig. 20; Table 16). After forest cover reduction, recharge increases 8% in the 2050s while in the 2080s, it is 11% less than historical conditions (Table 16). Basin recharge is not significantly different between historical conditions and both before and after forest cover reduction in the 2050s ($p\text{-value}=0.38$) (See Appendix B). In the 2080s, recharge is significantly different between the 1980s and before forest cover reduction, and not significantly different between the 1980s and after forest cover reduction, and not significantly different between both land cover conditions in the 2080s (See Appendix B). Recharge is expected to decrease with increased runoff due to decreased interception, and sublimation of snow from tree or vegetation (Konrad et al., 2014), however there is an increase in recharge in the 2050s after forest cover reduction by 4 cm in comparison to historic conditions (Table 16). Improved characterization of aquifer recharge in the CRBGs would offer further insight.

Figure 20: Mean basin recharge before and after forest cover reduction.

Table 16: Percent change of mean basin recharge before and after forest cover reduction.

	MIROC5			HadGEM2-ES		
	Historical 1980s	Before LCC 2050s	After LCC 2050s	Historical 1980s	Before LCC 2080s	After LCC 2080s
μ Recharge	44.55	43.47	48.31	43.73	32.14	38.99
Min	20.82	17.56	22.84	19.65	7.56	13.9
Max	69.88	70.65	74.09	64.26	51.48	62.13
% Δ		-2.42	8.44		-26.5	-10.84

In the 2050s, peak basin recharge occurs in April in historical conditions (11cm), and then occurs in January, decreasing 8% (10.3 cm) and 4% (10 cm) in recharge before and after forest cover reduction (Fig. 21; Table 17). Peak mean recharge in the 2080s, occurs in March (10.3cm) in historic conditions and shifts

to January decreasing 17% (9cm) and 10% (9.2cm), before and after forest cover reduction(Fig. 22; Table 17).

Figure 21: Mean basin monthly recharge for MIROC5 before and after forest cover reduction.

Figure 22: Mean basin monthly recharge for HadGEM2-ES before and after forest cover reduction.

Table 17: Peak mean basin recharge and peak month

	MIROC5			HadGEM2_ES		
	Historical	Before LCC	After LCC	Historical	Before LCC	After LCC
μ mo. Recharge (cm)	10.68	9.84	10.27	10.29	8.55	9.22
Peak mo.	April	January	January	March	January	January
% Δ		7.87	3.84		16.91	10.40

5.1.5 Base-Flow Index

Groundwater discharge to surface flow was determined by the BFI, the ratio of base flow to total stream flow. In the 1980s, mean BFI determined 72% groundwater contributions to surface flows (Table 18). This is sustained throughout the 21st century for all GCMs before forest cover reduction, dropping slightly to 69% for both scenarios at the end of the century (Table 18). After

forest cover reduction, MIROC5 predicts 70% groundwater contributions for both land cover conditions, a 3-5% decrease, while HadGEM2-ES predicts 67% and 66% groundwater contributions in both land cover conditions, a 13% increase from historical conditions (Appendix B). Before and after land cover change, there is minimal change under 5%, whereas in the 2080s after forest cover reduction, BFI increases by 13% at the end of the century (Table 18). This is individual model performance, and may vary across each GCM.

Table 18: Base-flow Index (BFI) and percent change before and after forest cover reduction.

Baseflow Index (%) Before Forest Cover Reduction				
	RCP 4.5	RCP 8.5	% Δ RCP 4.5	% Δ RCP 8.5
Historical	0.72	0.72		
2020s	0.70	0.69	-2.91	-4.16
2050s	0.69	0.69	-3.74	-4.44
2080s	0.69	0.69	-4.44	-4.85
Baseflow Index (%) After Forest Cover Reduction				
	MIROC5 RCP 8.5	HadGEM2-ES RCP 8.5	% Δ BFI MIROC5	% Δ BFI HadGEM2-ES
Historical	0.72	0.59		
2050s Prefire	0.69		-4.85	
2050s Postfire	0.70		-3.46	
2080s Prefire		0.67		13.10
2080s Postfire		0.66		12.76

