North American Climate in CMIP5 Experiments:

Part III: Assessment of 21st Century Projections

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Abstract

In Part 3 of a three-part study on North American climate in Coupled Model Intercomparison project (CMIP5) models, we examine projections of 21st Century climate in the RCP4.5 and RCP8.5 emission experiments. Aspects of North American climate change that are examined include changes in continental-scale temperature and the hydrologic cycle, extremes events, and storm tracks, as well as regional manifestations of these climate variables. We also examine changes in east Pacific and Atlantic tropical cyclone activity and North American intraseasonal to decadal variability, including changes in teleconnections to other regions of the globe.

Projected changes are generally consistent with those in CMIP3, although with better model agreement in some areas (e.g. projections of summer time precipitation decreases in the Caribbean and Southern Mexico). Although many projected changes in North American climate are robust across CMIP5 models, substantial disagreement in some areas helps to define priorities for future research. Areas of disagreement include projections of changes in snow water equivalent and diurnal temperature range on a regional basis, precipitation in the Southern U.S., Atlantic and east Pacific tropical cyclone activity, intraseasonal variability, and El Niño teleconnections. Model success in simulating historical climate as shown in Parts 1 and 2 of this study lend confidence to many of the projected results. However, model biases in other areas decrease confidence in projections, including changes in the timing of North American monsoon precipitation, growing season length along the West Coast, and the distribution of persistent drought and wet spells.
1. Introduction

The 21st century projections generated by the Coupled Model Intercomparison Project (CMIP5, Taylor et al. 2012) are analyzed here to assess climate change in North America (NA). This study accompanies two companion papers (Sheffield et al. 2012a,b) that assess CMIP5 models’ potential to accurately simulate regional climate in the 20th century. This study will begin with an examination of the changes in the continent-wide distribution of seasonal precipitation and temperature in two emissions pathways. It will then focus on a select set of regional climate features and their variability. These changes are considered in the context of the ability of models to accurately simulate current climate. As noted by Sheffield et al. (2012a), the ability of CMIP5 models to simulate current climate is generally comparable to that of CMIP3 models, with some improvement noted for individual models.

Projections of NA climate change have been conducted as part of previous climate assessments (e.g. CMIP3). Some salient points will be briefly summarized here. By 2080-2099, robust annual mean temperature increases are projected across NA with the greatest changes in Northern Canada and Alaska where 10°C mean wintertime temperature increases are projected in some scenarios (IPCC 2007). Robust increases in annual mean precipitation are projected for the Northern tier of the U.S., northward into Canada, and with robust decreases for the Southwest U.S., east Pacific warm pool, Caribbean, and adjacent land areas (e.g. Neelin et al. 2006; Seager et al. 2007; Seager and Vecchi 2010).

Beyond mean state changes, CMIP3 models predict a general increase in precipitation intensity (e.g. Diffenbaugh et al. 2005; Mahajan et al. 2012), particularly notable in the Northern Tier of the U.S. and Canada (Tebaldi et al. 2006). Increases in the duration and severity of
drought have been projected to occur in regions such as Central America and midlatitude NA (e.g. Sheffield and Wood 2008), of which increased temperatures and evapotranspiration are major components (Francina et al. 2010; Gutzler and Robbins 2011; Wehner et al. 2011). A general increase in heat waves, decrease in cold extremes, decrease in frost days, and increase in length of the growing season have been projected across large portions of NA (Meehl and Tebaldi 2004; Diffenbaugh et al. 2005; Biasutti et al. 2012; Christiansen et al. 2011; Diffenbaugh and Scherer 2011; Duffy and Tebaldi 2012). Decreases in the duration of the snowpack have been projected for many regions, in particular low altitude areas of the Pacific Northwest and Rockies (e.g. Brown and Mote 2009; Elsner et al. 2010). Such changes are projected to lead to earlier Spring snowmelt in many areas of the West (e.g. Hay et al. 2011). While model agreement is good on projected overall snow water equivalent declines in many areas by the end of the 21st century, some models show increases in snowpack along the Arctic Rim by 2100 (e.g. Hayhoe et al. 2004; Brown and Mote 2009), particularly at the height of the winter season even though the length of the snow season shortens (e.g. Raisanen 2008).

The projected response of NA climate in future emission scenarios is often more nuanced on the regional and local scales than for the continental scale features, especially when considering the evolution during the seasonal cycle. For example, Rauscher et al. (2008) noted an earlier onset of the midsummer drought in Mexico and Central America in model projections for the end of the 21st century. Previous studies project a redistribution of precipitation in monsoon regimes such as the SW U.S. with reduced Spring rainfall, and increased late rainy season rainfall (Seth et al. 2010, 2011; Biasutti and Sobel 2009). Ruiz-Barradas and Nigam (2010) showed that projections for the 21st century indicate a wetter north central U.S. during
Spring (increase in number of extreme Springs) and a drier SW U.S., but little consistency in summer rainfall tendencies among models in these same regions. The uncertainty in projected summer precipitation extends to adjacent land areas of the Gulf of Mexico (Biasutti et al. 2011). Changes in North Atlantic tropical cyclone activity are uncertain, as climate models produce differing patterns of tropical SST change, and different representations of tropical Atlantic SST relative to the tropical mean, which is a strong regulator of Atlantic hurricane activity (e.g. Knutson et al. 2010; Zhao and Held 2012). Past studies using CMIP3-class models have generally indicated that climate projections for the 21st Century at the local and regional levels remain a substantial challenge.

The present study provides a summary of projected 21st Century NA climate change in the updated state-of-the-art climate and Earth system models of CMIP5. The results contained herein are contributed by individual members of the CMIP5 Task Force of the National Oceanographic and Atmospheric Administration (NOAA) Modeling Analysis and Prediction Program (MAPP), and reflect each researcher’s individual expertise. Where appropriate, we make reference to individual papers submitted to this Journal of Climate special collection in parallel with this comprehensive study, and these individual contributions provide further depth to the findings that are summarized here.

Section 2 provides a brief introduction to the CMIP5 modeling project a well as the climate change experiments assessed. Section 3 presents an assessment of continental climate changes over the 21st Century. Section 4 assesses regional climate changes. How intraseasonal variability changes in the 21st Century is assessed in Section 5. Changes in Atlantic and east Pacific tropical cyclone activity are examined in Section 6. Multidecadal trends in interannual to decadal
hydroclimate variability are analyzed in Section 7. Conclusions are presented in Section 8.

2. CMIP5 Models and Experiments

We use projections from the CMIP5 multimodel dataset of historical climate and climate change experiments (Taylor et al. 2012). The experiments include both decadal predictions and long-term century-scale projections of climate. Table 1 summarizes information on the models used in this study. As noted in Taylor et al. (2012), one advantage provided by the CMIP5 experiments versus the CMIP3 effort is that the horizontal resolution of the atmospheric components of the coupled models has significantly improved. About one third of the models have atmospheric resolution of approximately 1.5° latitude or less, an improvement over CMIP3, as no model has resolution coarser than 3° latitude. This higher resolution is of some help in examining regional hydroclimate variables over NA, although still coarser than would be desirable in regions of complex topography and coastlines.

RCP8.5 and RCP4.5 will be highlighted, as they represent the core concentration pathways used for the CMIP5 project (Taylor et al. 2012). These experiments represent high concentration and moderate mitigation pathways in which radiative forcing due to anthropogenic factors reaches 8.5 W m\(^{-2}\) and 4.5 W m\(^{-2}\) by 2100, respectively (e.g. Meinshausen et al. 2011, Figure 4). Radiative forcing continues to grow beyond 2100 in RCP8.5, whereas in RCP4.5 stabilization at 4.5 W m\(^{-2}\) occurs around 2050 and remains fixed. In terms of the time evolution and value of globally-averaged radiative forcing at 2100, these pathways most closely resemble the A2 and B1 scenarios for CMIP3 used in the IPCC Assessment Report 4 (IPCC 2007, Figure 10.26). These climate change experiments will be compared to historical runs of the same
models in which they attempt to reproduce past climate from the mid 19th Century onward when being forced by observed trace gases, natural and anthropogenic aerosols, solar forcing, and other agents (Taylor et al. 2012). The number of models used in a particular analysis shown below is often limited by availability of data at the time of this study or local storage space, although we try to be as comprehensive as possible.

3. Continental Climate

a. Temperature and precipitation

We first examine projected changes on the continental scale at the end of the 21st Century relative to the 20th Century climate. Sheffield et al. (2012a) noted that CMIP5 models have success at the capturing the broad-scale features of NA surface temperature and precipitation in current climate, although with some regional-scale biases. Figure 1 shows the 15-member multimodel mean December-February (DJF) and July-August (JJA) precipitation changes during 2070-2099 for RCP8.5 relative to a 1961-1990 base period. Figure 2 shows the model agreement on the precipitation changes.

Figures 1 and 2 indicate increases in ensemble mean wintertime precipitation along the west coast of NA from California to Alaska, as well as along the NA East Coast from the mid Atlantic states northward. Model agreement for these changes is high north of 40°N, where all but one or two models agree on these changes for all locations. Comparison to a similar ensemble of 15 CMIP3 models indicates that while the large-scale pattern of precipitation increases at mid- to high- latitudes and precipitation decreases in the subtropics is similar between the two intercomparisons, one notable difference is that the boundary between these
changes has shifted slightly south, yielding projected precipitation increases over parts of California in CMIP5 (Neelin et al., 2012). Summertime ensemble mean precipitation changes are characterized by higher precipitation amounts in Alaska and the Yukon, where the models all agree on the sign of the changes. All models also suggest precipitation increases along the Arctic coast across the entire length of NA. The multimodel mean indicates reduced summertime precipitation in the east Pacific warm pool and the Caribbean, with agreement of all models in the vicinity of major Caribbean islands, the Yucatan, and in Southwestern Mexico adjacent to the east Pacific warm pool. The agreement on these changes for the Caribbean and Mexico was high for CMIP3 models (e.g. Neelin et al. 2006) and is reinforced as a region of even higher intermodel agreement in CMIP5.

