The Changing Character of Twenty-First-Century Precipitation over the Western United States in the Variable-Resolution CESM

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ABSTRACT

The changing characters of precipitation frequency and intensity have been comprehensively investigated from the recent historical period to the end of the twenty-first century over the western United States. Variable-resolution Community Earth System Model (VR-CESM) ensemble simulations are applied with a fine grid resolution of $\sim 0.25^\circ$ over the study area. Simulations are forced with prescribed sea surface temperatures, sea ice extent, and greenhouse gas concentrations from the representative concentration pathway 8.5 (RCP8.5) scenario. VR-CESM is shown to be effective at accurately capturing the spatial patterns of the historical precipitation climatology. The results of VR-CESM output provide significantly regional details with crucial enhancement of precipitation representations over complex terrain. In the Intermountain West and U.S. Southwest, a statistically significant increase in mean precipitation and rainy days through midcentury is observed, although this trend is tempered by the end of the century in response to a decrease in relative humidity. Over the Pacific Northwest, extreme precipitation events are observed to increase significantly as a result of increased cool season integrated vapor transport associated with a moistening of the cool seasons and drying through the warm seasons. In particular, extreme precipitation in this region appears to increase more rapidly than would be predicted by the Clausius–Clapeyron relationship. No clear climate signal emerges in mean precipitation or extreme events in the majority of California, where the precipitation climatology is attributed to large interannual variabilities that are tied closely to ENSO patterns.

1. Introduction

There is a substantial and growing interest in understanding the character of precipitation within a changing climate, motivated largely by its pronounced impacts on water availability and flood management in both human and natural systems (Hegerl et al. 2004; Kharin et al. 2007; Scoccimarro et al. 2013). Among past studies addressing precipitation, extremes have been a major focus, particularly drought and flood events (Seneviratne et al. 2012).

Overall, it is widely agreed that although atmospheric water vapor concentration is increasing, the impacts of a changing climate on the character of precipitation are far more complicated. Extreme precipitation events are particularly nuanced: Our best projections suggest that precipitation extremes will intensify even in regions where mean precipitation decreases (Tebaldi et al. 2006; Kharin et al. 2007).

Climate projections, particularly those addressing the frequency and intensity of rare events, are inevitably subject to large uncertainties. Nonetheless, climate models have been invaluable tools for developing insight into this problem (Easterling et al. 2000). In particular, global climate models (GCMs) have often been used to investigate changes in the mean, variability, and extremes of climate, as forced with predicted greenhouse gas (GHG) concentrations and aerosol emissions (Meehl et al. 2006). For example, in the context of globally averaged projections, Kharin et al. (2013) found that changes in extreme precipitation were amplified (by 2–3 times) compared to mean precipitation in the models in phase 5 of the Coupled Model Intercomparison Project (CMIP5). Generally, models show better agreement on the response of extreme

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precipitation in the extratropics, where large-scale precipitation is more prevalent, than in the tropics and subtropics (Kharin et al. 2013; Pendergrass et al. 2015).

Although quite a few studies have investigated climate extremes at the global scale (Seneviratne et al. 2012), efforts addressing extremes at local and regional scales are less common. It is well understood how increased GHG concentrations have contributed to the observed intensification of heavy precipitation events over the tropical ocean (Allan and Soden 2008) and the majority of Northern Hemisphere overland areas (Min et al. 2011), but changes are much more poorly understood at regional scales where meteorological variability is large (Trenberth 2011). Moreover, GCMs tend to underestimate observed precipitation magnitudes both in the aspects of mean precipitation and precipitation extremes, implying the necessity of downscaling techniques (Sillmann et al. 2013a). This issue of insufficient regional-scale climate information has been a major outstanding problem in climate science, as stakeholders and water managers typically require finescale information on climate impacts in order to effectively develop adaptation and mitigation strategies.

As a consequence, confidence in the assessment of regional extreme precipitation changes requires both high spatial resolution and a long integration period, both of which can make the computational cost untenable for global simulations. Dynamical downscaling with regional climate models (RCMs) has been one of the few tools available to achieve the frequency, intensity, and duration of extreme events at the needed scales. By only simulating a limited regional domain, RCMs better capture finescale dynamical features with high horizontal resolution (Bell et al. 2004; Frei et al. 2006; Rauscher et al. 2010; Wehner 2013). Finer resolution enables more accurate simulation of precipitation extremes, with better-resolved circulation patterns, cloud distributions, land use features, snowpack, and orographic effects (Leung et al. 2003a; Diffenbaugh et al. 2005; Salathé et al. 2008; Wehner et al. 2010). For example, Leung et al. (2003b) showed that the higher-resolution RCMs yield more realistic precipitation patterns and produce more frequent heavy precipitation over the western United States (WUS), consistent with observations. Diffenbaugh et al. (2005) studied both extreme temperature and precipitation events over the contiguous United States using an RCM configured at 25-km horizontal resolution, and demonstrated that finescale processes were critical for accurate assessment of local- and regional-scale climate change vulnerability. In the study of Ashfaq et al. (2016), a 7.4% increase in precipitation from extremes is projected with every 1°C rise in surface temperature over the contiguous United States using RegCM4 simulations driven by CMIP5 data, when comparing the historical period (1965–2005) to the near-term future period (2010–2050).

However, RCMs also have known issues associated with inconsistency between the lateral forcing data and the driven RCM. The menu of physical parameterizations and tuning parameters typically available to RCMs can also lead to overrunning of the model for a particular geographic region or climatological field (McDonald 2003; Laprise et al. 2008; Mesinger and Veljovic 2013). Consequently, there has been growing interest in utilizing variable-resolution-enabled GCMs (VRGCMs) to perform regional climate simulations. Unlike RCMs, which require GCM data to drive the simulation at lateral boundaries, VRGCMs use a unified model with enhanced resolution over a specific study region while keeping the coarse resolution at remaining global area (Staniforth and Mitchell 1978; Fox-Rabinovitz et al. 1997). VRGCMs have demonstrated a competitive ability for regional climate studies at a reduced computational cost, particular when compared to uniform-resolution GCMs (Fox-Rabinovitz et al. 2006; Rauscher et al. 2013).