5.1.6 Coefficient of Variation

High variability across the GCMs is generally observed in the summer months in both scenarios before forest cover reduction, where it is highest in the 2020s and 2050s in the RCP 8.5 scenario (Fig. 23). In the winter, variability is higher in the historical period for both RCP 4.5 and RCP 8.5, where higher uncertainty or wider range of GCM performance would result in a smaller mean and higher CV

value (Fig. 23). After forest cover reduction, variability is higher in the summer and increasingly in the 2005s and 2080s in RCP 8.5 (Fig. 24) In the winter after forest cover reduction, uncertainty is highest in the 2050s in RCP 4.5 (Fig. 24). In general, there is more certainty or increased confidence in GCM prediction of winter flows mid-century and onward.

Figure 23: Coefficient of Variation of seasonal flows before forest cover reduction.

Figure 24: Coefficient of Variation of seasonal flows after forest cover reduction.

6. Discussion

Model calibration may have been improved with a second stream gage with continuous historic flow available upstream on Meacham Creek, a tributary to the Umatilla River that provides a little over 50% of summer flows (CTUIR 2012). This would have required a second calibration for this drainage area and could be explored in future work. Consumptive use, which includes irrigation, municipal, and all other uses, such as domestic needs (Cooper 2002), was not calculated and deemed negligible because diversions at the USGS stream gage, Umatilla River West Reservation Boundary Near Pendleton, is minimal to total volume of streamflow. (Kate Ely, personal communication, April 23, 2013).

The temporal and spatial behavior of groundwater in a heterogeneous geologic structure of the CRBGs could not be delineated with PRMS alone and is outside the scope of this study. The CRBGs as a part of the CPRAS covers 70,811 km² and warrants a regional approach in understanding groundwater behavior (Cherkauer, 2004). By choosing to use PRMS, an assumption was made that the hydrogeology in the study area is similar to an unconfined groundwater system (Cherkauer 2004). Another assumption was made in calculating BFI, where well withdrawal is negligible to total flows (Cherkauer 2004).

6.1.1 Temperature and Precipitation

A projected warming rate in the western U.S. is 0.1-0.6°C/decade (Dickerson-Lange & Mitchell, 2014). In the URB, there is high uncertainty and

variability across the GCMs as can be seen in the wide variation of temperature and precipitation change and increase in each time period in the 21st century (Fig. 9). Mean temperature increases 3.3°C by the end of the century in RCP 8.5, similar to a +3.2°C increase by the 2080s predicted by Chang & Jung (2010) in the Willamette River, OR. Dickerson-Lange and Mitchell (2014) predicted a 1.8 to 3.5 °C mean increase in spring and summer temperatures by the 2050s in one scenario in Northwestern Washington. Precipitation is variable in summer flows, and increases as much as 11.3% in RCP 8.5 by the end of the century in the URB (Fig. 9, Table 6), where a 15-21% increase is seen in northwestern WA in two models (Dickerson-Lange & Mitchell, 2014). This is in agreement with Vynee et al. (2010), who observed approximately a 10% to 18% increase in precipitation by mid and end of the century in the URB. With increased temperatures and less snow to hold increased precipitation, the frequency and magnitude of floods is predicted to increase (Dickerson-Lange et al., 2014). It will become more important to understand temperature and precipitation trends, two primary factors in determining the timing and magnitude of streamflow in western rivers.

