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The 2015 drought in Washington State: a harbinger of things to come?

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Abstract

Washington State experienced widespread drought in 2015 and the largest burned area in the observational record, attributable in part to exceptionally low winter snow accumulation and high summer temperatures. We examine 2015 drought severity in the Cascade and Olympic mountains relative to the historical climatology (1950–present) and future climate projections (mid-21st century) for a mid-range global greenhouse gas emissions scenario. Although winter precipitation was near normal, the regional winter temperature anomaly was +2.1 °C ($+2.0\sigma$) in 2015, consistent with projections of a +2.3 °C ($+2.2\sigma$) temperature change and near normal precipitation in the future, relative to the climatology. April 1 snow water equivalent in 2015, −325 mm (-1.5σ), and the future, −252 mm (-1.1σ), were substantially lower than the climatology. Wildfire potential, as indicated by dead fuel moisture content, was higher in 2015 than mid-21st century mean projections. In contrast to most historical droughts, which have been driven by precipitation deficits, our results suggest that 2015 is a useful analog of typical conditions in the Pacific Northwest by the mid-21st century.

1. Introduction

As in the rest of the western US, rising temperatures affect winter snow accumulation and the timing of spring runoff in Washington state, with serious implications for water availability later in the year when water demand is highest (Barnett *et al* 2005, Anderson *et al* 2016). Temperatures across Washington and the northwestern US have increased by approximately 0.8 °C over the 20th century, with most of the warming over the past half century (Mote and Salathé 2010, Abatzoglou *et al* 2014). This has been associated with regional declines in spring snow water equivalent (SWE) (Mote *et al* 2005), earlier spring runoff (Stewart *et al* 2005), increased flood risk in warm transient basins (Hamlet and Lettenmaier 2007), earlier spring soil moisture recharge, increased evapotranspiration (Hamlet *et al* 2007), and higher fire activity (Westerling 2006, 2016). In the Pacific Northwest, there have been significant declines in annual streamflow since the mid-20th century and more extreme dry years (Luce and Holden 2009). Although interannual to decadal timescale variations, such as the El Niño-Southern Oscillation and the Pacific Decadal Oscillation, drive some of these observed changes, the long-term warming is largely controlled by anthropogenic climate change (Mote *et al* 2005, Stewart *et al* 2005, Abatzoglou 2010, Abatzoglou *et al* 2014). This anthropogenic signal has been attributed to ~60% of changes in western hydrology (Barnett *et al* 2008) and a doubling of burned area in western forests (Abatzoglou and Williams 2016).

Climate projections suggest substantial future changes in climate will occur across the western US. Rising regional temperatures are expected to result in earlier peak snowmelt by 30–40 days (Stewart *et al* 2004) and up to two months (Rauscher *et al* 2008) by the late 21st century (relative to the mid-20th century)