In this paper, we utilize the recently developed variable-resolution option in the Community Earth System Model (VR-CESM). VR-CESM is based on the CESM [and its predecessor, the Community Climate System Model (CCSM)], a family of models that have been used for decades to study the global climate (Neale et al. 2010; Hurrell et al. 2013). The overall performance of VR-CESM for modeling regional climate in the California and Nevada is detailed in Huang et al. (2016), where it was argued that VR-CESM has competitive biases in comparison to the Weather Research and Forecasting (WRF) Model (a traditional RCM) and the uniform-resolution CESM, when evaluating against high-quality observations and reanalysis. VR-CESM has already been used in a number of studies to simulate finescale atmospheric processes (Zarzycki et al. 2014, 2015; Rhoades et al. 2016; Huang and Ullrich 2016; Rhoades et al. 2017).

This study focuses on changes in the character of precipitation over the twenty-first century within the WUS, as predicted from long-term ensemble runs conducted with VR-CESM under a local grid resolution of $\sim 0.25^\circ$. The WUS is known to be particularly vulnerable to hydrological extremes, particularly floods and droughts (Leung et al. 2003b; Caldwell 2010), and hosts a variety of local features and microclimates associated with its rough and varied topography. Simulations of the future climate are performed in accordance with the representative concentration pathway 8.5 (RCP8.5) scenario, which describes a “business-as-usual” projection for GHGs (Riahi et al. 2011). In this study, we focus singularly on the RCP8.5 scenario because its midcentury results are similar to the more optimistic RCP2.6 scenario end-of-century results. Global sea surface temperatures (SSTs) and sea ice, which are

used to compute ocean–atmosphere fluxes, are prescribed in accordance with the widely used Atmospheric Model Intercomparison Project (AMIP) protocol (Gates 1992). By constraining atmospheric boundary conditions at the sea surface, we avoid model biases in SSTs that are known to be significant in the fully coupled configuration (Joseph and Nigam 2006; Grodsky et al. 2012; Stevenson 2012; Jha et al. 2014; Small et al. 2014). Of course, this choice requires us to accept potential uncertainties associated with our choice of SSTs.

Changes in the character of precipitation, in terms of frequency and intensity at multiple levels, have been assessed in our study from recent history through the end of the twenty-first century. A comprehensive set of indices for precipitation extremes has been evaluated from the ensemble simulations over the 26-yr periods corresponding to historical (1980–2005), midcentury (2025–50), and end-of-century (2075–2100) periods. Spatial inhomogeneity in local geography and temperature are observed to result in similarly inhomogeneous impacts on the precipitation field. Teleconnections [specifically El Niño–Southern Oscillation (ENSO)] are also observed to have a pronounced impact on precipitation features. Since only one SST dataset was used for this study, we note that our projections are conditioned on a particular future character of ENSO. This is a potentially large source of uncertainty, as at present there is no clear consensus on how ENSO may behave under a warming climate (Fedorov and Philander 2000; Latif and Keenlyside 2009; Guilyardi et al. 2009; Collins et al. 2010; DiNezio et al. 2012), and strengthening or weakening of this pattern will have clear consequences for our results (as discussed in section 6d).

This work builds on a number of previous studies that have explored the projected future change in WUS precipitation. For example, Kim (2005) applied downscaled climate change signals to selected indicators, and concluded that global warming induced by increased CO₂ is likely to drive increases in extreme hydrologic events in the WUS. Duffy et al. (2006) found that changes to mean precipitation predicted by the RCMs are not statistically significant compared to interannual variability in many regions over the WUS, although there is little consistency among the different RCMs as to responses in precipitation to increased GHGs. Gao et al. (2015) pointed out a potentially large increase in atmospheric river events by the end of the twenty-first century under the RCP8.5 scenario, with implications for large-scale and heavy precipitation events along the Pacific coast.

This paper is structured as follows. Section 2 describes the model setup. Section 3 describes the methodology and reference datasets employed. An assessment of the ability of the model to capture the historical climatology

of the WUS is given in section 4 with discussions of drivers of precipitation change in section 5. Results from the future mean climatological trend and projected changes to precipitation indices are in section 6. Section 7 summarizes the main points of the study along with a further discussion.

2. Model setup

CESM is a state-of-the-art Earth modeling framework, consisting of coupled atmosphere, ocean, land, and sea ice models (Neale et al. 2010; Hurrell et al. 2013). In this study, the Community Atmosphere Model, version 5 (CAM5) (Neale et al. 2010), and the Community Land Model, version 4.0 (CLM4) (Oleson et al. 2010), are used. CAM5 is configured with the spectral element (SE) dynamical core, which is known for its conservation properties, accuracy, and parallel scalability (Dennis et al. 2011; Taylor 2011) and incorporates the variable-resolution option (Zarzycki et al. 2014). CLM4 is employed in the “ungrid” configuration, which allows the land model and atmospheric model to utilize the same model grid and so eliminates the need for interpolation. The variable-resolution mesh used for this study is depicted in Fig. 1, in accord with our past studies (Rhoades et al. 2016; Huang et al. 2016; Huang and Ullrich 2016; Rhoades et al. 2017).

Simulations have been performed for the historical period (1979–2005; referred to as hist) and for two future periods: 2024–50 (mid) and 2074–2100 (end). Daily outputs are recorded for each period on the native SE grid and then remapped to a regional latitude–longitude mesh (Ullrich and Taylor 2015; Ullrich et al. 2016). The first year of each time period was discarded as a spinup period to allow adequate time for the initialized land and atmosphere to equilibrate. The 26-yr duration was chosen to provide an adequate sampling of annual variability for each time phase. As mentioned earlier, GHG concentrations are set based on RCP8.5. Historical SSTs and sea ice are prescribed at 1° resolution, as described by Hurrell et al. (2008). SSTs and sea ice for each future period are developed from fully coupled RCP8.5 climate simulations from CESM with bias correction applied (C. Hannay 2015, personal communication). Annually updated land surface datasets, which prescribe land-use characteristics, are interpolated from 0.5° to the land model grid.

