Uncertainty in future projections of the North Pacific subtropical high and its implication for California winter precipitation change

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Abstract

This study examines future projections of sea level pressure change in the North Pacific and its impact on winter precipitation changes in California. The multi-model analysis, based on the Coupled Model Inter-comparison Project phase 5 (CMIP5) models under the Representative Concentration Pathway 8.5 (RCP8.5) scenario, shows a robust sea-level pressure change in the late 21st century over the western North Pacific in which both the Aleutian Low and North Pacific Subtropical High (NPSH) shift poleward in concert with a widening of the Hadley Cell. This change is largely explained by a systematic increase of static stability in the subtropics. However, over the eastern North Pacific, the projected NPSH changes exhibit a substantial inter-model spread, resulting in uncertain projections of precipitation changes in California. This inter-model spread is associated with a Pacific Decadal Oscillation-like surface temperature change over the western North Pacific and the resulting meridional temperature gradient change. This result suggests that a better prediction of wintertime precipitation changes over the West Coast of North America in future climates may need a reduced uncertainty of atmosphere–ocean coupling in the western North Pacific.
1. Introduction

Projected changes in regional and global precipitation in a warming climate have been extensively documented in the literature because of their socio-economic importance (e.g., Held and Soden 2006; Meehl et al. 2007). Based on both Coupled Model Inter-comparison Project phases 3 and 5 (CMIP3 and CMIP5) model simulations, the Fifth Assessment Report (AR5) of the Intergovernmental Panel on Climate Change (IPCC) recently reported that the hydrological contrast between wet and dry regions and between wet and dry seasons will likely increase in a warmer climate (IPCC 2013). In terms of latitudinal distribution, this change, often referred to as the “rich-get-richer” mechanism, is characterized by more precipitation in the deep tropics and mid-latitudes but less precipitation in the subtropics (e.g., Held and Soden 2006). These projected precipitation changes can be qualitatively described by the Clausius–Clapeyron relation, which relates the increase in water vapor with increasing air temperature (e.g., Manabe and Wetherald 1975; Held and Soden 2006). However, precipitation changes are also known to be influenced by large-scale atmospheric circulation such as the Hadley Cell (HC) and extratropical storm tracks (e.g., Lu et al. 2007; Scheff and Frierson 2012a).

The response of the atmospheric general circulation to the increase in anthropogenic greenhouse gases is often characterized by a poleward shift of the extratropical storm tracks and an expansion of the HC (e.g., Gerber and Son 2014). These changes lead to a poleward shift of mid-latitude rain belts and a widening of the subtropical dry zone (Scheff and Frierson 2012a,b; Barnes and Polvani 2013; Neelin et al. 2013). Among them, the projected dryness in the subtropics associated with the HC expansion has been primarily attributed to the increase of static stability in the subtropics with a non-negligible contribution of the increased meridional
temperature gradient in mid-latitudes (e.g., Lu et al. 2008). In a warmer climate, static stability is anticipated to increase as a result of the quasi-moist adiabatic temperature adjustment to the surface warming (e.g., Frierson 2006). This can stabilize the baroclinicity in the subtropics resulting in a poleward displacement of the baroclinic instability and the related eddy-driven subsidence in the subtropics.

Many previous studies (e.g., Frierson et al 2007, Hwang et al 2013, Seo et al 2014) have focused on zonal-mean circulation and hydrological changes. However, for regional climate, zonally asymmetric changes are more important. In general, regional climate change is less robust than zonal-mean or global-mean climate changes. This statements becomes particularly true for locations downstream of the jet where baroclinic eddies are prevalent. Chang (2013) and Neelin et al. (2013) showed that future projections of atmospheric circulation changes downstream of the Pacific jet differ significantly among CMIP3 and CMIP5 models. Such differences in the projected circulation changes are inevitably reflected in the projected changes in surface temperature and precipitation downstream over land. For example, projected future precipitation change in California is known to include substantial uncertainty (Neelin et al. 2013).