Figure 3 shows the comparable multimodel mean changes for surface temperature for RCP8.5 during JJA and DJF. As expected, warming is projected across all regions of NA, with the greatest warming concentrated during wintertime at high latitude regions, where multimodel mean temperature increases peak near 15°C in the vicinity of Hudson Bay.

b. Moisture fluxes

CMIP5 model projected differences in vertically integrated moisture transport (MT) to 500 hPa (vectors) and its divergence (contours) are shown in Figure 4 for four coupled models, for RCP8.5 (2081-2100) minus historical (1981-2000) experiments. The MT is well simulated in these models in historical simulations (Sheffield et al. 2012a). While the number of models is too few to provide robust conclusions regarding changes in MT, this analysis provides process level support for model agreement/disagreement in precipitation projections. In summer, the models
suggest an increased transport in both (East coast and Great Plains) branches of moisture flow, with a poleward intensification in the coastal branch associated with a poleward shift in moisture convergence on the northwest flank of the Atlantic subtropical anticyclone and increased moisture divergence to the south. The increased Great Plains flux is consistent with projected strengthening of the Great Plains low-level jet, as described in more detail below. Three of the four models show increased divergence in the northern plains during summer, and all show increased divergence in the Pacific Northwest, where model agreement exists on precipitation decrease exists (Figure 2).

In winter, the models indicate stronger MT from the Pacific into the Northwest, but the latitude of the increased westerly transport is critical in determining whether much of California is in a region of increased or decreased divergence. Agreement also exists that increased MT would result in more convergence from the mid-Atlantic through eastern Canada, while the Gulf coast and Florida would see increased divergence of moisture.

c. Evapotranspiration and Runoff

Future changes in precipitation and how it is partitioned into evaporation and runoff have implications for water availability and the occurrence of extreme hydrological events such as floods and droughts. Evaluations of CMIP3 projections showed that changes in precipitation coupled with increasing atmospheric demand leads to drying of sub-tropical regions including Central America and the southwestern US (Wang, 2005; Seager et al. 2007; Sheffield et al., 2008), and increased potential for flooding because of more frequent and intense precipitation events, changes in snow accumulation and melt timing, and changes in antecedent soil moisture
conditions (e.g. Hamlet and Lettenmaier, 2007; Das et al., 2011). Here we analyze changes in the terrestrial water budget on a regional basis by calculating 30-year averages in annual mean precipitation, evapotranspiration and runoff between 1971-2000 and 2071-2100 for 13 climate models for six regions of NA (Figure 5). Changes in water storage (i.e. surface water, soil moisture, groundwater, etc.) over the 30-year periods are assumed to be small compared to the other terms and so precipitation should balance the sum of evapotranspiration and runoff.

For western and eastern NA regions an increase is projected in the multi-model mean annual precipitation under RCP8.5, consistent with Figure 1. The precipitation increases in these regions are apportioned more to evapotranspiration than to runoff, although models tend to overestimate evapotranspiration in historical simulations (Sheffield et al. 2012a). In the central region, annual mean precipitation increases are more modest. In high latitudes (Alaska/Northwest Canada and Northeast Canada) the multi-model mean precipitation is projected to increase, consistent with Figure 1, and is mostly partitioned into increases in runoff, rather than evapotranspiration. In Central America, precipitation is projected to decrease with most of the decrease manifesting in decreasing runoff.

d. Snow

Reductions in snow cover extent and amount are expected in the future as a result of increasing temperatures modified by changes in precipitation and their seasonal interactions. This has important implications for water supply, hydropower generation, and ecosystems, and feedbacks with the underlying soil and permafrost (Lawrence and Slater, 2010) and to the atmosphere through changes in albedo (Qu and Hall, 2007). CMIP3 projections (Räisänen, 2008)
indicated that warming reduces the snow season length from both ends across NA, but mid-
winter snow water equivalent (SWE) is expected to increase in more northerly colder regions
because of increased winter snowfall, but decrease in more southerly regions where temperature
effects on precipitation phase and melting dominate over any changes in precipitation amount.

Changes in snow are calculated using the CMIP5 model SWE values. Figure 6 (top)
shows the seasonal cycle of changes between 2099-2070 and 1971-2000 in monthly mean SWE
averaged over NA for a set of nine models. All models project a decrease in SWE throughout the
year with maximum changes during the peak of the snow season in Jan-Apr. The MME mean
decrease in SWE averaged over NA is about -30mm (with an inter-model range of about 25mm)
in the spring and about -5mm in the summer time (range is about 10mm). Spatially, the majority
of NA (south of 70°N) is projected to experience a decline in SWE where increasing
temperatures have a dominant effect by reducing the ratio of snowfall to rainfall and accelerating
melting (Figure 6, bottom). These reductions are concentrated in the Rocky Mountains to
southern Alaska, on the eastern provinces of Canada, and with a lower magnitude on the
Canadian Prairies. North of 70°N, SWE is projected to increase due to increasing precipitation,
which outweighs the effects of increasing temperature. Uncertainties across models are likely
associated with differences in the temperature projections, to which modeled snow is highly
sensitive (Räisänen, 2008). At higher latitudes, the sign of the change is uncertain in transitional
regions because of the competing effects of increasing snowfall and warming temperatures, and
in regions of increasing SWE where the magnitude of the precipitation increase is also quite
uncertain.
e. Growing season length

Projected warming is likely to be associated with impacts on temperature sensitive flora and fauna, as well as agriculture. We calculate changes in biophysical-relevant temperature thresholds including the last spring freeze, first autumn freeze and the growing season length in RCP8.5 at 2071-2100. By end of century the analysis of the 6 models (Figure 7) indicates that the growing season will increase across the NA continent, although substantial variability in the magnitude of these changes exists on a regional basis. The largest changes occur over the western US and northern Mexico, where increases of 2 months or more are projected. It should be noted that these same regions are characterized by some of the largest negative biases in historical simulations (Sheffield et al. 2012a). In the central US and Canada, increases of about 3-6 weeks are projected. The lengthening of the growing season is caused by earlier last spring freezes and later first autumn freezes (not shown), but the change in the former is generally larger.

f. Frost days

Figure 8 shows the change in the number of frost days (FDs) simulated by six models at 2071-2100. Maximum changes of over 80 days are limited to coastal regions of southern Alaska and western Canada. The rest of the continent shows changes of between 50 and 80 days in the western third from the US western mountains up through the Canadian Rockies to Alaska, and between 20 to 50 days over the eastern two thirds, with less change in the southern U.S.. Most uncertainty exists across the model projections in the West with decreases that range from about 50-60 days for CCSM4 to more than 100 days for IPSL-CM5A-LR. The highest agreement is in
the Canadian Northern Territories and for central America. Some of the differences can be explained by the historical biases in the models (Sheffield et al., 2012a), which may limit or enhance future changes even if the projected shift in temperature is the same across models. For example, CCSM4 has historically too few FDs in the central US, and projects less of a decrease in this region in the future. The IPSL-CM5A-LR and CSIRO-MK3-6-0 models project the largest decreases in FDs over the western U.S., but these two models also have the largest overestimation of historic FDs in this region. These results indicate that bias correction of the modeled extreme values of these type can help reduce the uncertainty in future projections.

\[g. \text{Daily temperature range}\]

One robust global climate change signal over the 20th Century was the widespread decline of daily temperature range (DTR), especially in winter, resulting from nighttime temperatures warming faster than daytime (Karl et al. 1993, Dai et al 1999, Easterling et al 1997, and Vose et al 2005). In RCP4.5, the CMIP5 models project sharp decreases in wintertime DTR in the mid 21st Century, most prominent in an east-west oriented band at northern latitudes where DTR decreases by more than 0.2 °C/decade (Figure 9). This decrease in DTR is largely due to preferential increases in nighttime temperature. In the southern U.S. and Mexico, DTR is projected to increase, although with larger uncertainty as indicated by the larger inter-model variance. During summer, DTR is projected to slightly increase (<0.1 C/decade) in the north central section of the U.S. In the southeastern U.S., the multimodel mean DTR signal is rather weak, which is also accompanied by larger variance among the models. The uncertainty in projected DTR trend is generally higher in the lower latitudes. If we view the signal (DTR trend)
to noise (inter-model variance) ratio as a simple measure of the confidence in the model projections, the northern Rocky Mountain region has smallest uncertainty in future projections.

h. Extreme events

CMIP5 historical and RCP 4.5 projection experiments are examined to assess projected changes in drought and persistent wet spells over the U.S.. The methodology used is the same as in part 1 of this study (Sheffield et al. 2012a), and repeated in the Figure 10 caption. Five models that contained three or more members for both the historical and projection experiments were chosen to analyze the response of extreme events to an increase in CO₂. We first calculated the difference in extreme precipitation event frequency of occurrence (FOC) when each experiment’s own climatology is used as a baseline. Each model shows little difference in the FOCs for both positive and negative events (not shown). Based on the new projected climatology, there is little change in the FOC pattern for both precipitation and soil moisture. The results become more interesting when the historical climatology is used as to assess future changes in dry and wet spells (Figure 10). Models tend to show a decrease in positive (wet) events in Mexico and the southwest and an increase in positive (wet) events in the eastern United States. The opposite is true for negative (dry) events. The CAN and IPSL, and MPI and CCSM4 to a lesser extent, also show increases in positive events in the northwest and northern Great Plains. These patterns of extreme event changes are consistent with differences in annual mean precipitation for each model (not shown). It should be noted that model ability to simulate the correct distribution of extreme precipitation event occurrence in the current climate is limited (Sheffield et al. 2012a).
Our analysis of extremes now extends to high temperatures using one ensemble member for nine models for RCP4.5 and RCP8.5 at 2071-2100. Daily maximum surface temperatures ($T_{\text{max}}$) are used to compute the number of days per year that exceed 90 F and 100 F, referred to as ND>90F and ND>100F, respectively.