Ensemble runs are needed to ensure that the sample adequately accounts for climate variability, especially for statistics associated with climatological extremes. However, the exact number of required ensemble members is heavily dependent on the variability of the

FIG. 1. (a) The approximate grid spacing used for the VR-CESM 0.25° mesh. (b) A depiction of the transition from the global 1° resolution mesh through two layers of refinement to 0.25°. (c) Topography height over the study area.

particular metric being examined, and so no standard ensemble criteria exist. Deser et al. (2012b) suggest that around three ensemble runs are required to detect a significant epoch difference for June–August (JJA) surface temperatures, whereas 10–30 ensemble members are needed for that for December–February (DJF) precipitation. In our study, the use of a single modeling system and prescribed SSTs does reduce the intrinsic variability of the climate system and so we found reasonably converged results with two ensemble members for the historical period and four ensemble members for each future period (see Figs. S1–S3 in the supplemental material for the intermember variability). Although a multimodel mean is often preferable for capturing climate variability, we also note that accuracy may be impacted by compensating results from models of differing quality. Further, since our study focuses primarily on understanding the drivers of changing extreme weather, the changes that emerge from a single model are indicative of the physical relationships that will hold in the coming century.

3. Methodology

a. Precipitation indices

To choose a comprehensive (but minimal) set that is informative to stakeholders and water resource managers, indices from throughout the literature were compiled to characterize precipitation (Tebaldi et al. 2006; Zhang et al. 2011; Sillmann et al. 2013b). The indices examined include those defined by the Expert Team on

Climate Change Detection and Indices (ETCCDI) (Karl et al. 1999) that are featured in earlier studies (Diffenbaugh et al. 2005; Dulière et al. 2011; Sillmann et al. 2013b; Singh et al. 2013) and others such as return levels and dry spell and wet spell characteristics defined by either percentiles or by selected thresholds.

From our standpoint, binning precipitation events by intensity is more intuitive than percentile, more relevant for societal impacts (Alexander et al. 2006), and more informative to water resources managers and climate adaptation strategies. Therefore, the indices we have chosen for this study attempt to provide a relatively comprehensive characterization of precipitation, and are summarized in Table 1. Indices related to dry spells of variable duration were also investigated in this study, but they only exhibited significant differences for extremely short (≤ 5 days) dry spells and so are not included in our results.

b. Impacts of ENSO

The impact of ENSO is emphasized in our study because of its significant influence on precipitation over a majority of our study area, particularly the U.S. Southwest (Cayan et al. 1999; Zhang et al. 2010; Deser et al. 2012a; Yoon et al. 2015). The phase of ENSO (i.e., El Niño and La Niña) is identified each year using the oceanic Niño index (ONI), defined as the 3-month running means of SST anomalies in the Niño-3.4 region [covering 5°N–5°S, 120°–170°W, based on NOAA (2013)]. An El Niño or La Niña episode is said to occur when the ONI exceeds +0.5 or −0.5 for at least five consecutive months for a water year (i.e., from July to June) (NOAA

TABLE 1. Precipitation indices employed in this study.

Index	Definition
Pr	Mean daily precipitation
R1mm	Number of days per year with $Pr > 1$ mm
SDII	Simple precipitation intensity index: Precipitation amount/R1mm
R5mm	Number of days per year with $Pr > 1$ mm and $Pr \leq 5$ mm
R10mm	Number of days per year with $Pr > 5$ mm and $Pr \leq 10$ mm
R20mm	Number of days per year with $Pr > 10$ mm and $Pr \leq 20$ mm
R40mm	Number of days per year with $Pr > 20$ mm and $Pr \leq 40$ mm
Rxmm	Number of days per year with $Pr > 40$ mm
F1mm	Fraction of precipitation contributed to the total precipitation for days of R1mm (similarly for F5mm, F10mm, F20mm, F40mm, and Fxmm)
P5mm	Precipitation amount from R5mm (similarly for P10mm, P20mm, P40mm, and Pxmm)

2013) (see Fig. S4 in the supplemental material for the ONI values). To adjust for the trend in the SST field associated with climate change, the anomaly is computed against the detrended mean SSTs from the periods 2020–50 and 2070–2100 for the mid and end periods respectively, using the aforementioned predicted SST dataset. As argued by Kao and Yu (2009), it may be desirable to use an extended Niño-3.4 region to determine the phase of ENSO; however, when employing SST anomalies integrated over an extended region of 105° – 170° W, we observed no significant impact on ONI statistics.

c. Assessing statistical significance

The Student's t test has been used to determine whether or not two datasets at each grid point are statistically equivalent when the sample population can be adequately described by a normal distribution. The normality of a dataset is assessed under the Anderson–Darling test. When the sample populations do not approximately follow a normal distribution, the Mann–Whitney–Wilcoxon (MWW) test is employed in lieu of the Student's t test. All tests are evaluated at the 0.05 (α) significance level. When comparing different time periods, statistical tests are conducted by treating all years from all ensemble members as independent samples (26×2 sample years for hist and 26×4 sample years for mid and end). Spatial correlation is assessed by computing the Pearson product-moment coefficient of linear correlation between climatological means from models and reference datasets.

d. Reference datasets

Gridded observational datasets and reanalysis of the highest available quality, with comparable horizontal resolutions to our VR-CESM simulations, are used for assessing the simulation quality. For comparison purposes, the reference dataset is interpolated to the model grid as needed, using bilinear interpolation methods. Multiple reference datasets are necessary because of the

underlying uncertainty in the precipitation field. The three datasets employed are described below.

1) UW GRIDDED DATA

The 0.125° UW daily gridded meteorological data are obtained from the Surface Water Modeling group at the University of Washington (UW), covering the period 1949–2010 (Maurer et al. 2002; Hamlet and Lettenmaier 2005). The UW dataset imposes topographic corrections by forcing the long-term average precipitation to match that of the Parameter-Elevation Regressions on Independent Slopes Model (PRISM) dataset.

2) NCEP CPC DATASET

The 0.25° National Centers for Environmental Prediction (NCEP) Climate Prediction Center (CPC) daily dataset provides gauge-based analysis of daily precipitation covering the period 1948–2006. It is a unified precipitation product that covers the conterminous United States and amalgamates a number of data sources at CPC via optimal interpolation objective analysis.

3) NARR

The ~ 32 -km NCEP North American Regional Reanalysis (NARR) provides 3-hourly precipitation snapshots, obtained by dynamical downscaling over North America and covering the period 1979 to the present (Mesinger et al. 2006).