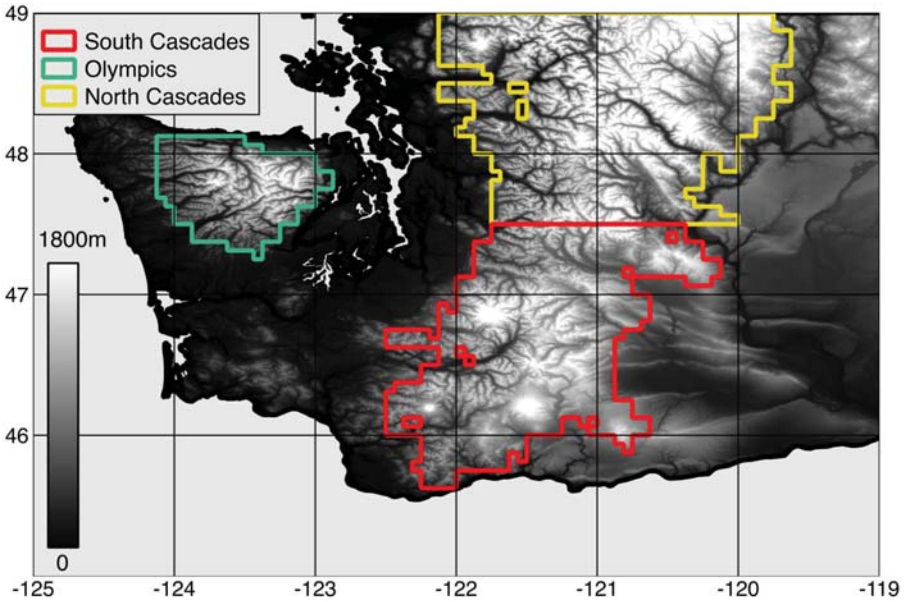

Figure 1. Digital Elevation Model (DEM) showing elevation (in meters) across Washington State with Cascade and Olympic mountain ranges (Jarvis *et al* 2008).

and could drive a 45% decline in net snow accumulation from 2000–2060 (Mankin and Diffenbaugh 2015). In the Pacific Northwest, Rupp *et al* (2016) estimated 1.5 °C–4 °C of warming will occur by the mid-21st century (compared to 1970–1999), with relatively small changes in annual precipitation. These changes are projected to result in substantial reductions in spring snowpack (~80% in the Cascades) as well as declines in summer soil moisture by the 2080s in Washington State (Gergel *et al* 2017). Consequences of these changes include insufficient reservoir storage (Barnett *et al* 2005), rising stream temperatures, and lower summer streamflows, which will negatively impact salmon populations (Mantua *et al* 2010), and climate and disturbance changes, which will alter forest composition (Littell *et al* 2010). Climate projections also suggest increased wildfire activity in Washington as a result of increased summer temperature and moisture deficits (Littell *et al* 2010, Sheehan *et al* 2015) with cascading impacts ranging from ecosystem services to regional air quality degradation from fire emissions (Liu *et al* 2016).

Given these concerns, we assess the severity of the 2015 drought in Washington in comparison to projected climate change using a suite of metrics. 2015 was a remarkable year across much of the West, with over 80% of long-term snow courses west of 115°W longitude reporting record low April 1 SWE (Mote *et al* 2016). The low snowpack in Washington was strongly related to exceptionally warm winter temperatures as winter precipitation was near normal across much of the state. The potential for water shortages was masked somewhat by the near normal winter precipitation (and above normal winter runoff in some watersheds as a result of high snowlines). However, by early May it was apparent that at least for some of the state's snow-dominant river basins, primarily those draining the eastern slopes of the Cascades, early snow runoff would lead to drought conditions later in the summer (in this region, $<$ 20% of the water year's precipitation falls after April 30). Warm winter temperatures persisted into the summer; as we show below, over parts of eastern Washington summer temperatures were the highest of the observational record. Washington state saw over 400 000 ha burned during 2015, the largest in the observational record, with most of the fire activity occurring in forested lands and fire suppression costs exceeding $560 million USD (US Department of Agriculture 2016). Other impacts include reductions of the yield and quality of crops across Washington State (resulting in economic losses of $633–$773 million USD) (Sandison 2017), and a historically low proportion of Sockeye Salmon that reached spawning grounds due to reduced early summer flows and exceptionally warm temperatures (Faulkner *et al* 2017).

Here, we contextualize the 2015 drought relative to the historical record from 1950 to present. We then compare 2015 with projections of future climate for the mid-21st century (2040–2069) to determine the extent to which 2015 can be considered typical of future hydroclimatic norms.

2. Data

2.1. Overview

Our study region includes the Olympic and Cascade mountain ranges of Washington, which, following the definition used by Mao *et al* (2015), we defined as having long-term historical April 1 SWE $>$ 50 mm (figure 1). We further divide the Washington Cascades at 47.53°N, into the South Cascades and North

Cascades, which in our partitioning have approximately equal areas. For both 2015 and the future climate, we examine differences in winter (November–March) temperature, winter (November–March) precipitation, April 1 SWE, total runoff (January–July), the center of mass of total runoff, and summer (June–September) dead fuel moisture as a proxy for fire risk (as in Gergel *et al* (2017)).