California usually receives half of its annual precipitation between December and February. As of the writing of this paper, California is entering the fourth year of one of its worst droughts in recorded history. To evaluate the degree to which greenhouse gas-induced global warming would affect California precipitation during the winter season and the related uncertainties, the present study examines the atmospheric circulation change over the North Pacific basin by using state-of-the-art climate models. Particular attention is paid to regional changes in the subtropical ridge, or the nominal local HC boundary, which controls the moisture transport and precipitation
in the region. More specifically, future projections of the Aleutian Low (AL) and North Pacific Subtropical High (NPSH), which are the two dominant winter climate systems affecting the hydro-climate in the North Pacific, and their impact on precipitation changes over California are analyzed. Only in the boreal winter season (December–February; DJF), when both the AL and NPSH are well defined, is considered.

This paper is organized as follows. The data and methodology used in this study are described in section 2. The model performance in reproducing the climatological NPSH is first evaluated by comparing a multi-model ensemble (MME) with observations in section 3a. The future projections are documented in section 3b. In section 3c, the inter-model spread in the projected NPSH changes and its uncertainty are discussed for the western and eastern North Pacific separately. A discussion and the conclusions of this study are then presented in section 4.

2. Data and methodology

As listed in Table 1, 36 CMIP5 models that have archived both the historical and Representative Concentration Pathway 8.5 (RCP8.5) simulations are used in this study. For a fair representation, only one ensemble member is used for each model. All datasets are first interpolated onto a $2.0^\circ \times 2.0^\circ$ grid prior to analysis. The long-term climatology is then estimated as the average over the period of 1959–1999 from historical simulations. The projected future change is evaluated by computing the linear trend using a least-squares fit over the period of 2007–2099 from RCP8.5 simulations. For reference, the model results are compared with the observed sea level pressure (SLP) and precipitation derived from the Hadley Centre’s monthly
mean SLP (HadSLP2; Allan and Ansell 2006) and the Global Precipitation Climatology Project (GPCP; Adler et al. 2003), respectively. Here, GPCP climatology covers a period of 1979-1999, which is shorter than the other datasets.

It is often useful to define a climate index to concisely quantify atmospheric circulation. An example is the Northern Annular Mode index for the extratropical circulation in the Northern Hemisphere. In this study, regional circulation changes across the North Pacific are primarily quantified by changes in the latitudinal location of the NPSH. Specifically, the latitudinal location of maximum SLP, referred to as the subtropical ridge, is tracked in time and is referred to as the NPSH latitude in this study. Needless to say, many definitions are used for atmospheric circulations. Among them, the NPSH latitude can be easily computed without complication caused by transients related to the storm tracks and can be directly compared with long-term observations and different model results. More importantly, the NPSH latitude is dynamically related to the poleward edge of the HC, as discussed by Hu et al. (2011) and Choi et al. (2014). According to Choi et al. (2014), the NPSH latitude is defined as the latitude where $d\text{SLP}/dy=0$, indicating a location geostrophically consistent with zero zonal wind near the surface.

To address the regional asymmetry of the NPSH and its long-term changes, two longitudinally averaged regions were considered across the dateline which is the center of the AL; i.e., the western North Pacific ($140^\circ-180^\circ$E; hereafter, W–NPSH) and the eastern North Pacific ($180^\circ-120^\circ$W; hereafter, E–NPSH). The inter-model spread of the projected changes in W–NPSH and E–NPSH latitudes is then evaluated by analyzing the static stability and vertical wind shear changes over the North Pacific. To further characterize the relationship between the inter-model spread in the projected circulation changes over the eastern North Pacific and the
uncertainty in the projected changes in California precipitation, a singular value decomposition (SVD) analysis is also performed.

3. Results

a. Climatology

Figure 1a shows the climatological DJF SLP in the historical simulations. The model bias, which is the difference from observations (HadSLP2) over the time period of 1959–1999, is represented in the figure by shading. In general, except over the continents, the SLP distribution is effectively reproduced by the models, although the CMIP5 MME exhibits a slightly stronger NPSH and weaker AL than those in observations. Solid and dashed thick purple lines in Fig. 1a denote the latitudinal locations of the NPSH as defined by the location of maximum SLP from the MME and observations. The NPSH is located between 25°N and 32°N in the observations except near the West Coast of North America. The longitudinal distribution of the NPSH latitude is reasonably well reproduced by the models, particularly for W–NPSH. Although the simulated E–NPSH is biased equatorward, the difference between the CMIP5 MME and observations is not statistically significant.