4. Assessment of precipitation character in VR-CESM

Before proceeding, we assess the ability of VR-CESM to represent the character of precipitation over the WUS. The indices defined in Table 1 are depicted in Figs. 2, 3, and 4 for VR-CESM and each of the reference datasets over the historical period (1980–2005). We

FIG. 2. (left)–(right) Mean precipitation and indices SDII, R1mm, and F1mm from (top)–(bottom) VR-CESM and reference datasets over the historical period, 1980–2005. Areas with statistically significance differences are marked with black stippling.

assume equal confidence in each of the reference datasets, and use the Student's t test (with UW, CPC, and NARR as the three statistical samples) to identify regions where VR-CESM deviates significantly from the reference mean. Regions where differences are statistically significant in the VR-CESM dataset are identified

with stippling. It is widely acknowledged that NARR is not as good at representing precipitation climatology as the gridded observations of CPC or UW (Bukovsky and Karoly 2007; Huang et al. 2016). However, the differences between NARR and gridded products also tend to be greatest in regions of high observational uncertainty,

FIG. 3. As in Fig. 2, but for indices (left)–(right) R4mm, F5mm, R10mm, and F10mm.

and so its inclusion is useful for quantifying the performance of VR-CESM against this uncertainty.

Overall, VR-CESM accurately captures the spatial patterns of precipitation (with spatial correlation coefficients larger than 0.9 as compared to the observations) and its indices. As expected, the majority of precipitation is distributed along the northwest coast and the mountainous regions

of the Cascades and the Sierra Nevada. Specially, regional biases are present with detailed descriptions as follows.

First, VR-CESM significantly overestimates precipitation (Pr) over dry regions with differences between 0.2 and 1.5 mm, and over the eastern flank of the Cascades and on both sides of the Sierra Nevada (with relative differences reaching 50%–150%). As with many

FIG. 4. As in Fig. 2, but for indices (left)–(right) R20mm, F20mm, R40mm, F40mm, Rxmm, and Fxmm.

regional models, VR-CESM is “dreary” and exhibits too many precipitation days ($R1mm$, $Pr \geq 1 \text{ mm day}^{-1}$, and $R5mm$, $1 \text{ mm day}^{-1} \leq Pr \leq 5 \text{ mm day}^{-1}$) (see Figs. 2 and 3) (Stephens et al. 2010). The spatial correlations are about 0.85, 0.75, 0.8, and 0.9 for $R1mm$, $R5mm$, $R10mm$, and $R20mm$ compared to the references. Nonetheless, over most regions, the relative contribution of each precipitation frequency subset to total precipitation (including $F1mm$, $F5mm$, $F10mm$, and $F40mm$) is fairly well represented by the model (with spatial correlations ranging from 0.7 to 0.8), suggesting that the model is effective at capturing the overall frequency distribution of the precipitation intensity.

Second, the spatial pattern of precipitation intensity (SDII) matches well between VR-CESM and references (with spatial correlations around 0.85) everywhere except in the Great Plains (the eastern edge of our domain) and in California’s Central Valley. The Great Plains is not a focus of this study and the suppressed intensity is primarily during the warm season

(April–September), which likely represents a failure of the convection scheme to adequately simulate variability in this region. This bias is also observed in CESM at 1° and 0.25° uniform-resolution CESM simulations (Small et al. 2014) and so is not a symptom of the variable-resolution transition region over the eastern edge.

However, the grossly exaggerated intensity over the western flank of the Sierra Nevada through California’s Central Valley does merit some additional discussion. Here, the overestimation of precipitation and enhanced intensity is associated with too many extreme precipitation events ($Pr > 20 \text{ mm day}^{-1}$) (see Fig. 4 for $R40mm$ and $Rxmm$). This bias is related to exaggerated orographic uplift (upslope winds), which also triggers a dry bias along the eastern flank of the Sierras. Similar biases in simulating extreme precipitation over topographically complex regions have also been found in high-resolution RCM simulations, and have been primarily attributed to excessively strong winds (Walker and

Diffenbaugh 2009; Singh et al. 2013). This bias is not alleviated by simply increasing the spatial resolution, as experimental VR-CESM simulations at 14, 7, and 3.5 km show only modest improvement (A. M. Rhoades 2016, personal communication). For VR-CESM, this issue may be further impacted by the diagnostic treatment of precipitation in CAM5 (Morrison and Gettelman 2008; Gettelman et al. 2008). Recent work by Rhoades (see personal communication above) has shown that using prognostic Morrison–Gettelman (MG2) microphysics routines can dramatically improve mountain precipitation climatology, as tested with an early version of CAM6. The bias might also be related to more complex dynamic processes or the lack of the scale-aware model parameterization schemes when producing the orographic forced precipitation.

The representation of precipitation in VR-CESM over California was also discussed in Huang et al. (2016), where it was observed that VR-CESM simulations at 0.25° adequately represented regional climatological patterns with high spatial correlation. VR-CESM demonstrated comparable performance to WRF with a similar grid resolution of 27 km (which was forced by ERA-Interim reanalysis), showing significant improvement contrasted to CESM at ~1°.

To better understand the impacts of resolution, the precipitation climatology of CESM at ~1° resolution is also assessed in the supplemental material (see Fig. S5 therein). Overall, we find that precipitation patterns over complex topography are poorly represented in the 1° dataset and do not capture the spatial patterns induced by orographic effects. Over the Cascades and the Sierra Nevada, total precipitation is grossly underestimated by the 1° data compared to gridded and reanalysis datasets. Precipitation has otherwise been smoothed out over the coastal areas and the mountainous regions of the northwest United States when simulated with CESM at coarse resolution. This result clearly underscores the benefits of high resolution (particularly the representation of topography) in simulating precipitation features. The ability for GCMs to simulate extreme precipitation also strongly depends on the horizontal resolution as discussed in Wehner et al. (2010) since precipitation typically intensifies at high resolution (Rauscher et al. 2016; O'Brien et al. 2016). A rough comparison is also provided (also see Fig. S5) for the output from a globally uniform CESM run at 0.25° spatial resolution with the finite-volume (FV) dynamical core (Wehner et al. 2014), exhibiting similar performance to VR-CESM. Overall, VR-CESM at 0.25° resolution appears to provide the best trade-off between accuracy and computational cost, as the coarser resolution does not correctly represent precipitation features and higher resolution does not appear to substantially

improve model accuracy (at least in this version of CAM).