6.1.2 Snow Water Equivalent and Precipitation

April 1st SWE is a function of winter accumulation and ablation. There is a clear pattern that SWE substantially decreases with each time period, indicating a hydrologic regime shift from a snow-rain dominant to rain-dominant basin. This is consistent with predictions in the Pacific Northwest (Stewart et al. 2005, Mote et al. 2005, Hamlet 2011; Jung & Chang, 2011; Safeeq et al. 2013; Dickerson-

Lange et al. 2014). Vynee et al. (2010) predicted SWE to decrease >50% by the 2080s in the URB. A considerable change in basin area-weighted SWE has been observed to affect mid-elevation areas in the rain and snow transition zone (Mastin, 2011). In post-fire conditions, there is a substantial decrease in SWE in the 2080s for both land cover conditions. This could be due to varying energy-balances at the land and atmosphere interface, including radiative fluxes and changes in albedo, which can significantly influence melting snow rate and the intensity of reflection by snow cover. Albedo was observed to be higher after a forest fire and lower after afforestation (Oris et al., 2014). Warming trends and increased precipitation should be observed to see at what elevations or aspects, snowpack contributes to aquifer recharge and summer baseflow. Analysis of montane snowpacks that store winter precipitation and provide water for the rest of year is required for adaptation planning in dam releases and flood control (Molotch et al., 2004; Jepsen, 2012).

6.1.3 Runoff Behavior

Precipitation and temperature are the main drivers of the magnitude and timing of streamflow (Dickerson-Lange & Mitchell, 2014). At the end of the 21st century, before forest cover reduction, mean CT occurs earlier in the year by 4.7 weeks. There is more variability in mean CT after forest cover reduction, which may be a result of individual model behavior. Post-fire parameters including an 80% decrease in both summer and winter cover density, and a 40% increase in the solar radiation transmission coefficient (Table 2) may have more effect on

peak discharge during individual precipitation events than CT, a day in the water year half the runoff has occurred. Runoff trends even if subtle, can be detrimental to fish habitat and growing seasons of wheat and green peas in the URB, for instance.

Seasonal flows in comparison to baseline conditions are predicted to change with each successive time period, where winter flows show an increase as great as 98% and summer flows a decrease as low as 66% at the end of the century before forest cover reduction (Table 9). The same holds true after forest cover reduction, with a greater increase in winter flows by 150% in the 2050s and 91% in the 2080s (Table 14). An increase in the ratio of winter rainfall to winter snowfall is observed here, where precipitation is not being held in the snowpack due to warming temperatures as seen across the western United States. In the region, Jung & Chang (2011) observed negative runoff trends in the spring and summer, and positive trends in the fall and winter in the Willamette River Basin. Similarly, Dickerson-Lange (2014) observed increases in winter discharge from 34 to 60% by midcentury and decreases in summer flows from -20 to -30% in Northwestern, WA. In the Deschutes Basin in central Oregon, winter flows are projected to increase 80 to 115% in the Cascade Range (Waibel et al., 2013).

6.1.4 Potential Basin Recharge & Base-flow

Contrary to a substantial decrease in SWE/P, mean basin recharge is projected to remain within range of historic levels with slight declines throughout the 21st century, with the exception of a sudden decrease in the 2080s for both

scenarios before forest cover reduction. Basin recharge estimates range from 36-41 cm/yr, within range of previous studies of 2.0 to 36.0 cm/yr (Spane & Weber, 1995; Bauer & Vaccaro, 1990; Brown 2009). After forest cover reduction, mean recharge increases and is greater than historic levels in the 2050s. This is based on individual model performance, otherwise a decrease would be expected due to a reduction of soil porosity and the formation of hydrophobic soils after a fire, reducing infiltration rates. Potential mean recharge for the CPRAS is estimated to be 11.6 cm/yr (Burns et al. 2012). Similar to potential basin recharge, groundwater contributions to surface flow is projected to be sustained where the BFI remains within historic levels with slight declines in both RCP 4.5 and RCP 8.5 scenarios before forest cover reduction, and in the 2050s after forest cover reduction. With the exception of a 13% increase in groundwater contributions to surface flows in the 2080s as predicted by HadGEM2-ES after forest cover reduction, this may be due to the predicted substantial increase in precipitation and increased winter runoff in the 2080s. Again, predictions vary by GCM and the ensemble mean could be different.

The decline of groundwater levels in the URB has been addressed by The City of Pendleton, where the Aquifer Storage and Recovery program (ASR) lowered the city's dependence on groundwater from 62% to 3%. Since then, groundwater declines were observed to be 340 cm/year and down to 200 cm/yr after ASR has been implemented in 2004 (Pendleton Public Works, Water Division, 2012).