2.2. Historical observations

The historical hydroclimatic data we used are from the Livneh *et al* (2013) database, extended from December 31, 2011 to December 31, 2013 by Livneh *et al* (2015) and for this study through 2015. The Livneh *et al* data include precipitation, daily air temperature maxima and minima, and other hydroclimatic data such as downward solar and longwave radiation and humidity at 1/16° latitude by longitude and daily time step over the conterminous US and Canada south of 53°N. The dataset also includes model-derived SWE, soil moisture, and runoff data (Liang *et al* 1994). Recent updates resolve certain issues in the original Livneh *et al* (2013) data set, and these corrections were incorporated in the extensions to 2015 as well. One minor difference in the recent updates is that the monthly orographic precipitation scaling procedure (used in all versions) uses a reference period of 1981–2010, rather than 1961–1990 in the earlier versions. The orographic adjustment procedure, which also was changed slightly in Livneh *et al* (2015), had a slight effect on precipitation totals.

We extracted SWE from the data set for April 1, which is a common indicator of summer water availability. Like soil moisture and runoff, the SWE values were produced by the variable infiltration capacity (VIC) macroscale hydrology model, which is a spatially distributed model that solves the water and energy balance at each grid cell. In addition to gridded temperature and precipitation, other forcing variables required to run the VIC model (as archived in the Livneh *et al* data set) include downward solar and infrared radiation, humidity, and surface wind climatology with sub-daily variables estimated from daily values (Bohn *et al* 2013).

2.3. Future projections

We used future climate projections from statistically downscaled output from the Coupled Model Intercomparison Project (CMIP5) for the midrange global emissions scenario (Representative Concentration Pathway (RCP) 4.5) (Taylor *et al* 2012). These data are publicly available from the Lawrence Livermore National Laboratory's Program for Climate Model Diagnosis and Intercomparison. Rupp *et al* (2013) evaluated the performance of 41 CMIP5 global climate models (GCMs) for the 20th century over the Pacific Northwest and found that the models in general better reproduced temperature-based metrics than precipitation.

As described in detail by Gergel *et al* (2017), daily temperature, precipitation, and wind were statistically downscaled from the native coarse resolution of the GCMs (table S1 available at stacks.iop.org/ERL/12/114008/mmedia) to 1/16° spatial resolution using the multivariate-adaptive constructed analogs (MACA) method (Abatzoglou and Brown 2012) and the Livneh *et al* (2013) data as a training dataset. The downscaling ensures that climate normals for individual downscaled variables such as temperature and precipitation are identical between the historical modeling experiments (1950–2005) and observational record used for model training (1950–2011). We used 10 GCMs identified by Rupp *et al* (2013) based on their ability to credibly simulate attributes of climate across the Pacific Northwest (table S1). Statistically downscaled (and bias corrected) output was then used to force the VIC model (Version 4.1.2). The downscaling and bias correct procedure assured that the historical portions of the climate scenarios (all of which were continuous from 1950 through 2099) were statistically consistent with the 2015 VIC forcings. The VIC model was run in energy balance mode using 1/16° spatial resolution daily data as described above. More details of the implementation of the statistical downscaling approach and the VIC model simulations are provided by Gergel *et al* (2017). The archived downscaled climate model output, along with the VIC simulations forced by downscaled climate model output, are available from the Integrated Scenarios of the Future Northwest Environment project (http://climate.nkn.uidaho.edu/IntegratedScenarios/index.php).