We have also assessed the impact of the ENSO within the historical VR-CESM runs by differencing the precipitation fields between the warm phase (i.e., El Niño) and cool phase (i.e., La Niña), compared to references (see Fig. S6 in the supplemental material). Notably, ENSO exhibits a weaker signal for observational precipitation, compared to VR-CESM, which might suggest that the model exaggerates ENSO's impact on precipitation, especially over the U.S. Northwest. Therefore, the improvement of ENSO in the model is directly proportional to the representation of ENSO-forced precipitation anomalies (AchutaRao and Sperber 2006).

5. Drivers of climatological precipitation change

The remainder of this paper now focuses on model predictions of precipitation change over the twenty-first century. Precipitation has been observed and modeled to be changed at both global and regional scales under climate change. The observed intensification of heavy precipitation events over the recent past for the majority of Northern Hemisphere land areas is primarily attributed to increases in GHGs (Min et al. 2011). GHGs drive radiative changes in the lower troposphere, increase SSTs, and lead to increased evaporation, all of which then impact the character of precipitation events (Allen and Ingram 2002; Sugi and Yoshimura 2004). Several studies have argued that precipitation extremes will intensify continuously through the end of the twenty-first century in both dry and wet regions, although the extent of this change will be spatially heterogeneous (Donat et al. 2016).

In accordance with the Clausius–Clapeyron (C–C) relationship, saturation vapor pressure in the atmosphere is expected to increase by ~7% for each 1°C increase in temperature (Allan and Soden 2008). As long as a source of water vapor is present, a corresponding increase in atmospheric water vapor content is expected. Naturally, evaporation over the ocean will intensify with climate warming, but increases in water vapor content over land may be constrained by soil moisture (Cayan et al. 2010). When specific humidity is high, heavy rain events become more probable, even if total precipitation is decreasing (Allen and Ingram 2002; Trenberth 2011). This suggests that global total precipitation is expected to increase at a slower rate than precipitation extremes (Allan and Soden 2008). In accordance with previous studies (e.g., Allan and Soden 2008; O'Gorman and Schneider 2009; Min et al. 2011), changes to extreme precipitation follow the C–C relationship more closely than total precipitation amount

FIG. 5. (top)–(bottom) The 2-m average temperature (Tavg) and relative humidity (RH), and mean precipitation (Pr) averaged over (left)–(right) the historical time period, along with average mid-hist and end-hist differences for October–March (first, third, and fifth columns) and April–September (second, fourth, and sixth columns). Areas with statistically significance differences are marked with black stippling.

(Trenberth et al. 2003). However, there is still substantial uncertainty regarding the magnitude of this change, since precipitation extremes are also dependent on factors such as the vertical velocity profile and thermodynamic effects (O’Gorman and Schneider 2009).

With overland water vapor constrained by soil moisture content, changes to moderate or heavy precipitation events (primarily during the cool season) over the WUS are mainly the result of increased large-scale vapor transport from the eastern Pacific Ocean rather than directly from evaporation, typically associated with atmospheric rivers (ARs) and/or orographic uplift (Trenberth et al. 2003; Neiman et al. 2008). Warming may lead to enhancement of the storm track, which would increase ARs along the U.S. West Coast with increased air water vapor content in the future (Dettinger 2011; Gao et al. 2015).

In addition, the precipitation of the WUS has strong interannual variability caused by large-scale atmospheric circulation mainly associated with ENSO (Leung et al. 2003b). As a significant driver of precipitation, ENSO modulates the storm-track behavior over the WUS with a northwest–southwest precipitation dipole (Gershunov and Barnett 1998), as discussed in section 6d. The projected SSTs used in this study emerge from one possible realization of ENSO. However, there is still substantial uncertainty regarding how El Niño will change under global warming (Fedorov and Philander 2000; Guilyardi et al. 2009), which is one of the main

sources of uncertainty in our results. Capotondi (2013) showed that the diversity of El Niño characteristics in CCSM4 is comparable to what was found in observations, although, as found by Deser et al. (2012c), the overall magnitude of ENSO in CCSM4 is overestimated by ~30% over the preindustrial time period.

6. Results

a. Mean climatology

Differences in the mean climate of the WUS, as predicted by VR-CESM across time periods, are depicted in Fig. 5. Since the character of WUS precipitation has a strong seasonal contrast, changes to mean precipitation (Pr), near-surface (2 m) temperature (Tavg), and near-surface relative humidity (RH) are depicted for what we refer to as the cool season (October–March) and the warm season (April–September).

As a result of enhanced GHG concentrations, mean annual Tavg increases by between 1.5 and 3.5 K from hist to mid (mid-hist) and between 4 and 7.5 K from hist to end (end-hist). Despite the large spatial variation in mean seasonal temperatures, the observed differences in mean temperature across time periods are fairly uniform, particularly over the ocean and in coastal regions. Away from the coast, there is a weak gradient in the temperature change field, with the largest increase in

temperatures occurring toward the northeast during the cool season and toward the north during the warm season. The increases in temperature are about 0.5 and 1.0 K larger during the warm season compared to the cool season for mid and end, respectively.

Overall, future RH is constrained closely to hist since it is governed by competing increases in temperature and atmospheric water vapor content. Although RH increases monotonically over the ocean in response to increased evaporation, over land the character is more heterogeneous: in general, RH tends to increase in regions where the T_{avg} increase is constrained below ~ 2 K but decrease when the T_{avg} anomaly exceeds ~ 2 K. The decrease in these regions is on the order of 2% and 3%–6% for mid and end, respectively. In fact, trends in RH are spatially consistent with temperature increase but opposite in magnitude with a spatial correlation coefficient of approximately 0.8. This suggests that continental evaporation and oceanic water vapor transport are insufficient vapor sources when the temperature reaches a certain level, consistent with the observation of Joshi et al. (2008). This effect has also been found in results by Rowell and Jones (2006) over continental and southeastern Europe, and by Simmons et al. (2010) over low-latitude and midlatitude land areas.