The latitudinal locations of W–NPSH and E–NPSH in the individual models are further compared with observations in Fig. 2a. In the observations, denoted by the red dot, the W–NPSH, averaged over 140°–180°E, and the E–NPSH, averaged over 180°–120°W, are located at approximately 25.6°N and 27.5°N, respectively. As stated above, the CMIP5 MME reproduces these locations reasonably well; i.e., 25.6°N for W–NPSH and 27.1°N for E–NPSH. Although a
weak equatorward bias is found in E–NPSH, the inter-model spread is substantially large, and
the observed location of E–NPSH is within the uncertainty range of the model simulations.

It is evident in Fig. 2a that the W–NPSH and E–NPSH latitudes are significantly correlated,
with a correlation coefficient of 0.72, indicating a congruency between the upstream and
downstream bias in the subtropical high. However, the model-to-model spread of the E–NPSH
latitudes is larger than that of W–NPSH latitudes, with standard deviation of the former and latter
at 1.52 and 0.97, respectively. In other words, the spread in model bias increases downstream of
the Pacific jet where synoptic-scale eddy activity is more prevalent. This is expected from the
more vibrant eddy feedback to the mean flow downstream of the North Pacific storm track (e.g.,
Chang et al. 2002). This difference in model bias between the W–NPSH and E–NPSH latitudes
is not simply explained by a linear relationship (gray line in Fig. 2a). This may suggest that the
W–NPSH and E–NPSH latitudes are not controlled by exactly the same mechanisms.

Figure 1b presents MME precipitation and its departure from the observation in DJF. The
observed climatology is computed for the period of 1979–1999, whereas the MME is computed
for 1959–1999. Although not shown, the results are not sensitive to analysis period (e.g., MME
for 1979–1999). In Fig. 1b, two wet regions are identified in mid-latitudes. The first is over the
western North Pacific along the Kuroshio–Oyashio extension, and the second is along the
upslope of Rockies. Across the Kuroshio–Oyashio extension, a large amount of heat and
moisture are released, enhancing local precipitation and the Pacific storm tracks (e.g., Kwon et al.
2010). The MME overestimates this precipitation by more than 20%. Unlike the western North
Pacific, precipitation over the western United States and Canada is localized along the Rockies,
which is caused by uplifting by the windward slope of the Rockies on the incident storm tracks
from the Pacific Ocean (Chang 2013; Neelin et al. 2013). The MME shows a bias toward more
diffused distribution of precipitation toward the downstream of the Rocky Mountains, likely
resulting from the smoothed topography in the models and biases in atmospheric circulation (e.g.,
storm track activity) over the eastern North Pacific.

Focusing on the NPSH, we evaluate the relationship between California winter precipitation
and the location of the E–NPSH (Fig. 2b). Here, California precipitation is defined by averaging
the precipitation in the red box in Fig. 1b (135°–115°W, 32°–39°N), similar to Neelin et al.
(2013). The MME precipitation (about 3.00 mm/day) shows a wet bias in comparison to the
observation (about 2.22 mm/day). The model-to-model spread further indicates that California
precipitation in the model is highly correlated with the latitudinal location of the E–NPSH with a
correlation coefficient of −0.83 (Fig. 2b). This result explains about 70% of the variance. That is,
a negative bias in California precipitation in the model is related to a poleward bias of the E–
NPSH latitude. Similar analyses are also performed for the intensity of the E–NPSH (i.e.,
maximum value of SLP over the eastern North Pacific) and the latitudinal location of the AL;
their correlations with California winter precipitation are −0.68 and −0.52, respectively. This
result indicates a strong control of the location and intensity of the E–NPSH on the winter
precipitation over California in the models.

b. Future projections

Future changes in winter SLP and precipitation under the RCP8.5 scenarios are shown in
Fig. 3. The regions in which more than 66% and 90% of the models (≥24 and 32 models) have
the same sign are shaded and hatched, respectively. Contour lines represent the model-to-model spread quantified by one standard deviation about the MME. The robust SLP trends are found only over 140°E–160°W with a dipolar structure about 45°N (shading in Fig. 3a). A significant decrease in high-latitude SLP likely results from the fact that Arctic warming is predicted to be much stronger than the globally-averaged temperature change. On the poleward side of the NPSH, shown by the thick purple line in Fig. 3a, a relatively robust increase in SLP is also found. Near the continent, no robust change in SLP is found over the eastern North Pacific, where the magnitude of the inter-model spread is larger than the MME change.