For RCP4.5, all selected models project an increase of ND>90F across much of the U.S. and Mexico (Figure 11). The largest increases are projected to occur in the Midwest and northeast US, where ND>90F is projected to increase by 50% to 100% by the late 21st century relative to its current climatology. An increase of ND>90F by 10%-50% is also projected over the Southwest and South Central U.S., where ND>90F is climatologically high during the 20th century. These models also consistently project a weak increase of ND>90F over Wyoming, Utah and Colorado. Generally, the models that underestimate climatological ND>90F (Figure 11) tend to project a weaker increase of ND>90F (e.g., MPI, IPSL, MRI), whereas models that overestimate climatological ND>90F tend to project a stronger increase of ND>90F (e.g., HadGEM2, MIROC5). These discrepancies are greatest near the Pacific Coast. The CCSM4 has the least bias and projects an increase of 50-100% of ND>90F over Midwest, Northeast and Pacific Coast, and up to 50% over the Southwest and South Central U.S.

Under RCP8.5, except for MRI which severely underestimates historical ND>90F, all other models project that ND>90F over the majority of the U.S. south of 40°N will exceed 60-90 days. The spatial pattern of increasing ND>90F is generally similar to that projected under RCP4.5, but with 10-30% greater magnitude.

4. Regional Climate
a. North Pacific and North Atlantic Storm Tracks

The projected change in Northern Hemisphere storm tracks has been examined based on 6-hourly data available for 16 models from the CMIP5 archive (Figure 12). Near the tropopause (250 hPa), the models project a strengthening of the storm track (ST) on its poleward flank, and a slight weakening on its equatorward flank during Northern Hemisphere winter. In summer, the models project a significant decrease in upper tropospheric ST activity south of the ST peak, and weak increase over the high latitudes. These results are consistent with previous studies based on CMIP3 (e.g. Yin 2005; Teng et al 2008) that indicate a poleward shift of the STs under global warming.

In the mid-troposphere (500 hPa), the models mainly project a decrease in ST activity. Some indication exists that over the Central and Western Pacific, the decrease is largest on the equatorward flank of the ST in winter, while over the Eastern Atlantic, the decrease is largest again on the southern flank of the ST in summer, but the shift is not as clear as that found near the tropopause level. Near the surface, the models project mainly a decrease in ST activity extending from Alaska across Canada and the U.S., and extending into the Atlantic in winter, with not much indication of a poleward shift. In summer, the maximum projected decrease is found mainly on the equatorward flank of the ST, indicating weakening and poleward shift of the ST. Projected changes from spring and fall lie between those for winter and summer, but results from Spring (Fall) more resemble those from winter (summer).

Here we provide a complementary analysis to that above using the Hodges (1994; 1995) cyclone-tracking scheme on 6-hourly mean sea level pressure data to assess changes in extratropical cyclone activity along the East Coast. Sheffield et al. (2012a) presented the
historical (1979-2004) predictions of western Atlantic extratropical cyclones during the cool season (November to March) for the mean of 15 CMIP5 models, which shows substantial skill at simulating the distribution of cyclone activity, although with modest under prediction of amplitude. Colle et al. (2012) highlighted the details of this historical cyclone analysis and the 21st century predictions in this region using 15 CMIP5 models. Figures 13b-d shows the difference in cool season cyclone track density for the multimodel mean between each of the three separate 30-year future periods (2009-2038, 2038-2069, and 2069-2098) for RCP8.5 and the historical (1979-2004) period (Figure 13a). Only a slight decrease in cyclone activity is projected over parts the western Atlantic storm track for 2009-2038 (Fig. 13b); however, Colle et al. (2012) showed that this reduction may be more widespread if one only considers the highest resolution CMIP5 models. The reduction in cyclone density is more apparent in the multimodel mean for the 2038-2069 period, with a reduction of 0.2 – 0.5 cyclones per cool season per 50,000 km² (5-15% change), primarily along the southern half of the western Atlantic cyclone storm track, which is near the Gulf Stream boundary. Meanwhile, a slight increase in cyclone density is projected to the north over parts of the northern New England and Nova Scotia, enhanced in the highest resolution models (not shown). These future changes relative to the historical period continue to increase in size and amplitude (10-20%) for the 2069-2098 period. The cyclone density also decreases around the Great Lakes and to the east of southern Greenland.

A gradual reduction in the maximum intensity of cyclones occurs within the dashed box region of Fig. 13a for the three future periods (Fig. 14); however, this reduction is delayed around 990 hPa during the first two 30-year periods. By 2069-2098, a 0.5 to 1.5 (5-10%) reduction in the number of cyclones is projected between the 960 and 1010 hPa pressure bins. In
contrast, Colle et al. (2012) showed a 5-10% increase in the number of 960-980 hPa cyclones along the U.S. East coast (not shown).

b. Northeast U.S. and Western Atlantic precipitation

In Sheffield et al. (2012a), several CMIP models were evaluated to determine how well they can simulate precipitation over the Northeast U.S. and western Atlantic during the historical period (1979-2004) cool season (November-March). Figure 15 shows the mean CMIP precipitation for this historical period, as well as the precipitation difference (in mm/season and percent change) for the early 21st century (2009-2038) minus historical and late 21st (2069-2098) minus historical for these same models. For the early 21st century (Fig. 15b), the largest precipitation increase (80-150 mm) is from the southeast U.S. northeastward into the western Atlantic in association with the mid-latitude storm track, while the largest percentage increase is over eastern Canada (15-25%). As illustrated in Fig. 13b, a slight decrease in extratropical cyclones occurs in this region, which suggests that the same or fewer storms will have more precipitation. By the later 21st century (Fig. 15c), the precipitation change from the historical is 15-25%, while the largest change of 50-85% (40-100 mm/season) is around eastern Canada. Over the Northeast U.S., the mean precipitation increases 25-35% by the later 21st century. Over inland locations of the Northeast U.S. (Fig. 15d), the number of relatively light precipitation events (< 10 mm/day) increases from 5-20%, with much of this change occurring by the early 21st century. The number of relatively heavy precipitation events (> 20 mm/day) over the Northeast U.S. doubles by the early 21st century, and increases by 5-6 times by the later 21st century. The results suggest that a potential exists for a dramatic increase in the number of
extreme rainfall events over the Northeast U.S. during the next 50-75 years.

c. Western Water

Changes in snow are important in western regions because of the implications for water resources and winter tourism (see Section 4c). Figure 16 shows changes in April SWE over the western US from nine models for RCP8.5. April SWE is an indicator of total snow accumulation over the winter-time and the potential water resources availability for the coming year, which is especially important in California and the Colorado River basin. All models project a decrease in April SWE of up to 100mm in the central Rockies and Canadian Rockies. The spatial resolution of the models is generally too coarse to discern the details of the Sierra Nevada range on the California-Nevada border, but the models project a decrease in April SWE over this area. This broad decline is consistent with high-resolution hydrological changes using downscaled projections (Hayhoe et al., 2004). The decrease in SWE is driven by higher temperatures that increase the ratio of rainfall to snowfall (winter time precipitation is projected to remain the same or increase slightly. See Fig. 1) and increase melt efficiency, therefore moving the spring melt earlier in the season. Consequences from the shift in snowmelt timing may also because water rights can be month dependent (Hayhoe et al., 2004).

d. North American monsoon

We first concentrate on the core of the NA monsoon (NAM) region (23°N-30°N, 105°W-110°W) to assess projected behavior in the subset of models that best simulate the historical precipitation climatology in this region. In Sheffield et al. 2012a, the seasonal cycle in
precipitation of 21 CMIP5 models over the historical period (1979-2005) were evaluated to identify models that have a reasonable representation of the seasonal cycle in precipitation over the core NAM region. The results of this analysis indicate that eight of the 21 models have small (lag=0) phase errors with respect to the observations. Unfortunately, only six of these eight models provide precipitation data for RCP 4.5 from 2070-2100 at the time of this analysis.

The RMS of the future minus historical monthly rainfall climatology and the annual mean rainfall differences are shown in Figure 17 for these six models. The results suggest that, even for models that correctly capture the timing of the seasonal cycle of precipitation in the region, large discrepancies exist in what these models project for the monthly mean magnitude (range of 0.1 to 0.5 mm/day) and annual mean changes (range of -0.25 to 0.15 mm/day) in precipitation for the monsoon region. In some cases the projections reflect model historical period biases, for example in the CanESM2 (dry) and UKMO HadGEM2 (wet) models (Sheffield et al. 2012a). For models with little or no historical bias (CNRM-CM5 and CSIRO-Mk3-6-0), agreement is somewhat better and suggests little to moderate drying for the NAM region, consistent with the continental scale drying shown in Figure 1.