2.4. Fire risk

As in Gergel *et al* (2017), we quantified changes in fire risk based on the dead fuel moisture (DFM) content. Based on equations developed by Simard (1968), the US National Fire Danger Rating system first quantifies the maximum and minimum daily equilibrium moisture content (EMC) of dead woody material in steady-state conditions based on temperature and relative humidity (Cohen and Deeming 1985). DFM is then calculated for multiple sizes of fuel classes; we focus here on 100 and 1000 hour time lags, which represent larger diameter fuels (1–3″ and 3–8″, respectively) that have a slower response to reaching equilibrium with changing environmental conditions than fine fuels (1 and 10 hour time lags) (Cohen and Deeming 1985). The 100 hour time lag is based on 24 hour average environmental conditions while the 1000 hour time lag is based on a seven day average (Cohen and Deeming 1985). Previous studies have demonstrated strong relationships between 100 and 1000 hour DFM averaged over the fire season and burned area extent and the occurrence of very large wildfires across the Northwest (e.g. Abatzoglou and Kolden 2013, Stavros *et al* 2014). We performed the DFM calculations at 1/16° spatial resolution with temperature (minimum

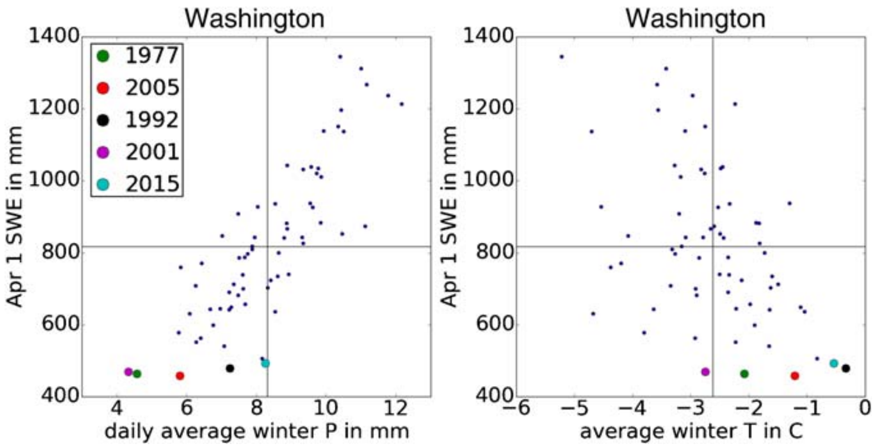

Figure 2. 1950–2015 April 1 snow water equivalent (SWE, mm) versus (*a*) daily winter (November–March) precipitation rate (mm day^{-1}) and (*b*) average winter (November–March) temperature (°C). 2015, as well as the four years with the lowest SWE, are highlighted by colored circles. Horizontal and vertical lines mark the climatological mean.

Table 1. Mean winter (November–March) temperature (°C), winter (November–March) precipitation rate (mm/day), April 1 SWE (mm), total six month (January–July) runoff (km^3), Julian day with center of mass (COM) of total runoff (day of year), and summer (June–September) dead fuel moisture (DFM, %) by location. For 2015 (b) and future projections (c), the standardized anomalies relative to the 1950–2015 climatology are given in parentheses. Regions are denoted in figure 1.

(a) 1950–2015				
	All regions	South cascades	North cascades	Olympics
Temperature (°C)	−2.6	−1.5	−4.2	0.2
Precipitation (mm d^{-1})	8.3	7.8	7.3	14.5
SWE (mm)	817.6	683.5	913.9	901.7
Total Runoff (km^3)	53.45	19.70	23.13	10.62
COM (days)	205.5	201.9	222.1	178.3
DFM (%)	16.8	16.2	16.7	19.6
(b) 2015				
Temperature (°C)	−0.54 ($+2.0\sigma$)	0.88 ($+2.4\sigma$)	−2.3 ($+1.8\sigma$)	1.7 ($+2.0\sigma$)
Precipitation (mm d^{-1})	8.3 ($+0.0\sigma$)	7.8 ($+0.0\sigma$)	7.7 ($+0.2\sigma$)	12.6 (-0.6σ)
SWE (mm)	492.4 (-1.5σ)	290.7 (-1.8σ)	712.7 (-0.9σ)	298.7 (-1.8σ)
Total Runoff (km^3)	49.11 (-0.5σ)	17.92 (0.5σ)	23.30 ($+0.1\sigma$)	7.89 (1.3σ)
COM (days)	168.9 (-3.1σ)	158.3 (-3.3σ)	188.4 (-2.7σ)	141.3 (-2.6σ)
DFM (%)	13.9 (-1.8σ)	13.4 (-1.6σ)	13.7 (-1.7σ)	16.4 (-1.7σ)
(c) 2040–2069				
Temperature (°C)	−0.4 ($+2.2\sigma$)	0.7 ($+2.1\sigma$)	−1.9 ($+2.2\sigma$)	2.2 ($+2.6\sigma$)
Precipitation (mm d^{-1})	9.0 ($+0.4\sigma$)	8.3 ($+0.3\sigma$)	7.9 ($+0.4\sigma$)	15.9 ($+0.4\sigma$)
SWE (mm)	565.5 (-1.1σ)	399.0 (-1.3σ)	715.7 (-0.9σ)	540.0 (-1.1σ)
Total Runoff (km^3)	50.3 (-0.4σ)	17.9 (-0.5σ)	22.2 (-0.3σ)	10.2 (-0.2σ)
COM (days)	182.2 (-2.0σ)	176.7 (-1.9σ)	200.4 (-1.8σ)	155.9 (-1.6σ)
DFM (%)	15.7 (-0.7σ)	15.1 (-0.6σ)	15.5 (-0.7σ)	18.6 (-0.5σ)