Figure 4a shows the inter-model spread of projected changes in the NPSH in terms of the linear trends of W–NPSH and E–NPSH latitudes. As in the inter-model spread in the climatological latitudes (Fig. 2a), the projected changes in W–NPSH latitudes are closely linked to those of E–NPSH latitudes with a correlation coefficient of 0.71. Most models show a poleward shift in the W–NPSH (MME trend is 0.11°/decade). However, both poleward and equatorward shifts are found in the E–NPSH latitude trend. More importantly, the range of inter-model spread in E–NPSH latitude changes (-0.22 to 0.31) is much larger than that in W–NPSH latitude changes (-0.05 to 0.27). The net result is almost no change in the MME E–NPSH latitude (0.03°/decade). This result indicates that the uncertainty in the projected changes of NPSH latitudes increases downstream of the Pacific jet.

The red and blue circles in Fig. 4 indicate the MMEs of 20 and 16 models which show poleward (i.e., northward) and equatorward (i.e., southward) shifts of E–NPSH latitude, referred to as E–NPSH–N and E–NPSH–S. Their trends are 0.12°/decade and −0.08°/decade, respectively. This spread in the projected changes in E–NPSH latitude is largely consistent with the fact that
the poleward expansion of the Northern Hemisphere HC is less significant than that of the Southern Hemisphere HC in the future climate (e.g., Lucas et al. 2013).

The projected changes in precipitation (Fig. 3b, shading) show a significant increase in high latitudes but a decrease in low latitudes around the axis of maximum precipitation (Fig. 1b). The dipole pattern of precipitation change is commonly explained by the so-called thermodynamic response to global warming (Held and Soden 2006; Trenberth 2011). However, there exists a large model-to-model spread, particularly near the California region, as shown by the red box in Fig. 3b. Previous studies (Neelin et al. 2013, Scheff and Frierson 2012b; Gao et al. 2014) showed that future projection of California precipitation change is quite uncertain because of the geographical location of California and the large inter-model differences in dynamical response to global warming such as displacements of the Pacific jet, subtropical dry zone, storm tracks, and expansion of the HC.

We examined whether the inter-model spread in the projected California precipitation change is affected by the uncertainty in the projected changes in the E–NPSH latitude. This relationship is illustrated in Fig. 4b. As in climatology (Fig. 2b), the projected changes in California winter precipitation are linked to those in E–NPSH latitudes with a correlation coefficient of −0.53. Although this number is smaller than climatological relationship (i.e., −0.83), it is still statistically significant at the 99% confidence level. Thus, a considerable inter-model spread in the projected California precipitation changes can be attributed to the uncertainty in the projected changes in the E–NPSH latitude.
c. Uncertainty of future projections

What causes the inter-model spread in the projected changes in W–NPSH and E–NPSH latitudes? To better understand the uncertainty in the projected circulation changes, this section addresses the controlling factors of the NPSH change and its zonal asymmetry under global warming. From a zonal-mean perspective, the poleward boundary of the HC, adjoining the subtropical ridge, reaches a latitude where baroclinic instability occurs (Held 2000). Lu et al. (2008) evaluated the changes in baroclinicity under global warming and its impacts on the HC boundary by using the baroclinic criticality (Phillips 1954):

\[ C = \frac{f^2(u_{500} - u_{850})}{\beta g H(\theta_{500} - \theta_{850})/\Theta_0}, \]

where the zonal wind \((u)\) shear and potential temperature \((\theta)\) gradient are taken between 500 hPa and 850 hPa. \(H\) represents the air column thickness between 500 hPa and 850 hPa; \(H\) and others are simply set to be a constant. Lu et al. (2007, 2008) documented that the increased static stability on the equatorward side of the westerly jet reduces baroclinicity in the subtropics and leads an expansion of the HC. This scaling analysis has been mainly applied to the zonal-mean circulation.