While the analysis described above suggests that projecting annual mean changes over the core monsoon region remains a challenge for global models, changes on a seasonal basis may be more robust, as suggested in Figures 1 and 2. In Figure 18 the multi-model average monthly precipitation from thirteen models in the CMIP5 database capture a northward migration of precipitation in the NAM region (longitudes 102.5° to 115°W) during the warm season. The models' historical precipitation (b) begins later and is weaker than the observed estimate (a). Model projections from RCP8.5 are consistent with the CMIP3 results (see introduction) and
demonstrate reduced precipitation between 10°-25°N through the cold season extending into June and July, and increased precipitation in September and October, although this tendency also resembles the bias of the models in the historical simulations. These reductions in precipitation between 10°N-25°N in June and July are consistent with reduced Mexico and Caribbean precipitation in this latitude band seen in Figures 1 and 2 for JJA.

e. Great Plains Low Level jet

The Great Plains low level jet (LLJ) is a basic component of the warm season circulation in NA that provides a moisture source for Great Plains precipitation. It emerges in April, strengthens and peaks in June and July, and vanishes in late September (e.g. Sheffield et al. 2012a). Figure 19 reviews the ability of five climate models to capture the intensity and seasonal cycle of meridional flow in the LLJ region. As shown in Figure 19, the NCAR CCSM4, MPI_ESM, and HadESM2-CC models are best able to capture the amplitude of the LLJ during boreal summer, and vertical cross sections and horizontal maps of the jet verify this behavior (not shown here). Figure 20 shows how the jet strength evolves for the NCAR CCSM4, MPI_ESM, and HadESM2-CC models in RCP4.5 and RCP8.5. The projections generally indicate a strengthening of the LLJ by about 15-25% in these three models by 2071-2100 in RCP8.5.

The development of warm season precipitation anomalies or extreme wet and dry conditions in the Great Plains and the Midwest is largely a result of the weather/precipitation systems that develop and depend strongly on the dynamic stability of the large-scale circulation (e.g., Moore et al. 2012; Veres and Hu 2012). This notion explains the apparent discrepancy
between the predicted strengthening LLJ shown in Fig. 20 and the decrease in summer precipitation in the Great Plains shown in Figures 1 and 2. This discrepancy suggests changes in the future summer circulation regime that would produce less precipitation in a moist environment with possibly a different intensity distribution of precipitation events in the Great Plains.

f. Arctic sea ice

Under RCP4.5 (Figure 21), the multi-model ensemble mean does not reach ice free-conditions by 2100, though the minus 1 standard deviation bound reaches nearly ice-free conditions (less than $1 \times 10^6$ km$^2$) around 2050. In contrast, the multi-model mean ice extent drops below $1 \times 10^6$ km$^2$ around 2060 for the RCP8.5 emission scenario, with the minus one standard deviation bound showing completely ice-free conditions by 2050. The September ice extent is stabilized around $4.0 \times 10^6$ km$^2$ under RCP2.6 (not shown).

g. Warming hole in the Eastern U.S.

During the second half of the 20th century, the central-eastern U.S. experienced cooling trends while global mean temperatures warmed. We refer to this cooling region as a “warming hole” (WH), following Pan et al. (2004). A number of studies have attributed the mechanisms for this abnormal trend to large-scale decadal oscillations such as PDO and AMO (Robinson et al., 2002, Kunkel et al., 2006, Meehl et al., 2012). Other studies indicate that regional scale processes such as hydrological cycle (Pan et al, 2004) and land surface interaction (Liang et al. 2007) may play roles in the WH. Most CMIP3 models have had trouble reproducing the WH
(Kunkel et. al., 2006). In Sheffield et al. (2012b), it was shown that the 95% uncertainty range of historical CMIP5 simulations does bracket the observed negative temperature trend, although the multi-model mean time series had limited skill at reproducing the warming hole.

Here we assess whether any evidence for an eastern U.S. WH exists in 21st Century projections. Figure 22 shows the multimodel mean DJF and JJA temperature trends during 2015-2050 and 2051-2098 over NA for RCP8.5. The multi-model average in RCP8.5 shows relatively slower warming in eastern U.S. compared to western U.S. during the winter (DJF), but little signal in summer. Figures 23a and 23c show time series of 30 year running trends (warming rate) of Eastern U.S. winter temperature for RCP8.5 and RCP4.5. The multimodel median of RCP8.5 simulations indicate a continued higher warming (~ 0.5 °C/decade) through the 21st century, whereas RCP4.5 indicates a decline in warming rate from ~ 0.4 °C/decade in the first half to ~ 0.2 °C/decade in the second half of the 21st Century (3rd panel), consistent with the stabilization of forcing at 2050 in this pathway. Almost the entire uncertainty range in RCP 8.5, and the majority of the uncertainty range in RCP 4.5, is above the zero line for 30 year running winter temperature trend in eastern U.S. (Figure 23a,c); hence it is very likely in RCP 8.5, and likely in RCP 4.5 that the negative temperature trend in eastern U.S. will disappear in the projected 21st century climate.

Figure 23b and Figure 23d show the difference in trends between the east and west U.S. For the east – west temperature trend difference, the uncertainty range is equally distributed above and below the zero line (Figures 23 b,d). A small negative difference (0.1 °C) between east and west temperature trends in Figure 22 could be the result of the selected time period. The multi-model median curve in Figure 23b and Figure 23d varies between ±0.1 °C. Hence, the
chance of relatively slower warming in the eastern U.S. compared to western U.S. is the same as
the opposite.

5. Tropical Intraseasonal Variability

a. Mid-summer drought

For most of Central America and southern Mexico, climatological precipitation is a
maximum in June and September, bracketing a period of reduced rainfall during July-August
known as the midsummer drought (MSD; Portig et al. 1961, Magaña et al. 1999). A previous
assessment of CMIP3 model performance at simulating the MSD and future projections
(Rauscher et al. 2008) suggested that many CMIP3 models are capable of simulating the MSD
despite an overall dry bias, and that the MSD is projected to become stronger with an earlier
onset. An updated evaluation of this feature (Sheffield et al. 2012b) indicates that many CMIP5
models are also able to capture both the spatial and temporal aspects of the MSD.

The CMIP5 multimodel mean projection for changes in precipitation during each of the
summertime months (June through September) is shown in Figure 24. During each of the
summertime months, the east Pacific ITCZ is projected to shift southward in concert with a
drying over the east Pacific warm pool (EPWP), Central America/southern Mexico, and the
Caribbean with enhanced drying over the major Caribbean islands of Cuba, Hispaniola, and
Jamaica. The strongest drying is projected to occur during July and August, which are the
months during which the MSD occurs in many regions throughout the Inter-Americas region.
The precipitation changes shown are generally consistent with the JJA average precipitation
patterns shown in Figure 1, although Western Mexico is projected to experience wetter
conditions during the late summer (September), consistent with the analysis of Figure 18.

Next, a simple algorithm for detecting and quantifying the climatological MSD strength is used that does not assume a priori which months are maxima and which months constitute the MSD (Karnauskas et al. 2012). Consistent with the month-by-month rainfall projections, the CMIP5 multimodel mean provides a very robust projection of a stronger MSD for most regions that experience an MSD today (Figure 25). The maximum MSD increases from ~2.5 mm/day to ~3 mm/day (4 mm/day) in the RCP4.5 forcing experiment, and nearly doubles to >4 mm/day in the RCP8.5 forcing experiment. The multimodel mean projection is qualitatively consistent in each of the CMIP5 models that best replicate the observed MSD (Sheffield et al. 2012b). The stronger MSD is a result of early and mid summer rainfall being reduced relative to the late summer peak (see also Figure 18). The extension of the MSD northward along the Gulf coast of Mexico and into the U.S. is projected to strengthen in both forcing experiments. Some regions that did not previously exhibit a MSD (e.g., Panama) develop a moderate MSD under either forcing experiments.

b. Intraseasonal oscillation

Some have argued that the leading mode of intraseasonal variability (ISV) in the east Pacific warm pool is a regional manifestation of the Madden-Julian oscillation (Maloney et al. 2008). Sheffield et al. (2012b) documented the ability of CMIP5 models to capture this leading mode of 30-90 day precipitation variability over the east Pacific and adjacent regions of Mexico and Central America and the Caribbean during June-September. This analysis was done using complex empirical orthogonal function (CEOF, Barnett 1983; Horel 1984) analysis that was
previously used to document the leading mode of ISV over the east Pacific during boreal
summer (Maloney et al. 2008). Based on a pattern correlation and assessment of amplitude of the
leading CEOF as compared to TRMM observations, seven models were judged to produce
realistic ISV.

Figure 26 shows the amplitude of the leading CEOF mode in one model with poor
simulation of ISV (CCSM4) and three models with realistic ISV (HadGEM2_CC,
HadGEM2_Es, MPI_ESM_LR) for 1981-2005, the 2076-2099 projection in RCP8.5, and the
difference in amplitude relative to present day. In the models with realistic ISV, robust changes
include increased amplitude of the leading CEOF on the southern fringe of the amplitude
maximum, with inconsistent changes in amplitude elsewhere. A plot of intraseasonal
precipitation variance supports these results (Jiang and Maloney 2012). Interestingly, some of
these models also exhibit significant decreases in ISV in the Caribbean. These changes in
variance mimic the mean precipitation changes shown in Figures 1 and 24.