and maximum), precipitation, and specific humidity from the Livneh *et al* data set and future climate simulations, and calculated relative humidity (minimum and maximum) from the MTCLIM algorithm (Bohn *et al* 2013).

3. Results and discussion

3.1. Historical perspective of the 2015 drought

2015 was an exceptional (although not unprecedented) year in terms of April 1 SWE and winter temperature, while not remarkable in terms of precipitation (figure 2, table 1). Averaged across the Olympic and Cascade mountain ranges, April 1 SWE in 2015 was 490 mm, or −330 mm (-1.5σ) lower than the 1950–2015 climatology (the fifth lowest year since 1950). Average winter (November–March) temperature was the second highest at −0.54°C, or +2.1°C ($+2.0\sigma$) warmer than the climatology. By contrast, winter (November–March) precipitation was 8.3 mm d^{-1}, which is nearly equal to the long-term average. For comparison, for the period of record from 1950–2015, the lowest April 1 SWE was in 2005 with a −360 mm (-1.6σ) anomaly, the lowest average winter precipitation anomaly was 2001 at −4.0 mm day^{-1} (-2.4σ) anomaly, and the highest winter temperature was in 1992 with a +2.3°C ($+2.3\sigma$) anomaly.

The temperature, precipitation, and SWE anomalies varied strongly across our three regional domains in 2015 (table 1, figure S1). As shown in figure 3, 2015 had the lowest April 1 SWE since 1950 in the Olympics at −600 mm less than the climatology (-1.8σ). The average winter temperature in the Olympics was the