In response to these changes to T_{avg} and RH, from hist to mid, mean precipitation over the entire study domain exhibited a 0.2 – 0.6 mm day^{−1} increase during the cool season (about 10%). The largest changes were over the northwest, where cool-season precipitation emerges from large-scale patterns (i.e., atmospheric rivers and associated storm systems) (Trenberth et al. 2003; Neiman et al. 2008). Over the warm season, where precipitation in the WUS is primarily from convection, the increase was around 0.2 mm day^{−1} (about 10%) through the Intermountain West and U.S. Southwest with drying through the U.S. Northwest (a decrease in mean precipitation of 0.2 mm day^{−1}). These trends largely hold and intensify through end (with relative changes of about 20%–30% compared to hist), except in the Intermountain West and Southwest regions where precipitation again falls to historical levels. The statistical significance of these results is depicted in Fig. 6.

The increase in cool season precipitation in the U.S. Northwest is largely driven by increased integrated vapor transport (IVT) (see Fig. 7) during extreme precipitation events. As observed in previous studies, IVT is particularly useful for understanding extreme precipitation events that arise from large-scale meteorological features (Ralph et al. 2004; Leung and Qian 2009; Dettinger 2011). IVT is composed of column water vapor content and horizontal wind advection, both of which could be impacted by the climate change signal as plotted

in Fig. 7. Over the eastern Pacific, we observe increases in both water vapor content and wind speed, which are in turn responsible for increases to IVT in the Pacific Northwest. However, over the continent, we see a weakening of the westerlies overland driven by a reduced meridional temperature contrast. The increased cool-season IVT does not manifest strongly along the Pacific coast of California, where IVT is much smaller on average and is primarily modulated by ENSO.

Changes in precipitation over the Intermountain West and U.S. Southwest during the warm season are primarily associated with convective processes and so are directly impacted by variations in RH. As shown in Fig. 5, RH increases through midcentury in this region (although with modest significance) and then significantly decreases through the end of the century over most of the study area (except over where soil moisture was already low in hist). This results in a modest increase in precipitation through midcentury followed by a return to historical precipitation amounts by end of century. Further climate warming is expected to further decrease RH and drive increased aridity in this region.

b. Precipitation indices

This section now analyzes changes to precipitation character with respect to our predefined indices (as given in Table 1). For each index, the changes from past to future, as annually averaged over different time periods for all ensemble members, are plotted in Fig. 6 (for the indices that quantify precipitation days) and Fig. 8 (for the indices describing precipitated water amounts).

On comparing hist and mid, it is clear that the number of rainy days and frequency of nonextreme precipitation events (≤ 10 mm day^{−1}) has increased significantly (about 10%–15%) over the U.S. Southwest and Intermountain West, which is less obvious between mid and end. On the contrary, the frequency of nonextreme precipitation has decreased significantly over the U.S. Northwest region and the eastern areas of Montana, Wyoming, and Oregon (by about 10%). The increase in the frequency of these nonextreme precipitation events explains the observed change to mean precipitation exhibited in Fig. 5, which is largely associated with warm season mesoscale storm systems.

Although essentially all regions exhibit an increase in high-rain events ($Pr > 10$ mm day^{−1}), this increase is only statistically significant through the Intermountain West and in the Pacific Northwest (for $Pr > 20$ mm day^{−1}). When comparing mid to end, there is a clear and significant increase in extreme precipitation events over the U.S. Northwest coast [$\sim (20\%$ – $30\%)$] and eastern flank of the Cascades ($> 40\%$). The increase in the U.S. Northwest is accompanied by a decrease in nonextreme

FIG. 6. (left)–(right) Differences of precipitation indices (a) Pr (mm day^{-1}), R1mm, R5mm, and R10mm and (b) SDII, R20mm, R40mm, and Rxmm between (top)–(bottom) mid-hist, end-mid, and end-hist averages in (a) and (b). Areas with statistically significance differences are marked with black stippling.

FIG. 7. Differences in (top) pointwise IVT averaged over days with (a) $10 < \text{Pr} \leq 40$ mm and (b) $\text{Pr} > 40$ mm for the cool season (October–March) averaged over 26 yr for (left)–(right) hist, mid–hist, and mid–hist in (a), and (left)–(right) hist, end–hist, and end–hist in (b); and in (bottom) for specific humidity and horizontal wind patterns at 850 hPa for (left)–(right) hist, mid–hist, and end–hist averaged over all days over the cool seasons. The minimum wind vector length is set to 0.5 m s^{-1} for better visualization.

precipitation days, as shown above, indicative of drying over the warm season. The positive signal observed in winter precipitation extremes was also observed in Dominguez et al. (2012) but over entire WUS, and in the

context of return levels toward the latter half of the twenty-first century using an ensemble of RCMs.

Perhaps surprisingly, no significant changes in mean precipitation or precipitation extremes are predicted for

FIG. 8. Differences of rainy amount represented by the indices (left)–(right) P5mm, P10mm, P20mm, P40mm, and Pxmm between (top)–(bottom) mid–hist, end–mid, and end–hist averages. Areas with statistically significance differences are marked with black stippling. Areas with no data are indicated in gray.

California. In fact, the precipitation signal under a warmer climate is conceded to be more ambiguous for California (Neelin et al. 2013) in light of the extreme interannual variability (Dettinger 2011). Our results show a small decrease in extreme precipitation over the Sierra Nevada (although the decrease is not statistically significant). On the contrary, Kim (2005) found that under global warming, heavy precipitation events increase in the mountainous regions of the Northern California coastal range and the Sierra Nevada. This leads us to the likely conclusion (particularly in light of VR-CESM's own biases in this region) that projections in this region are highly dependent on model formulation and the representation of the large-scale circulation patterns effects—particularly ENSO, as further discussed in section 6d.