6. East Pacific and Atlantic Tropical Storm Track and Cyclone Activity

a. Tropical storm track

The first approach we use to examine tropical storm activity is to calculate tropical storm
track statistics from 6-hourly 850 hPa relative vorticity following the method of Hodges (1995),
which captures traveling synoptic disturbances characteristic of the West Atlantic, Caribbean and
tropical East Pacific during summer (e.g. Serra et al. 2010). These systems contribute to the
majority of weather across the Intra-Americas Sea and are the seeds of a majority of West
Atlantic and East Pacific tropical cyclones. Figure 27 shows the 7-model mean for the future

period (May-November 2070-2100) and the future minus historical period differences in tropical
storm track density for each model. Several of these CMIP5 models (CanESM2, CCSM4,
HadGEM2-ES, MIROC5 and MRI-CGCM3) indicate a southward shift of the main storm track
as well as an increase in track density for the future projections, consistent with results found
using CMIP3 models (Bengtsson et al. 2006; Colbert et al. 2012). This southward shift is also
observed in the pattern of mean precipitation change seen in the JJA analysis in Figures 1 and 24.
No significant changes are observed in the mean strength of the tropical storm track based on
these seven models (not shown).

b. Tropical Cyclones in the North Atlantic and Eastern North Pacific

Using the Camargo and Zebiak (2002) algorithm we analyzed track changes in tropical
cyclone (TC)-like structures in RCP4.5 and RCP8.5 in four of the CMIP5 models shown by
Sheffield et al. 2012b to produce the best simulation of TC tracks in the current climate (GFDL-
ESM2M, MPI-ESM-LR, MRI-CGCM3, and MIROC5), although all of these models still
severely underestimate the number of TCs per year in the Atlantic and East Pacific.

Figure 28 shows the mean track density in the North Atlantic for the historical
simulations for the period 1951-2000, as well as the difference of the track density between the
period 2051-2100 and the historical track density for RCP4.5 and RCP8.5. While the MRI model
shows notable increases in TC track density for the Atlantic north of 25°N, the MIROC model is
characterized by substantial decreases in this same region. Changes in the GFDL and MPI
models fall in between. To be more quantitative, Figure 29 shows the number of TCs in the
Atlantic and eastern north Pacific for these four models in the historical runs, and RCP4.5 and
RCP8.5. For the North Atlantic, the MRI model exhibits a significant increase in Atlantic TC numbers in RCP8.5 relative to the historical period, whereas the MIROC5 model exhibits a significant decrease in Atlantic TC numbers in both RCP4.5 and RCP8.5. For the East Pacific, results are also mixed, with significant increases in TC numbers in future climate in the MPI model in the RCP4.5 scenario, and significant decreases in the MIROC5 model.

c. Downscaling with a high resolution model

To complement the analysis in Section 5b, we use a dynamical downscaling approach in which a high resolution global atmospheric model (GFDL HIRAM; Zhao et al. 2009; Zhao and Held 2012) is integrated and forced by CMIP3 and CMIP5 coupled model projected SSTs and sea ice concentrations. Recent studies suggest that when forced by the observed SSTs and sea-ice concentrations, a global atmospheric model with a resolution ranging from 50km to 20km can accurately simulate many aspects of TC/hurricane frequency and its variability for the past few decades during which reliable observations are available (e.g., Sugi et al. 2002; McDonald et al. 2005; Yoshimura et al. 2006; Oouchi et al. 2006; Bengtsson et al. 2007; Gualdi et al. 2008; LaRow et al. 2008; Zhao et al. 2009).

We first generate a present-day control simulation by prescribing climatological SSTs and sea-ice concentration (seasonally varying with no interannual variability) using time-averaged (1982-2005) Hadley Center Global Sea Ice and Sea Surface Temperature (HadISST) data (Rayner et al. 2003). For the CMIP3 global warming experiments, we add the SST anomalies (also seasonally varying with no interannual variability) projected by the coupled models to the climatological SSTs and double the CO₂ concentration. For the CMIP5 high-
resolution time-slice simulations with prescribed SSTs and sea-ice concentrations, the
specifications for both the present-day and the future projection experiments also include
interannual variability, and feature future SSTs from two GFDL coupled models (ESM and
CM3). The results from downscaling the GFDL CMIP5 projections include changes in both
SST anomalies and different specifications for greenhouse-gases and aerosols depending on the
pathway used (Held et al 2012). The storm detection and tracking algorithm we use in the
analysis is described in (Zhao et al. 2009).

The GFDL C180HIRAM simulations with CMIP3 model forcing produce a large inter-
model spread (standard deviation of fractional changes ~ 0.35 ) in the N. Atlantic hurricane
frequency response to warming (Figure 30a). For example, the two Hadley Center models
produce the largest decrease while the ECHAM5 model generates a modest increase of
hurricanes. In contrast, the two GFDL CMIP5 models tend to produce an increase especially in
the near decade (2026-2035), and in the CM3 projections for RCP4.5. However, for RCP8.5,
both the CM3 and ESM produce insignificant change at the late 21st century.

Zhao and Held (2012) found that most of the inter-model spread in the N. Atlantic
hurricane frequency response among the CMIP3 models can be explained by a simple relative
SST (RSST) index defined as the tropical Atlantic SST minus tropical mean SST. Under global
warming scenarios the SST difference between the MDR and the other tropical ocean basins
varies from model to model with implications for hurricane activity (Latif et al. 2007; Swanson
2008; Vecchi et al. 2008; Wang and Lee 2008; Xie et al. 2010). The RCP4.5 projections for both
near decades and late 21st century from the CM3 and ESM models show consistent relationship
between the N. Atlantic hurricane frequency and the RSST (Figure 31a). However, the results
from the two late 21st century runs with RCP8.5 show a marked departure from the regression line associated with the largest reduction (38% for CM3-2090-RCP8.5 and 25% for ESM-2090-RCP8.5) in global mean hurricane frequency. This departure is most likely a result of the larger direct effect of the atmospheric greenhouses-gases concentration (RCP8.5) that can suppress global and regional TC/hurricane frequency and therefore shift the hurricane frequency-RSST regression line downward (Held and Zhao 2011). In general, the CMIP5 downscaling results continue to suggest a large uncertainty in future projections of N. Atlantic hurricane frequency, consistent with the analysis in Section 5a and 5b.

The GFDL CMIP5 downscaling results tend to produce a reduction in E. Pacific hurricanes (Figure 30b, 31b). A negative correlation generally exists between the response of E. Pacific and the N. Atlantic hurricane frequency. When the fractional changes are plotted against the E. Pacific RSST index, we also see a strong correlation between E. Pacific hurricane frequency and E. Pacific RSST. Again, the departure for the two RCP8.5 models at the late 21st century supports that the global mean reduction due to the direct effect of GHG tends to systematically move the regression line downward.

d. Analysis of Atlantic wind shear and relative SST changes

As discussed in Section 5c, at least to the middle of the 21st Century models indicate a strong relationship between Atlantic hurricane activity and Atlantic RSST. Other studies suggest the strong influence of Atlantic vertical shear (VS) variations in future climate on tropical cyclone activity (e.g. Vecchi and Soden 2007). Here we assess the trends over the 21st Century in Atlantic main development region (MDR; 85°W-15°W, 10°N-20°N) RSST and VS in CMIP5.
models for RCP4.5. A broader set of CMIP5 models is used here than that used in Section 5c, so as to obtain a more complete picture of shear and RSST changes across the ensemble. Recall that RCP4.5 features a ramp up in radiative forcing to 4.5 W m\(^{-2}\) at 2050, and stabilization thereafter, and hence the SST change and radiative forcing is more muted than that for the end of century RCP8.5 time slice experiment discussed in Figure 31 where the RSST change becomes less strongly related to Atlantic tropical cyclone activity.

A scatterplot of the MDR RSST and VS trends in the 21\(^{st}\) century CMIP5 simulations is shown in Fig. 32. The individual models show different responses of the MDR RSST and VS trends, and hence suggest inconsistent changes in tropical cyclone activity, consistent with the conclusions in Sections 5a-c. The linear fit of all models (the solid line in Fig. 32) shows that the VS trend decreases with the MDR SST trend. This is consistent with a recent modeling study (Lee et al. 2011), showing that a slower warming in the tropical North Atlantic compared with the tropical Indo-Pacific Oceans increases the VS and static stability in the MDR for Atlantic hurricanes.

7. Interannual to Decadal Hydroclimate

a. Interannual variability

The large internal variability of the El Niño/Southern Oscillation (ENSO) makes it difficult to detect changes in ENSO properties over periods of a couple of centuries (Wittenberg 2009, Stevenson et al. 2012). Even without significant changes in ENSO itself, the long-term trend in tropical SSTs and the associated changes in the midlatitude base state may result in
substantial changes in ENSO-related teleconnections (Meehl et al. 2006, Meehl and Teng 2007, Lau et al. 2008, Kumar et al. 2010, Stevenson et al. 2012). In this analysis, we evaluate whether the NA seasonal temperature and precipitation patterns associated with ENSO are projected to change significantly in the 21st century and beyond in CMIP5 models in RCP4.5 and RCP8.5. It should be noted that CMIP5 model performance in accurately simulating ENSO teleconnections in current climate is relatively poor (Sheffield et al. 2012b). We do not distinguish different types of ENSO episodes, such as CP from EP El Niño episodes that may change in proportion with climate change (Yeh et al. 2009), but rather consider a single broad class of ENSO events identified by the Niño3.4 SST index. We identify ENSO episodes in the same way as described in the historical analysis (Sheffield et al. 2012b).

On the basis of these calculations, we find that the frequency of ENSO episodes in the RCP4.5 and RCP8.5 simulations is approximately the same as in the historical simulations. A hint of increased ENSO amplitude exists, as the ensemble and episode mean peak Niño3.4 SST amplitude of El Niño episodes for the historical, RCP4.5, and RCP8.5 episodes is 1.42, 1.54, and 1.62°C, respectively; for La Niña episodes the mean peak amplitude is -1.32, -1.42, and -1.54°C, respectively. However, large inter-model variability exists, so these differences are not robust.