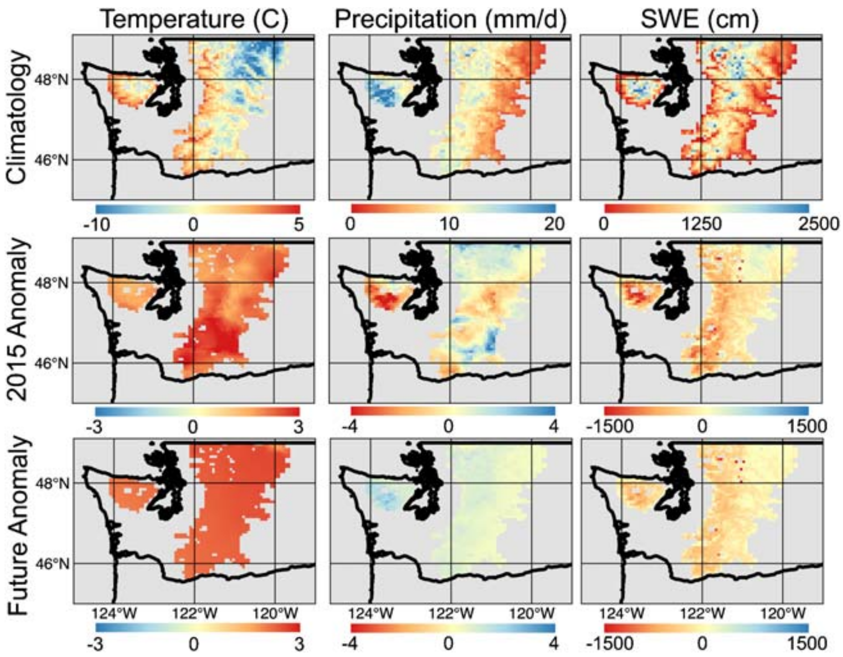

Figure 3. 1950–2015 climatology of winter (November–March) temperature (°C), winter (November–March) precipitation rate (mm day^{-1}), and April 1 snow water equivalent (SWE, mm day^{-1}) (top panel), 2015 anomaly relative to climatology (middle panel), and future anomaly (2040–2069) from the midrange emissions scenario relative to climatology (bottom panel). Note the common color bar for absolute values and anomalies.

second highest at +1.5 °C warmer than the climatology (+2.0σ) but average winter precipitation was not extremely low (the 18th lowest within 1950–2015 at −2 mm day^{-1} lower than the climatology). In the South Cascades, April 1 SWE was the second lowest on record at −390 mm (−1.8σ) lower than the climatology, winter temperature was the warmest on record (+2.4 °C (+2.4σ) above the climatology), and the average daily winter precipitation rate was close to the long-term average. The North Cascades experienced their second warmest winter on record at +1.9 °C (+1.8σ) warmer than the climatology, but April 1 SWE and precipitation were not extreme (13th and 40th over 1950–2015, respectively).

Total runoff from January–July 2015 was −4.3 km^3 lower (−0.5σ) than the climatology across the region, with the largest changes in the Olympics, with a −7.9 km^3 (−1.3σ) anomaly (table 1). The mean timing of the center of mass of total runoff (calculated as by Stewart *et al* (2005)) for the 2015 water year (October to September) had even larger changes. Averaged across the region, the center of mass was 37 d earlier (−3.1σ) in 2015 than the climatology. This anomaly was largest in the South Cascades, which was 44 d earlier (−3.3σ) in 2015 relative to the climatology.

Summer (June–September) fire danger during 2015 was exceptional as viewed through the lens of both 100 and 1000 hour DFM; 2015 had the fifth lowest DFM values since 1950. (100 and 1000 hour DFM showed similar spatial and temporal patterns so we only show the 100 hour DFM results). Averaged across our study area, DFM values were lower (corresponding to higher fire risk) in 2015 by −2.9% (−1.8σ) relative to the climatology (table 1, figure S3). In 2015, the largest DFM declines relative to the climatology were in the Olympics (−3.2%) versus −3.0% and −2.7% for the North and South Cascades, respectively.

3.2. Future projections compared to the 2015 drought

Averaged over 2040–2069, the multi-model mean from 10 downscaled CMIP5 models (table S1) for average winter temperature, winter precipitation, and April 1 SWE over our entire domain are −0.4 °C, 9.0 mm d^{-1}, and 570 mm (table 1). This corresponds to warmer winter temperatures by +2.3 °C (+2.2σ), (slightly) increased average winter precipitation by +0.7 mm day^{-1} (+0.4σ), and lower April 1 SWE by −250 mm (−1.1σ) relative to the 1950–2015 climatology. Using 2015 as a benchmark, future temperatures are comparable to 2015 but with more homogenous warming across the region (figure 3). Winter precipitation is higher in the future simulations than in 2015 (+0.7 mm d^{-1} (+0.7σ) vs. −0.1 mm d^{-1} (−0.03σ) anomalies relative to the climatology), with the largest difference in the Olympics, where the future precipitation anomaly is +1.4 mm (+0.4σ) vs. −1.9 mm (−0.6σ) for 2015. The April 1 SWE anomaly in 2015 was lower than the future (−330 mm vs. −250 mm relative to the climatology). The largest difference in April 1 SWE between 2015 and the 2040–2069 mean was in the Olympics, which had a −600 mm (−1.8σ) anomaly in 2015 vs. −360 mm (−1.1σ) in the future.