6.1.5 Future Work

A significant next step would be to characterize at different spatial and temporal scales, the behavior of ground- and surface-water interactions using a numerical groundwater model and further analyze the effects of fire on runoff with a landscape simulation model. A dynamic global vegetation model may help identify parameters that account for regrowth, fire type and intensity, to improve understanding of the effects of fire-burns on a watershed system. Soil water repellence for example, has been found to last anywhere from one to six years, where a shorter temporal scale may best capture watershed response to fire (Henderson & Golding, 1983; Macdonald & Huffman, 2004). PRMS files may be adapted to combine with MODFLOW, a numerical groundwater model, to input to GSFLOW, a coupled groundwater and surface-water flow model, to increase understanding of the spatial and temporal behavior of groundwater. In addition, spatial analysis at a finer scale of water budget variables according to aspect and elevation will only enhance localized efforts for management of ecosystem services.

7. Conclusions

Increasing global mean temperature and changing precipitation are driving factors in runoff behavior. The uncertainty in the affects of climatic change and variability, and anthropogenic influences on a hydrologic regime makes it imperative to study their effects on natural resources. Using PRMS, a runoff model was calibrated for the upper URB, to characterize trends in runoff, snowpack, recharge, and other components of the water budget to understand water availability in changing climate and forest cover reduction. The effects of fire and climate shifts on runoff behavior, is largely understudied in the URB making this study unique.

A hydrologic regime shift is observed in the URB, from a snow-rain dominated to rain-dominated basin. Increased winter flows and decreased summer flows are predicted throughout the 21st century. By the end of the century, a 1.6°C increase in RCP 4.5 and 3.3°C increase in RCP 8.5 in basin temperature is observed. With increasing temperatures, less precipitation is held in the snowpack and peak basin runoff occurs earlier in the year in both land cover conditions, with decreased summer flows throughout the 21st century. This is amplified in winter flows in the RCP 8.5 scenario mid-century and onward. The ratio of SWE/P is shown to significantly decrease in both scenarios across the century before forest cover reduction. After forest cover reduction, similar trends in mean CT, seasonal flows and SWE/P, are observed with a substantial decrease in SWE/P by the 2080s and increase in winter flows in the 2050s.

Mean basin recharge is sustained throughout the 21st century with slight declines in each subsequent time period before forest cover reduction, while after post-fire simulation, basin recharge increases. Similarly, groundwater contributions is predicted to remain constant with slight declines throughout the 21st century before forest cover reduction, and a 13% decline in RCP 8.5 after forest cover reduction. Due to the complexity of groundwater behavior in the CRBGs, basin recharge should be explored further with a numerical groundwater flow model.

This study provides further insight to secure freshwater resources for ecosystem function and cultural resources in the URB. It is a viable method in which to improve adaptive capacity, including flood control, dam releases, and in-stream flow restoration practices. Collaborative partnerships across the region with entities such as CTUIR, the Columbia Basin Ground Water Management Area, and U.S. Forest Service will only improve resource assessments and plan for extreme climatic events.

The methods used in this study are adaptable and replicable in western watershed basins where temperature and precipitation are the driving factors of runoff behavior, and snow water equivalent is a sensitive, strong indicator of a changing hydrologic regime. The ensemble mean change provides a consensus of trend, variability and range of GCM projections (Mote et al., 2011). In the face of climate uncertainty, watershed modeling proves to be a valuable tool in

identifying runoff trends to assist in making informed water management decisions and can only strengthen climate preparedness.