We calculate SAT and precipitation composites and assess significant differences in the same way as in the historical analysis and described in the Figure 33 caption (Sheffield et al. 2012b). Overall, we find that some models project significant changes in ENSO-related SAT and precipitation patterns, but most do not indicate significant changes in the 21st century. Moreover, because pattern changes vary from model to model, these changes are not robust in the ensemble mean. As an example, Figure 33 illustrates which models project significant SAT and
precipitation patterns in the winter of the mature ENSO episode [designated as DJF(0/1)]. The two HadGEM2 models project significant SAT and precipitation pattern changes in the 21st century, but most other models do not. For the models with simulations that extend beyond 2100, a tendency exists for significant changes from 2101-2300. In the ensemble calculations (bottom row), however, all pattern changes are not statistically significant.

b. Decadal variability

Similar to our analysis for ENSO, we evaluate whether NA temperature and precipitation patterns associated with the Pacific Decadal Oscillation (PDO) are projected to change in the CMIP5 models. We define the PDO as the leading empirical orthogonal function of extended winter (November-April) monthly mean sea surface temperature (SST) anomalies in the North Pacific poleward of 20°N. We follow the same procedures for calculating SST anomalies, performing the EOF analysis, and calculating the PDO index as described in the historical analysis (Sheffield et al. 2012b), except that we now base the EOF analysis on the period from 2006-2090. Sheffield et al. (2012b) showed that CMIP5 models have success in reproducing the PDO temperature teleconnection pattern, but mixed success at capturing the teleconnection pattern in precipitation.

Overall, we find that SAT and precipitation pattern changes for individual models are generally not significant, but some robust changes are apparent in the ensemble mean (Figure 34). The most notable change is found in spring (MAM) in the 21st century in RCP8.5 in temperature. Most models do not project significant SAT or precipitation pattern changes in any epoch, which probably relates to the limited sample size and the relative stability of the PDO-
related variability. However, in the ensemble mean of RCP8.5, where the sample size is enhanced, the change of the 21\textsuperscript{st} century SAT pattern is statistically significant at the 99\% level. Significant changes are also indicated in summer and fall of the 21\textsuperscript{st} century, but as described below, the magnitude of the statistically significant change is strongest in spring.

To determine the nature of the projected changes we show in Figure 35 the ensemble mean MAM SAT regressions for the historical run and RCP8.5. Figure 35 indicates a clear weakening of the PDO-related SAT variability in the warmer climate, with largest weakening over western NA. We find that the PDO-related North Pacific SST variability does not show a similar weakening, however. This finding suggests that increased ocean stratification, which increases the SST response to a given a wind forcing, and/or the decrease in mean temperature gradients may be responsible for the weakened PDO response. This weakening tendency is also evident in summer and fall, but of weaker magnitude. Similar changes in winter also are significant, but only after 2100.

8. Conclusions

We have examined 21\textsuperscript{st} Century projections of NA climate in CMIP5 models under the RCP4.5 and RCP8.5. In terms of robust projected changes, models indicate that warming will occur everywhere in NA, with the greatest warming occurring in northern Canada where multimodel mean temperature increases of 15\textdegree C can be found during boreal winter in RCP8.5 at the end of the 21\textsuperscript{st} Century. It is very likely that the negative temperature trend (warming hole) in the eastern U.S will disappear in the 21\textsuperscript{st} Century. Models agree that more precipitation will
fall north of 40 degrees in winter, with the largest increases along the West coast and northeast U.S. and Canada. The potential exists for a large increase in the number of extreme rainfall events over the Northeast U.S., since many of the CMIP5 models for RCP8.5 suggest a 5-6 fold increase in increase heavy precipitation events by the late 21st century. Model agreement is good on reduced summertime precipitation in the Caribbean and Southern Mexico. Projected precipitation increases are robust across models for Alaska and the Arctic rim during boreal summer.

Further robust projected changes include a decrease in snow water equivalent throughout the year, with the greatest decreases in January-April. Especially large decreases are noted in the Rockies and Northwest Canada. The growing season is projected to lengthen across NA by the end of the 21st Century, with the largest changes in western U.S. and Northern Mexico where increases of two months or more are anticipated. In the agriculturally important regions of the central U.S. and Canada, increases of 3-6 weeks are projected. Frost days are projected to decrease across NA. An increase in the frequency of extended wet periods is projected for the Eastern U.S.. Warm temperature extremes are projected to increase across NA, with the number of 90 degree days to increase in the Midwest and Northeast U.S. by 50-100% relative to the current climatology. Models that can resolve the Great Plains low level jet in current climate project an increase in strength of the jet in the 21st Century. A poleward shift of the Atlantic storm track is evidenced in the upper troposphere during all seasons. During winter, a general decrease of storm track activity, especially in the Southern half of the Atlantic storm track, is projected. Models tend to agree on a drier early monsoon season (June and July) and wetter late season (September and October), consistent with a robust strengthening of the midsummer
drought in Central America and the greater Caribbean region. Where comparisons are done, RCP4.5 changes are often a more muted version of RCP8.5 change. One exception is in hurricane activity, where the direct effect of greenhouse gases on TC/hurricane frequency appears to become increasingly important relative to other factors (such as RSST) in the RCP8.5 at the end of the 21st Century.

Although many projected changes in NA climate are robust across CMIP5 models, substantial disagreement in some areas helps to define priorities for future research. The sign of mean precipitation changes across the southern U.S. is inconsistent among models. Models disagree on annual mean precipitation changes in the NA monsoon region. Models disagree on snow water equivalent changes on a regional basis, especially in transitional regions where competing effects occur due to greater snowfall and warming temperatures. In the southeastern U.S., the multimodel mean diurnal temperature range (DTR) signal is rather weak, accompanied by larger variance among the models. The western U.S. is characterized by large intermodel variability in the number of frost days in the Western U.S., where multimodel mean decreases in frost days (greater than 70 days in RCP8.5) are also largest. Models do not agree on how intraseasonal variability will change over the Caribbean and Gulf of Mexico, which may have implications for future modulation of hurricane activity. Projected changes in seasonal mean Atlantic and east Pacific tropical cyclone activity are inconsistent among models, which disagree on the sign and amplitude of changes in environmental factors that modulate hurricane activity. Models are highly inconsistent in projecting how the ENSO teleconnection to NA will change.

Model success in simulating historical climate as shown in Parts 1 and 2 of this study (Sheffield et al. 2012a,b) lend confidence to many of the projected results. However, model
biases in other areas decrease confidence in projections, including changes of the timing of NA
monsoon precipitation that exhibits projected changes in the same sense as historical biases.
Growing season length is projected to increase most strongly over along the West Coast, where
models tend to display large negative biases in historical runs. Models have substantial
difficulties in simulating the historical distribution of persistent drought and wet spells, which
produces less confidence in the pattern of extreme precipitation event changes in future climate.

This study represents an initial effort by the NOAA MAPP CMIP5 Task Force to provide
an assessment of projected NA climate change over the 21st Century, although the analysis here
just touches on what is available in the CMIP5 archive. More in-depth analysis of NA climate
change is contained elsewhere in this special collection, and analysis of NA climate change in
the CMIP5 archive will be expanded in years to come.

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Table 1. CMIP5 models evaluated and their attributes.