In addition to comparing mean future conditions, we also analyzed conditions in 2015 relative to

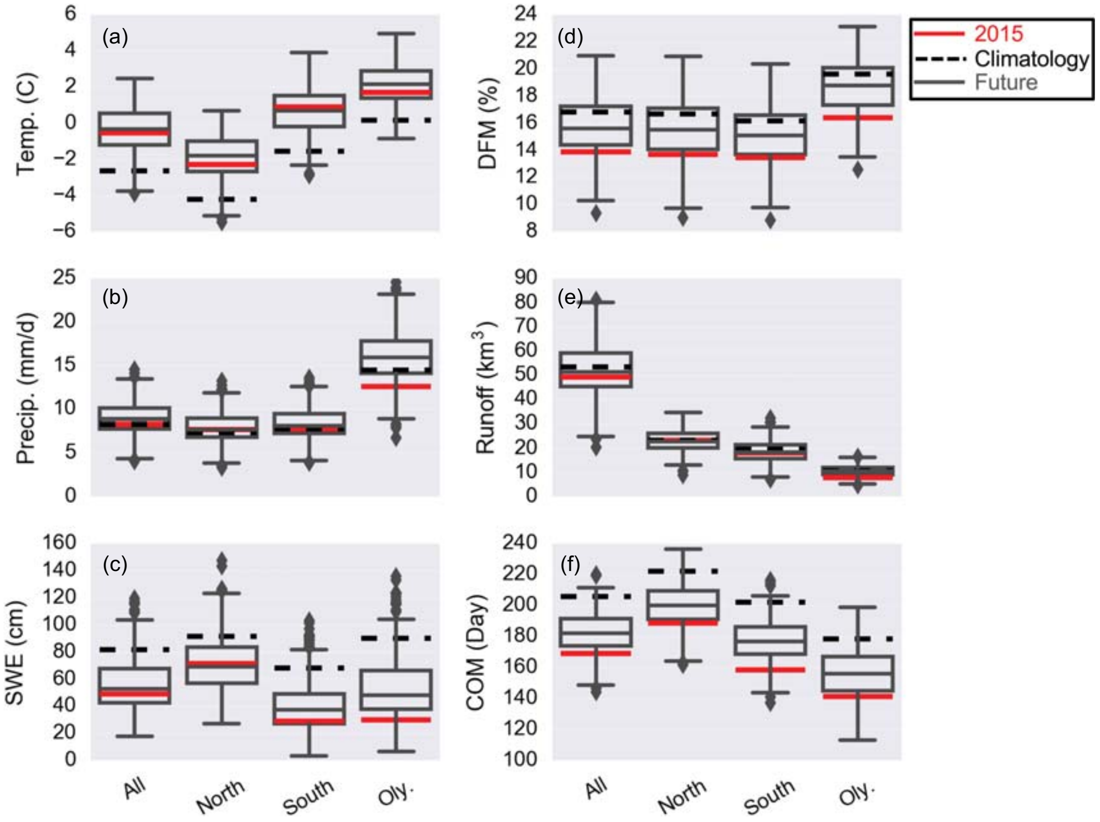

Figure 4. Distribution of all individual models and years over 2040–2069 (gray line) compared to the 1950–2015 climatology (black dotted line) and 2015 (red line). Calculated for winter (November–March) temperature (°C) (*a*), winter (November–March) precipitation (mm day^{-1}) (*b*), April 1 snow water equivalent (SWE, mm) (*c*), average summer (June–September) dead fuel moisture (DFM) content (*d*), January to July total runoff (km^3) (*e*), center of mass (COM) for total runoff (Julian days) (*f*). The whiskers represent 1.5 times the interquartile range.