For the most extreme precipitation events ($\text{Pr} > 40 \text{ mm day}^{-1}$), there is a statistically significant increase along the U.S. Northwest coast ($\geq 60\%$), the Cascades ($\sim 50\%$), and the northern Rockies ($\geq 60\%$) by the end of the century. Significant increases are also apparent along the Klamath Mountains in California of about 20%–40% from hist to end. Changes in accumulated precipitation for these events (see Fig. 8) are consistent with the change in their frequency. With a projected amplification of temperatures in these regions of 4–5 K over the cool season, this increase of precipitation extremes is in excess of the 7% per degree change that would be anticipated from the C–C relationship. In this case, the probable cause of this excess is due to the intensification of the storm track along the coast, as further discussed in section 6a.

c. Regional precipitation frequency distributions

To further investigate the regional heterogeneity of changing precipitation, frequency distributions of daily rainfall based on the days (with $\text{Pr} \geq 0.1 \text{ mm day}^{-1}$) are plotted in Fig. 9 for (a) the Pacific Northwest, including Washington and Oregon, (b) central and Southern California, (c) the Intermountain West, including Nevada and Utah, and (d) the U.S. Southwest, including Arizona and New Mexico. Frequency plots are developed using simulation outputs at all grid points within each region. Results here mirror our earlier discussion. Over the U.S. Northwest, precipitation intensity increases with a shift toward a greater frequency of the most extreme precipitation days, especially by the end of the century, accompanied by a reduction in nonextreme precipitation days. No significant shifts can be observed for the California region. Over the Intermountain West, there is a similar trend toward more extreme precipitation as in the Northwest, but with no reduction in warm season nonextreme precipitation days. As for the Southwest, precipitation is more frequent, but the

response is weaker than that observed in the Intermountain West.

As a supplement to our results, the 95th percentile precipitation (P95) for all days over each simulation period is plotted in Fig. 10, to provide additional clarity in how the most extreme precipitation events are changing. Again, the shift to more extreme precipitation is most pronounced over the U.S. Northwest as warming intensifies through the end of the twenty-first century (P95 increased for about 20%–30%). For dry regions, including the U.S. Southwest and Intermountain West, precipitation tends to be more extreme (P95 increased for about 15%) with the increase of both the mean precipitation and number of rainy days (see Fig. 6) from hist to mid. However, this trend is suppressed over Southern California and the remaining southwestern region when the warming persists through the end period. In this region, where convective precipitation dominates, the increase in humidity does not keep pace with increases in saturated vapor pressure.

d. Disentangling the direct climate signal from ENSO and PDO

As discussed earlier, this study assumes a fixed pattern of SSTs that is consistent across all ensemble members and incorporates certain assumptions on the character of ENSO through the end of the century that arise from the coupled model. The phase of ENSO is well known to have important repercussions for precipitation extremes (Larkin and Harrison 2005; Allan and Soden 2008; Maloney et al. 2014; Yoon et al. 2015). In particular, Cai et al. (2014) found a significant increase in extraordinary precipitation events through the eastern Pacific Ocean in the twenty-first century within the CMIP5 ensemble, associated with increasing frequency of extreme El Niño events due to greenhouse warming. To better understand how ENSO has impacted our results, we now turn our attention to understanding how precipitation extremes behave in response to the phase of ENSO.

In our study, mean SSTs over the Niño-3.4 region are 26.83° , 28.62° , and 30.54°C for hist, mid, and end, respectively. Based on the ONI values, the mean SST anomalies over the Niño-3.4 region are 1.38, 1.71, and 2.30 K during El Niño years, and -1.16 , -1.62 , and -1.43 K during La Niña years, again for hist, mid, and end. It is apparent that within this dataset the magnitude of SST anomalies associated with the warm phase of ENSO has intensified. The spatial pattern of SST anomalies averaged over the warm and cool phases of ENSO can be found in Fig. S7 of the supplemental material. The calculated ONI values suggest an increasing frequency of El Niño through mid and an almost doubled frequency of La Niña during mid and end compared to hist (see Fig. S4).

FIG. 9. Frequency distribution of days (with $\text{Pr} \geq 0.1 \text{ mm day}^{-1}$) over the three time periods [1980–2005 (blue), 2025–2050 (orange), and 2075–2100 (red)] from all simulations dataset in four regions (with logarithmic vertical scale): (a) Washington and Oregon, (b) California (except the northern part, i.e., latitude no larger than 38°), (c) Nevada and Utah, and (d) Arizona and New Mexico.

Differences in mean precipitation and associated indices taken between the warm phase (i.e., El Niño) and cool phase (i.e., La Niña) of ENSO are provided in Fig. 11 for the cool seasons from hist, mid, and end. During the El Niño phase, intensified mean precipitation is expected over California and the U.S. Southwest (Hamlet and Lettenmaier 2007), accompanied by reduced precipitation intensity over the U.S. Northwest. In the La Niña phase, this pattern is reversed, with wetter conditions in the Northwest and a drier Southwest.

Consequently, ENSO is associated with a northwest–southwest precipitation dipole, triggered by ENSO’s modification of the storm track (Gershunov and Barnett 1998; Leung et al. 2003b), along with modulation of the enhanced precipitation variability (Cayan et al. 1999; Kahya and Dracup 1994). The dipole effect intensifies as a compound result of the climate warming effect and the changed magnitude of the ENSO anomalies. Strengthening storm patterns associated with ENSO are also found by Maloney et al. (2014) over California using CMIP5 output under

FIG. 10. The 95th percentile (P95) of precipitation constructed from all days for the historical period, 1980–2005 and (left)–(right) the changes over each future time period.

RCP8.5. This dipole is also apparent in the frequency of rainy days and extreme precipitation events.

The impact of ENSO can also be seen in the IVT difference that arises between El Niño and La Niña phases in each time period (see Fig. 12) and the accompanying 850-hPa wind patterns. During the El Niño phase, there is an increase in onshore moisture flux over California that triggers a returning circulation through the U.S. Northwest. This suggests that understanding moisture flux regulation from ENSO is a very important contributor to the character of future precipitation extremes.

Based on the above results, it is apparent that the magnitude of the effects of ENSO is comparable to or even higher than the impacts of climate forcing; that is, shifts in the future character of ENSO would have more dire implications for precipitation extremes than shifts

in mean climatological forcing. To investigate this further, linear regression has been applied at each grid point using a simple linear model that incorporates the phase of ENSO (using the Niño-3.4 SST anomaly) and the underlying climate forcing yearly (from mean GHG concentration). The precipitation indices are used as response variables. The significance of these two factors was then obtained from analysis of variance (ANOVA) output (see Fig. S8 in the supplemental material). The magnitude of the response associated with each factor was also computed (see Fig. S9 in the supplemental material). As expected, the ENSO forcing matches most closely with the difference between El Niño and La Niña (see Fig. 12). Hence, we observe that ENSO is a major driver of precipitation character through California, the Intermountain West, and the U.S. Southwest and does have an impact on mean precipitation through the

FIG. 11. Differences of (left)–(right) precipitation indices Pr and R1mm, R10mm, R20mm, R40mm, and Rxmm between warm and cool phases of ENSO over (top)–(bottom) each time period.