In this study, the scaling analysis of Lu et al. (2007, 2008) is applied to regional circulation changes. Specifically, the relative importance of zonal wind shear \((u_{500} - u_{850})\) and static stability \((\theta_{500} - \theta_{850})\) changes in the NPSH latitude changes is evaluated. The two factors are computed around NPSH latitudes \((20^\circ-40^\circ)\) over the western and eastern North Pacific. Figure 5 exhibits their linear trends against the linear trends of the W–NPSH and E–NPSH latitudes. The projected changes in the W–NPSH latitudes are strongly correlated with the increase in the subtropical
static stability \((\theta_{500} - \theta_{850})\) with a correlation coefficient of 0.72 (Fig. 5a). Only a weak relationship is found between the W–NPSH and the wind shear \((u_{500} - u_{850})\) changes (Fig. 5c). This result indicates that the inter-model spread in W–NPSH latitude (Fig. 4a) is mainly due to uncertainty in the projected changes in static stability over the western North Pacific. However, this is not the case in the E–NPSH latitude changes. The inter-model spread of the projected change in E–NPSH latitude is strongly related with those in the vertical wind shear. The correlation coefficient between the two is \(-0.83\) (Fig. 5d). More importantly, E–NPSH–N and E–NPSH–S are well explained by different signs of vertical wind shear changes (i.e., E–NPSH–N for \(-0.01\) m/s/decade and E–NPSH–S for \(0.11\) m/s/decade). In contrast, only a weak relationship is found with static stability change \((r = 0.26;\) Fig. 5b). This result suggests that, while uncertainty in W–NPSH latitude changes is primarily associated with uncertain thermodynamic changes in a warmer climate, that in E–NPSH latitude changes is likely related with uncertain dynamical circulation changes.

As described earlier, projected changes in California winter precipitation are largely explained by those in E–NPSH latitudes. To better understand their relationship, the SLP and precipitation changes are separately examined for the models with E–NPSH–N and E–NPSH–S in Fig. 6. The two groups of models show substantial differences (see the Fig. 6c). For example, the E–NPSH–N models (Fig. 6a) exhibits strengthened SLP on the poleward side of the climatological NPSH, consistent with a poleward displacement of the E–NPSH. In these models, precipitation changes over California are generally weak and insignificant. The area-averaged precipitation over the red box in the figure is only 0.14 mm/day/century (see the red dot in Fig. 4b). In contrast, the E–NPSH–S models (Fig. 6b) show eastward and southward expansion of the AL in the 21st century, indicating an equatorward shift of storm tracks. Accordingly, the
California winter precipitation is predicted to increase substantially (0.86 mm/day/century; see the blue dot in Fig. 4b). This result confirms that projected changes in winter precipitation in California are to a large extent modulated by the atmospheric circulation (Chang et al. 2012; Neelin et al. 2013).

To further identify the atmospheric circulation changes associated with E–NPSH changes, the correlation map between the skin temperature (Ts) trends over the ocean and the E–NPSH latitudes trends is analyzed in Fig. 7a. Similar maps are also shown for SLP and zonal wind at 250 hPa (Figs. 7b and 7c). These figures illustrate the relationship in inter-model differences (not the MME mean changes). Figure 7a indicates that the inter-model spread in the projected change in the E–NPSH latitude is strongly associated with uncertainty in the Ts change over the western North Pacific instead of the eastern North Pacific. Such Ts trends resemble the Pacific Decadal Oscillation (PDO)-like sea surface temperature changes in the mid-latitudes. This PDO-like Ts changs tends to decrease (increase) latitudinal temperature gradient north (south) of 30°–40°N, leading to a poleward shift in the westerly jet, as shown in Fig. 7c. They are also related with an enhanced anti-cyclonic circulation over the North Pacific centered at 160°W and 40°N (Fig. 7b), a pattern that is identical to the difference between the models with E–NPSH–N and E–NPSH–S (see Fig. 6c). The associated zonal wind changes are weak but qualitatively similar to Fig. 7c, indicating a reduced vertical wind shear in consistent with Fig. 5d. A quasi-barotropic zonal wind change is common in the atmosphere–ocean coupling for internal variability (Nakamura et al. 1997; Yang et al. 2002; Barton and Ellis 2009; Sung et al. 2014). This result suggests that mid-latitude air–sea coupling can serve as a major source of uncertainty for the projected changes in E–NPSH latitudes and California winter precipitation.
To elucidate the leading patterns of model-to-model spread in atmospheric circulation linked to surface temperature trends over the North Pacific, we also performed a SVD analysis (Bretherton et al. 1992) to winter SLP and Ts trends under the RCP8.5 scenario (Fig. 8). Here, the leading SVD modes shown in Figs 8a and 8b denote the model-to-model spread deviated from the MME. The leading mode explains 62.3% of the total covariance. For the Ts and SLP variances, the leading SVD mode explains 43.4% and 31.6% of the total variance of each variable. A PDO-like warming emerges again over the western North Pacific (Fig. 8a). The magnitude of warming reached 1.2 K/century near the Kuroshio–Oyashio extension region. Although the center of the surface circulation change differs slightly from that shown in Fig. 6c and Fig. 7b, an enhanced anti-cyclonic circulation, corresponding to the PDO-like warming, is also evident (Fig. 8b). This result verifies that the PDO-like warming in the North Pacific and the related feedback processes are among the factors that give rise to inter-model spread in the projected changes of the E–NPSH latitude and California winter precipitation.