<table>
<thead>
<tr>
<th>Model</th>
<th>Center</th>
<th>Atmospheric Horizontal Resolution (lon. x lat.)</th>
<th>Number of model levels</th>
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Figure 1. CMIP5 15 member multi-model, multi-run ensemble mean precipitation change for RCP8.5 for 2070-2099 relative to 1961-1990 base period for December-February (DJF) and June-August (JJA). Models used: BCC-ESM-1, CanESM2, CCSM4, CNRM-CM5, CSIRO, FGOALS-S2, GFDL-CM3, GISS-E2-R, HADGEM2-CC, INMCM4, IPSL-CM5A-MR, MIROC5, MPI-ESM-LR, MRI-CGCM3, NORESM1-M. Red line is the 4 mm/day 1961-1990 climatology. For models that have more than one run, the average is taken over runs prior to taking the multi-model mean.
**Figure 2.** DJF and JJA plots of model agreement on sign of end-of-century precipitation change for the CMIP5 RCP8.5 scenario for the years 2070-2099, relative to a base period of 1961-1990. Red colors indicate a higher number of models (out of 15) that agree on a negative precipitation change; blue colors imply agreement on a positive precipitation change (shown as few models with negative change). Agreement is calculated on the sign of change alone; areas with fewer than 10 models agreeing on sign are unshaded.
Figure 3. As in Figure 1, except for surface temperature change.
Figure 4. Mean JJA (left) and DJF (right) moisture transport integrated vertically to 500 hPa (VIMT shown as vectors, Kg/ms), and moisture divergence (color contours, Kg/m²s x10⁴) computed from four coupled models from CMIP5 (a-h) 2081-2100 minus 1981-2000 from the RCP8.5 and Historical experiments. The models used are CanESM2, CNRM-CM5, GFDL-ESM2M, and MIROC5, for which one realization of the required 6-hourly fields were available. This figure should be compared with the corresponding analysis from Sheffield et al. 2012a.
Figure 5. 30-year means from the historical (1971-2000) and RCP8.5 experiment (2071-2100) for regionally averaged runoff and evapotranspiration (mm/day). Six regions were defined for the North American continent: Central America, Western North America, Central North America, Eastern North America, Alaska/Northwest Canada, and Northeast Canada. The circles represent individual climate models. The large triangles represent the multi-model values. Precipitation balances runoff plus evapotranspiration over decadal time scales by assuming that there is no change in water storage. The diagonal lines represent contours of precipitation. A shift in the multi-model mean towards the upper-right indicates an increase in precipitation. A shift to the bottom-left indicates a decrease in precipitation. Values are calculated for 13 climate models (CanESM2, CSIRO-Mk3-6-0, GFDL-ESM2G, GISS-E2-H, GISS-E2-R, IPSL-CM5A-LR, IPSL-CM5A-MR, MIROC-ESM, MIROC-ESM-CHEM, MPI-ESM-LR, MRI-CGCM3, NorESM1-M, NorESM1-ME) using one ensemble member each.
Figure 6. Changes in SWE (mm) from 9 CMIP5 models (one ensemble member each) from 1971-2000 to 2071-2100 for the RCP8.5 scenario. (top) Mean monthly change in SWE averaged over N. America (25N-80N, 170W-65W), and (bottom) spatial distribution of change in winter-spring (NDJFMAM) SWE (shading) and coefficient of variation (CV) of changes in SWE across models (contours). Some of the models have spuriously high snow accumulations at isolated grid cells and these are filtered out.
Figure 7. Projected changes in growing season length for 6 CMIP5 models (CCSM4, CSIRO-Mk-6-0, HadGEM-ES, IPSL-CM5A-LR, MPI-ESM-LR, NorESM1-M; all single ensemble member r1i1p1) for RCP8.5. Changes are calculated as the difference between the mean growing season length for 2071-2100 and 1971-2005. We define the growing season length following Schwartz et al. (2006), which is the number of days between the last spring freeze of the year and the first hard freeze of the autumn in the same year. A hard freeze is defined when the daily minimum temperature drops below -2°C. The end of century changes are calculated based on the change in 30-yr average growing season length (2071-2100 - 1971-2005).
Figure 8. Change in the annual number of frost days (Tmin < 0°C) between 1979-2005 and 2071-2100 for the RCP8.5 scenario from six CMIP5 models (CSIRO-Mk3-6-0, MPI-ESM-LR, NorESM1-M, HadGEM-ES, CCSM4, IPSL-CM5A-LR). The first ensemble member is used for each model. Also shown are the MME mean change and the standard deviation of the change across the models. Frost days were calculated on the model grid and then the results interpolated to 2.0° resolution. Frost days are calculated as the annual number of days with Tmin less than 0°C (Frick et al., ...
2002) and then averaged over 1979-2005 for the "historical" and 2071-2100 for the RCP8.5 scenario.
Figure 9. Trend of daily temperature range ($T_{\text{max}} - T_{\text{min}}$) during 2006-2055 averaged among 22 models with 63 members in the RCP45 experiment. The contours are the inter-model variance of the range.
Figure 10. The difference in frequency of occurrence for RCP4.5 projection runs calculated using the RCP4.5 projected climatology minus historical runs calculated using the historical climatology defined by SPI6 averaged over positive (wet) events for (a) CAN, (b) CCSM4, (c) CSIRO, (d) IPSL, (e) MPI. (f)-(j) same as (a)-(e) but for negative (dry) events. Meteorological drought is measured by precipitation (P) deficit, and the index used to classify drought is the 6-month Standardized Precipitation Index (SPI6). Agricultural drought is measured by total soil moisture (SM), and the index used for drought classification is the SM percentiles [Mo 2008]. The SPI6 is computed by following the method outlined by Mckee et al. [1993, 1995]. The SM anomalies are normally distributed so SM percentiles are computed.
based on a normal distribution function with parameters determined for each climate model. The Frequency of Occurrence (FOC) is the number of extreme events that last at least 9 months divided by the total number of events. An event is defined as extreme if the SPI6 reaches the threshold of +/- 0.8 or the SM percentile is above (below) 80%(20%).
Figure 11. Climatological distribution of the number of days per year when temperature is greater than 90°F over US and Mexico for the period of 1979-2005 derived from 9 CMIP5 models and Global Historical Climatology Network daily maximum surface temperature, as well as model projections for RCP4.5 and RCP8.5 for 2071-2100. $T_{\text{max}}$ were mapped to a 2.5°x2.5° grid.
Figure 12. a) Solid contours: Winter (DJF) climatological (1980-1999) storm track activity, as indicated by variance of 24-hour difference bandpass filtered v at 250 hPa level, based on ensemble mean of 16 CMIP5 models (contour interval 100 m²s⁻²). The filter used is the 24-hour difference filter (Wallace et al. 1988), which highlights synoptic variability with periods of 1.2 to 6 days. The models examined are BCC-CSM1.1; CanESM2; CNRM-CM5; CSIRO-Mk3.6.0; GFDL-ESM2M; HadGEM2-CC; HadGEM2-ES; INM-CM4; IPSL-CM5A-LR; IPSL-CM5A-MR; MIROC5; MIROC-ESM; MIROC-ESM-CHEM; MPI-ESM-LR; MRI-CGCM3; and Nor-ESM1-M. Colored lines: Projected change (2081-2100 mean minus 1980-1999 mean) based on RCP8.5 (contour interval 20 m²s⁻²). Shades: grid boxes over which more than 80% of CMIP5 models agree on the sign of the projected change; b) Same as a), but for summer (JJA; contour intervals 50 m²s⁻² and 10 m²s⁻²); c) Same as a), but for 500 hPa level (contour intervals 50 m²s⁻² and 10 m²s⁻²); d) Same as c), but for JJA (contour intervals 25 m²s⁻² and 5 m²s⁻²); e) Same as a), but for variance of SLP (contour intervals 20 hPa² and 5 hPa²); f) Same as e), but for JJA ((contour intervals 10 hPa² and 2.5 hPa²).
Figure 13. (a) Cyclone track density for the mean (shaded) and spread (contoured every 0.3) for 15 CMIP5 models (listed in Colle et al. 2012) showing the number of cyclones per cool season (November to March) per 50,000 km$^2$ for 1979-2004. Same as (a) except the difference in cyclone density (Future – historical) and percentage change between (b) 2009-2038, (c) 2038-2068, and (d) 2069-2098 and the historical 1979-2004 period. The models used to generate the ensemble mean include the CESM, EC-EARTH, MRI-CGCM3, CNRM-CM5, MIROC5, HadGEM2-ES, HadGEM2-CC, INMCM4, IPSL-CM5A-MR, MPI-ESM-LR, NorESM1, GFDL-ESM2M, IPSL-CM5A-LR, BCC-CSM1, and MIROC-ESM-C.
Figure 14. (a) Number of cyclone central pressures at their maximum intensity for the 1979-2004 cool seasons within the dashed box region in Fig. 14a for a 10 hPa range centered every 10 hPa showing the CMIP5 mean historical (black), (b) 2009-2038 (green), (c) 2038-2068 (black), and (d) 2068-2098 (red) years for the 15 CMIP5 models in Figure 14. (b) Same as (a) except the difference and standard deviation between three future periods and the historical 1979-2004 period.
Figure 15. (a) Multimodel mean daily precipitation (mm/season) averaged for select CMIP5 members for the historical (1979-2004) period during the cool season (November – March). (b) Precipitation difference (shaded in mm/season) and the percentage change (solid every 10 percent) between the (b) 2009-2038 and the historical 1979-2004 period for the cool season (November – March). (c) Same as (b) except for the 2069-2098 period. (d) Difference in the number of precipitation days and percentage change for each amount bin between the 2009-2038, 2038-2068, and 2069-2098 and the historical 1979-2004 period for the land area only in the black box in (b). The models analyzed include BCC-CMS, GFDL-ESM2M, HadGEM2-CC, HadGEM2-ES, INMCM4, IPSL-CM5a, MIROC5-ESM, MICRO5, and NorESM.
Figure 16. Average April SWE (mm) from 9 CMIP5 models for (a) 1971-2000, (b) 2071-2100 (RCP85) and (c) their difference and ratio. The models are: CanESM2, CCSM4, CNRM-CM5, GFDL-ESM2M, GISS-E2-R, inmcm4, MIROC5, MIROC-ESM, MIROC-ESM-CHEM.
Figure 17. Bars represent the (a) RMS and (b) annual mean differences between the future (2071-2100) and historical (1979-2005) periods for the core NAM region (23°N-30°N, 105°W-110°W) for RCP4.5.
Figure 18. Precipitation annual cycle averaged for longitudes representing the North American Monsoon (102.5 to 115. W), for the 1986-2005 climatological period from CMAP precipitation, version 2 from Xie and Arkin 1997 (a) and the ensemble mean of 13 CMIP5 models (b), in units of mm/day with thick black lines identifying contours greater than 4 mm/day. Multi-model climatology contours (black lines) are reproduced in (c) which also shows ensemble mean precipitation % difference (colors). Areas of significant change are stippled. The models employed in the analysis are CCSM4, CNRM-CM5, CSIRO-Mk3, CanESM2, GFDL-CM3, GFDL-ESM2G, GISS-E2-R, IPSL-CM5A-LR, MIROC-ESM, MIROC5, MPI-ESM-LR, MRI-CGCM3, NorESM1-M.