FIG. 12. Changes of IVT (color shading) for simulations between the warm and cool phases of ENSO of wet season (October–March) over rainy days (annually averaged), overlaid with seasonal mean wind patterns at 850 hPa (m s^{-1}). (Note: The minimum wind vector is set to be 0.5 m s^{-1} ; therefore, winds $< 0.5 \text{ m s}^{-1}$ are also plotted at the minimum length for better visualization.)

Cascades. In contrast, the impacts of climate forcing are visually similar to the pattern of the difference between the different time periods (see Fig. 6), and primarily impacts both extreme and nonextreme precipitation in the U.S. Northwest and Intermountain West.

We have also assessed the impacts of the Pacific decadal oscillation (PDO) on precipitation and observed only a weak correlation between the PDO pattern and precipitation. That is, precipitation features did not change substantially between the cool phase or warm phase of PDO when examining historical simulations. However, when in phase with ENSO, PDO did have an observable impact over the Northwest. This coupled effect has been found by studies such as that of Gershunov and Barnett (1998), who observed that ENSO and PDO can “reinforce” each other, with PDO responding to the same internal atmospheric variability as ENSO (Pierce 2002). In our simulations, there were roughly an equal number of positive and negative PDO years in the data from each time period. Since SSTs were fixed among ensemble members, the 26-yr simulation period might be insufficient to account for the variability of PDO. Therefore, in this study, we draw no conclusions on the impact of PDO.

7. Discussion and summary

In this study, an ensemble of 26-yr simulations has been conducted using VR-CESM with the finest local grid resolution of $\sim 0.25^\circ$ to assess the changing character of precipitation over the twenty-first century in the WUS. Climate forcing for future projections is prescribed under the RCP8.5 “business as usual” scenario. Our results were generally consistent with previous studies that addressed changes in precipitation over the twenty-first century using CMIP5 simulations (i.e., Sillmann et al. 2013b). However, by incorporating high resolution over the WUS, significantly more regional detail emerges with regards to a crucial enhancement of

precipitation representations, especially in regions of complex topography.

Evaluated against historical gridded observations and reanalysis data, VR-CESM was found to accurately capture the spatial patterns of precipitation, including its frequency and intensity, despite exhibiting an overestimation of precipitation over the eastern flank of the Cascades, throughout California’s Central Valley and along the Sierra Nevada. Nonetheless, there was pronounced improvement in the representation of precipitation features when compared with coarse-resolution simulations ($\sim 1^\circ$).

Mean precipitation and distributions of both non-extreme and extreme precipitation events have been investigated, as projected by the VR-CESM model under climate forcing. Although constrained by water influx and soil moisture, changes in extreme precipitation are hypothesized to follow the C-C relationship more closely than total precipitation amount ($\sim 7\% \text{ K}^{-1}$). In general, this only seemed to be the case over the Intermountain West; the northwest exhibited an enhanced response from extreme precipitation ($\sim 10\% \text{ K}^{-1}$), whereas California and the U.S. Southwest observed essentially no response.

From VR-CESM, the warming response to RCP8.5 climate forcing exhibited a roughly uniform character, although warming was more pronounced away from the coast and to the north. In the future, relative humidity (RH) was projected to be constrained by competing increases in both temperature and atmospheric water vapor content. RH tended to increase in regions where average temperature increase was below $\sim 2 \text{ K}$ and decrease when average temperature increase exceeded $\sim 2 \text{ K}$. This suggests that continental evaporation and oceanic water vapor transport are insufficient vapor sources to maintain RH levels above a certain threshold temperature. In response, mean precipitation increase is fairly inhomogeneous, with a more pronounced increase

in the northwest where water vapor transport is largely enhanced.

Over the Intermountain West and U.S. Southwest, an increase in warm season RH through midcentury led to a statistically significant increase in precipitation and nonextreme rainy days due to increased convection. This increase levels off through the end of the century, when continuing warming is observed to drive a reduction in RH. Nonetheless, there is a significant increase in extreme precipitation episodes ($Pr > 10 \text{ mm day}^{-1}$) over the Intermountain West that is not observed in the Southwest.

Over the U.S. Northwest, there is a clear shift from non-extreme precipitation events to extreme precipitation events associated with a moistening of the cool seasons and drying through the warm seasons. Specifically, although the total number of annual precipitation days remains relatively constant, low-rain days tend to decrease and heavy-rain days are projected to increase. In each case, the change is on the order of 10 days yr^{-1} . The increase of heavy precipitation frequency is driven by the increased IVT over the eastern Pacific in association with atmospheric river (AR) episodes, and intensified drying over the warm season is caused by a reduction in RH. The more frequent cool season precipitation extremes in this region tend to result in higher runoff events, which are in turn associated with a greater chance of flooding, particularly from rain-on-snow events.

Over California, except along the northernmost coast, there is no apparent climate signal in the mean precipitation or extremes. Interannual variability in this region associated with ENSO dominates precipitation patterns throughout the historical period and the twenty-first century. ENSO drives precipitation behavior by modulating the midlatitudinal storm track in this region. In particular, during the El Niño phase, there is an increase in onshore moisture flux over California that triggers a returning circulation through the U.S. Northwest. The results over California highlight the importance of understanding the response of ENSO to climate change (which is still largely inconsistent in CMIP5 climate models and so is a key source of uncertainty in our results), since variations in the magnitude or structure of ENSO will have immediate consequences for precipitation in this region.

The projected SSTs utilized for this study through the end of the century suggest that SST anomalies associated with ENSO will intensify. The impacts of ENSO are wide-reaching, with a statistically significant response observed in the character of precipitation throughout California, the Intermountain West, and the U.S. Southwest, as well as impacting mean precipitation through the Cascades. In contrast, the significance of climate forcing (when compensating for ENSO) largely

matched the differences observed between time periods, and had its greatest impact on both extreme and non-extreme precipitation in the U.S. Northwest and Intermountain West.

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