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APPENDIX A: Summary of Statistical Analyses Comparing Key Climate Indices Before Forest Cover Reduction

Calculated p-values in RCP 4.5							
	1980s vs. 2020s	1980s vs. 2050s	1980s vs. 2080s	2020s vs. 2050s	2020s vs. 2080s	2050s vs. 2080s	Test
Basin Runoff	--	--	--	--	--	--	9.78E-01*
Mean CT	--	--	--	--	--	--	2.52E-01*
Winter Flows	0.00	0.00	0.00	0.00	0.00	6.87E- 01	2.20E-16*
Summer Flows	0.00	0.00	0.00	3.38E- 03	1.81E- 05	4.92E- 01	2.20E-16*
SWE/P	(254.12, 74.04)	(400.62, 74.04)	(511.09, 74.66)	(146.50, 73.42)	(256.97, 74.04)	(110.48, 74.04)	2.20E-16**
Basin Recharge	(50.00, 74.041)	(59.97, 74.04)	(134.44, 74.66)	(9.97, 73.41)	(84.43, 74.04)	(74.46, 74.04)	3.90E-05**
Calculated p-values in RCP 8.5							
Basin Runoff	--	--	--	--	--	--	3.60E-01*
Mean CT	--	--	--	--	--	--	9.78E-01*
Winter Flows	0.00	0.00	0.00	8.90E- 06	0.00	2.11E- 04	2.20E-16*
Summer Flows	(35.34, 23.50)	(55.01, 23.5)	(80.17, 23.70)	(19.67, 23.30)	(44.83, 23.50)	(25.17, 23.50)	2.20E-16**
SWE/P	(230.97, 74.04)	(425.31, 74.04)	(662.39, 74.66)	(194.34, 73.41)	(431.43, 74.04)	(237.09, 74.04)	2.20E-16**
Basin Recharge	(60.91, 74.041)	(96.48, 74.04)	(176.17, 74.66)	(35.57, 73.41)	(115.27, 74.04)	(79.69, 74.04)	7.91E-09**

*ANOVA

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APPENDIX B: Summary of Statistical Analyses Comparing Key Climate Indices After Forest Cover Reduction

Calculated p-values in the RCP 8.5 Scenario					
Index	GCM Model	Hist. vs. Before LCC	Hist. vs. After LCC	Before LCC vs. After LCC	Test
Mean CT	MIROC5	0.00	0.00	3.66E-01	2.20E-16*
	HadGEM2-ES	0.00	0.00	5.29E-01	8.15E-16*
Winter Flows	MIROC5	0.00	0.00	6.36E-01	6.49E-12*
	HadGEM2-ES	3.37E-04	3.71E-05	8.22E-01	9.88E-06*
Summer Flows	MIROC5	0.00	0.00	9.99E-01	2.04E-10*
	HadGEM2-ES	--	--	--	4.80E-01**
Basing Recharge	MIROC5	--	--	--	3.78E-01*
	HadGEM2-ES	2.03E-03	3.26E-01	1.01E-01	2.95E-03*
SWE/P	MIROC5	(35.49, 19.75)	(41.13, 19.75)	(5.63, 19.58)	3.17E-10**
	HadGEM2-ES	(42.52, 19.47)	(43.45, 19.47)	(0.93, 19.471)	3.39E-14**

*ANOVA

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APPENDIX C: Comparison of Model Performance

GCM Model	Precipitation (cm)			Temperature (°C)		
	μ Obs.	μ Exp.	Obs-Exp	μ Obs.	μ Exp.	Obs-Exp
BCC-CSM1.1	1.80	1.78	0.01	60.94	58.35	2.59
CanESM2	1.80	1.78	0.01	60.99	58.35	2.64
CNRM-CM5	1.79	1.78	0.01	60.99	58.35	2.64
CSIRO-Mk3.6.0	1.80	1.78	0.01	60.11	58.35	1.76
GFDL-ESM2M	1.79	1.78	0.01	60.62	58.35	2.27
HadGEM2-ES	1.79	1.78	0.01	60.64	58.35	2.29
INM-CM4	1.80	1.78	0.01	60.98	58.35	2.63
MIROC5	1.80	1.78	0.02	59.81	58.35	1.46
MRI-CGCM3	1.80	1.78	0.01	60.71	58.35	2.35
NorESM1-M	1.79	1.78	0.01	60.60	58.35	2.25