Figure 19. Climatology of LLJ strength as a function of month between 1971-2000 for NCEP-NCAR reanalysis and the CCSM4, GFDL, GISS, HadESM2-CC, and MPI-ESM models. The LLJ was computed as the 925 hPa meridional wind averaged over 27.5-32.5N and 95-100W.
Figure 20. Monthly climatology of 925 hPa meridional wind averaged over 27.5-32.5N and 95-100W for 1971-2000, 2035-2064, and 2071-2100 for RCP4.5 and 8.5 for the CCSM4, HadGEM2-CC and MPI ESM models. The monthly climatology from NCEP-NCAR reanalysis during 1971-2000 is shown on each plot.
Figure 21. CMIP5 multi-model ensemble mean September Northern Hemisphere sea ice extent from 1900 to 2100, based on the historical (gray) and future RCP4.5 (blue) and RCP8.5 (red) emission scenarios. The observed ice extent from 1953-2011 is shown in black and the shading represents ± one standard deviation of the multi-model ensemble means. In deriving the multi-model ensemble mean for each emission scenario, only models that have at least 75% of their distribution of September ice extent within the observed range of 6.13 to 8.43x10^6 km^2 from 1953-1995 are included. The rejected models include CanESM2, GISS-E2-R, GISS-E2-H, CSIRO Mk3-6 and FGOALS-g2, resulting in a total of 16 models for the historical and RCP4.5 scenario and 14 models for the RCP8.5 scenario.
Figure 22. Winter and summer temperature trend in RCP8.5 for two time periods: (1) 2015-2050, (2) 2051-2098. Multi-model average is calculated from a total of 39 ensemble members from 16 climate models. Numbers in each panel show global mean temperature trend (land only, 60S to 60N).
Figure 23. RCP 8.5 (a) 30 years running trend in East USA (defined in Figure 17) DJF Temperature Trend (b) 30 year running Trend difference between East and West USA. (c) and (d), same as (a) and (b), but for RCP4.5. The East and West USA are defined in Figure 16. Shaded region show 95% uncertainty range calculated from 39 ensemble members from 16 models in RCP 8.5 case, and from 22 models (ensemble average if model has more than one ensemble) in RCP 4.5 case. X-axis represents start of the 30 year period. For example, the trend corresponding to 2030 represent trend from 2030 to 2059. Five core models are also shown as an example, only the first ensemble member from each model in RCP 8.5 case, and ensemble average of each model in RCP 4.5 case.
Figure 24. CMIP5 multimodel mean projection of precipitation change (mm/day) during each of the summertime months (June-September) based on the mean of the last two decades of the 21st century (2080-2099) of the RCP8.5 forcing experiment minus the mean over the historical experiment (1860-2005). The numbers of models included in the analyses are 23 and 20 for Historical and RCP8.5, respectively.
Figure 25. CMIP5 multimodel mean projection of the MSD (mm/day) averaged over the historical experiment, the RCP4.5 forcing experiment (2080-2099), and the RCP8.5 forcing experiment (2080-2099). The numbers of models included in the analyses are 23, 17, and 20 for Historical, RCP4.5, and RCP8.5, respectively.
Figure 26. The leading complex EOF amplitude of 30-90 day precipitation anomalies during boreal summer during 1981-2005, 2076-2099 in RCP8.5, and the difference for the CCSM4, HadGEM2_CC, HadGEM2_ES, and MPI_ESM_LR models. The total variance explained by the leading CEOF mode for the end of the 20th Century and 21st Centuries are given as follows: CCSM4 (12.1%, 14%), HAD_GEM2_CC (22.1%, 25.7%), HAD_GEM2_ES (20.5%, 21.4%), MPI_ESM (25.1%, 26.4%). The color bar for the actual CEOF amplitudes are shown on the right, and the color bar for the difference is shown on the bottom. The time series of the CEOFs were normalized, and so amplitude information is contained in the spatial pattern.
Figure 27. CMIP5 7-model mean track density for the RCP 4.5 future (May-Nov 2070-2100) projections. Contours every four density units starting with two. Also shown are the seven CMIP5 track density differences for the future minus historical (May-Nov 1979-2005) period. Colored areas indicate differences exceeding one standard error of the historical mean for that model. Contour lines are every one density unit with no zero contour shown.
Figure 28. Left panels: mean track density in the North Atlantic for the historical simulation in the period 1951-2000 for the GFDL-ESM2M (1 ensemble member all pathways), MPI-ESM-LR (3 ens. Members all pathways), MRI (5 ens. members historical, 1 ens, member RCP4.5 and RCP8.5), and MIROC5 (1 ens. member historical, 3 ens. members RCP4.5, 1 ens. member RCP8.5) models. The difference of the track density in the period 2051-2100 and historical track density is shown in the middle panels for RCP45 and the right panels for RCP85. The Camargo and Zebiak (2002) method defines the existence of a TC based on low-level vorticity, sea level pressure, definition of a warm core based on temperature anomalies, and surface wind speed. The storms also have to last at least 2 days to be considered in our analysis.
Figure 29. Box-plot distributions of the number of tropical cyclones (NTCs) per year in each model in the 3 different scenarios for the last 50 years in each century: historical (Hist, 1951-2000), RCP45 and RCP85 (2051-2100). The box denotes the range of the 25th and 75th percentiles of the distributions, while the median is marked by a red line. The values outside of the middle quartile are marked by whiskers and red crosses. We calculated the statistical significance of differences in the distributions of the number of TCs in the present and future using a Wilcoxon rank sum test for medians.
Figure 30. a) Fractional change in N. Atlantic hurricanes frequency downscaled using the GFDL C180HIRAM. Blue bars show results using the CMIP3 models projected SST warming anomalies (and their ensemble mean) at the late 21st century (2080-2100 relative to 2000-2020) for the A1B scenario. The control experiment was integrated for 20 years while most of the warming experiments were carried out for 10 years due to constraints of computer time. Red, green and black bars show results using GFDL CMIP5 model (CM3 and ESM) projected SST warming anomalies (CM3-2030-RCP4.5 and ESM-2030-RCP4.5: 2026-2035 averaged SST anomalies from CM3 and ESM RCP4.5 experiments with radiative gases at RCP4.5 2026-2035 values. CM3-2090-RCP4.5 and ESM-2090-RCP4.5: 2086-2095 averaged SST anomalies from CM3 and ESM RCP4.5 experiments with radiative gases at RCP4.5 2085-2095 values. CM3-2090-RCP8.5 and ESM-2090-RCP8.5: As in CM3-2090-RCP4.5 and ESM-2090-RCP4.5 experiments except using RCP8.5 model projected SST anomalies with
radiative gases at RCP8.5 2086-2095 values. The GFDL C180HIRAM present-day experiments contain a 3-member ensemble simulation for the period of 1981 to 2008. The CMIP5 SST anomalies are computed relative to 1981-2008 average. b) As in a) except for the E. Pacific.
Figure 31. a) the fractional change in N. Atlantic hurricane frequency against changes in a relative SST index defined as the Atlantic Main Development Region (MDR) [80°W-20°W, 10°N-25°N] SST minus tropical mean [30°S-30°N] SST in ASO season. b) As in a) except for the E. Pacific, the E. Pacific Main Development Region is defined as [160°W-100°W, 7.5°N-15°N].
Figure 32. Scatterplot plot of the vertical wind shear trend versus the relative SST trend in the hurricane main development region (MDR; 85°W-15°W, 10°N-20°N) during the Atlantic hurricane season of June to November (JJASON). The relative MDR SST trend is calculated as the difference between the MDR SST trend and the global tropical (30°S-30°N) SST trend. SST is from the NOAA extended reconstructed SST version 3 (Smith et al. 2008).
Figure 33. Field significance of the changes in (a) El Niño and (b) La Niña SAT (left columns) and precipitation (right columns) composites in CMIP5 models (rows) from the historical simulations to different epochs under two emissions scenarios. An El Niño (La Niña) episode is defined as all times such that the three-month running mean Niño3.4 SST, detrended and high-pass filtered in this case, is greater than 0.5°C (less than -0.5°C) for at least five consecutive three-month running seasons. The model names are given at the left, and the bottom row corresponds to the CMIP5 ensemble. Each column corresponds to a different variable, emissions scenario, and epoch, where the first number above the column indicates the emissions scenario and the second number indicates the epoch (e.g., “4.5/21” indicates RCP4.5 in the 21st century and “4.5/22+23” indicates RCP4.5 from 2101-2300). We determine that the model patterns are significantly different from the observed patterns if the patterns are statistically distinguishable in a test of field significance based on the “false discovery rate” (FDR) (Benjamini and Hochberg 1995, Wilks 2006). Light, medium, and dark red correspond with 90%, 95%, and 99% significance levels, respectively, blue corresponds with statistically insignificant ($\alpha_{global} > 0.1$) changes, and white indicates no available data.
**Figure 34.** Field significance of the changes in PDO SAT (left columns) and precipitation (right columns) regressions in CMIP5 models (rows) from the historical simulations to different epochs under two emissions scenarios. The model names are given at the left, and the bottom row corresponds to the CMIP5 ensemble. Each column corresponds to a different emissions pathways and epoch, where the first number above the column indicates the emissions scenario and the second number indicates the epoch (e.g., “4.5/21” indicates RCP4.5 in the 21st century and 4.5/22+23 indicates RCP4.5 from 2101-2300). To assess significance, we regress detrended SAT and precipitation upon model PDO indices and then evaluate whether regression pattern changes are significant based on an FDR. Light, medium, and dark red correspond with 90%, 95%, and 99% significance levels, respectively, blue corresponds with statistically insignificant ($\alpha_{global} > 0.1$) changes, and white indicates no available data.
**Figure 35.** Ensemble regression of SAT anomalies with the PDO index during MAM in (a) historical simulations and (b) RCP8.5 simulations between 2006 and 2100, and (c) the difference between (a) and (b). The contour interval is 0.1°C in (a) and (b) and 0.05°C in (c). Stippling in (c) indicates where the differences are statistically significant at the 95% confidence level.