individual years and models from the 2040–2069 ensemble (10 models × 30 (29) years (water years)). Individual future years show considerable variability, but the core areas of predicted change are consistent with the 2040–2069 mean (figure S2). Winter temperatures in 2015 were generally lower in the Olympics and North Cascades than future years, while winter temperatures in 2015 were slightly warmer than future years in the South Cascades. Winter precipitation in 2015 was generally higher in future years in the North Cascades and Southeastern Cascades and lower elsewhere (blue shades). SWE in 2015 was generally lower in the Olympics and South Cascades, and somewhat higher in the North Cascades relative to future years. Comparing with all future years from the ten individual models over the region as a whole (figure 4), winter temperature, winter precipitation, and April 1 SWE in 2015 ranked in the middle tercile of the distribution (41, 36, and 39 percentiles, respectively), only slightly below the mean of the future regional distribution. That said, 2015 winter precipitation and April 1 SWE were considerably lower in the Olympics relative to the individual future models and years.

Total January to July runoff in 2015 was close to the future mean for all years and models for the region as a whole, but 2015 was lower than the future mean in the Olympics. The mean timing of the center of mass of total runoff for each future water year (October to September) was 23 d earlier (−2.0σ) in the future relative to the climatology, primarily reflecting warmer future conditions. Future fire risk increased as estimated by declining summer 100 hr DFM, −1.1% (−0.7σ) relative to the climatology. Projected future declines were slightly less in the Olympics (−1.0%) than in the North and South Cascades (−1.2% and −1.1%, respectively). In the analysis reported above, we found that the advancement in the center of timing of runoff and fire risk in 2015 both exceeded changes projected for average conditions by the mid-21st century. However, conditions comparable to those observed in 2015 in terms of runoff timing and fire risk become more probable by the mid-21st century, in the 14th and 19th percentile, respectively, of the future distribution.

4. Conclusions

The 2015 drought in Washington State had widespread implications for water resources and fire activity. Using

2015 conditions as a benchmark for assessing modeled future changes in regional climate for the mid-21st century, we conclude that:

1. Across the Cascade and Olympic mountain ranges, 2015 had some of the lowest spring SWE and warmest winter temperatures on record, but precipitation was not much different from the climatology. When standardized relative to the climatology, 2015 winter temperatures were the warmest in the Southern Cascades and spring SWE was the lowest in the Southern Cascades and Olympics.
2. Averaged across the domain, winter temperatures in 2015 were similar to those projected in the future. In part as a result of projected increases in future winter precipitation, 2015 spring SWE reductions were larger than the future. As assessed by the DFM, fire risk was higher in 2015 than the mid-21st century than in 2015, but both are higher than the climatology.
3. Furthermore, our results (see figure S1) show that the drought conditions of 2015 occurred despite near-normal winter precipitation. Given the general similarity between 2015 and projected future conditions, this suggests a transition from precipitation to temperature control in future droughts. For instance, aside from 2015, figure S1 shows that low snowpacks associated with most previous droughts have occurred primarily as a result of anomalously low winter precipitation (e.g. 1977, 2001, 2005) which is not the case for 2015, or for projected future conditions.

While we demonstrate that 2015 may be representative of expected conditions in the mid-21st century, there are additional factors to consider. First, the impact of chronically warmer conditions, in contrast to an isolated warm year, could amplify the impacts on hydrology and natural resources, and certainly if 2015 is taken as representative of average conditions, many years would be drier (particularly summers). Furthermore, changes in average conditions may exacerbate summer dryness due to interannual carryover effects. For example, in the Central Great Plains of the US, soil moisture deficits are more severe during consecutive years of drought than an isolated year of extreme drought (Livneh and Hoerling 2016). The extent of these impacts will likely differ for managed systems like agriculture (Peck and Adams 2010) as contrasted with natural ecosystems, which have varying capacity for resiliency to multi-year droughts (Asner *et al* 2016). Second, projections of increased primary production from CO_2 fertilization, coupled with decreased water availability (Mankin *et al* 2017), could further increase the area susceptible to future fires, especially due to enhanced seasonal understory and/or grassland growth. Further examinations of past drought conditions, such as 2015, can provide the context to better anticipate and plan for these future changes.

Acknowledgments

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