4. Summary and Discussion

This study assesses large-scale atmospheric circulation change over the North Pacific in a future climate and its uncertainty with implications for the winter precipitation in California. Since California is located at the transitional region between the Pacific storm track and a subtropical dry zone, its precipitation changes are likely controlled by multiple processes (e.g., Neelin et al. 2013). Accordingly, model biases and uncertainty in thermodynamical and dynamical processes can easily lead to considerable uncertainty and inter-model spread in the projected changes in Californian winter precipitation. Two indices, W–NPSH and E–NPSH, are
defined as proxies for the latitude of the local HC edge to quantify the North Pacific circulation and its response to global warming climate change. The location of the E–NPSH is highly correlated to California winter precipitation in both the mean model bias and future projections under the RCP8.5 scenario.

The CMIP5 models show strong agreement on the poleward displacement of the W–NPSH latitudes under the RCP8.5 scenario. Most models predict a decrease in high-latitude SLP but an increase in mid-latitude SLP over the W–NPSH region. This poleward shift of the W–NPSH latitude, or local expansion of the HC, is primarily explained by an increase of static stability in the subtropical North Pacific. Its inter-model spread is also well explained by uncertainty in the projected changes in static stability.

The E–NPSH latitudes do not show systematic changes. Only about half of the models (56%) predict a poleward shift of E–NPSH latitude, while others predict an equatorward shift. The net result is almost no change in E–NPSH latitude in MME with a large model-to-model spread. This uncertainty in the projected changes of the E–NPSH latitudes, which leads to the uncertain change in California winter precipitation, is largely explained by uncertain changes in vertical wind shear on the equatorward side of the Pacific jet. In contrast to the static stability change, the signs of the projected changes of vertical wind shear are not robust among the models. This model-to-model spread in the E–NPSH latitude change is dynamically related to the basin-wide quasi-barotropic atmospheric interaction to the PDO-like surface warming. This finding suggests that a certain amount of inter-model spread in the projected changes of the E–NPSH latitude and the California winter precipitation is rooted in atmosphere–ocean coupled mid-latitude dynamics over the western North Pacific. As such, to better understand and predict
precipitation changes over the western North America in the future climate, uncertainty of atmosphere–ocean coupling in the western Pacific needs to be taken into account.

One might wonder whether the spread in the projected shift in E–NPSH can be related to the mean bias of the circulation index, as has been found for the Southern Hemisphere in previous studies (Kidston and Gerber 2010, Son et al. 2010). To this end, we examine in Figure 9 the relationship between the projected shift of E–NPSH latitudes and the model bias (i.e., model climatology minus observation) in the latitudes of E–NPSH and W–NPSH. It is found that an equatorward bias in W–NPSH latitudes is linked to a greater poleward shift of E–NPSH with a correlation coefficient of –0.49 (Fig.9a). Somewhat surprisingly, this correlation is higher than the one against the bias in E–NPSH latitude (Fig. 9b; $r = –0.38$). This result again suggests that faithful representation of the air-sea coupling over the western North Pacific, which results in the correct position of the W–NPSH, is needed to reduce the uncertainty of the changes in California winter precipitation. From Figure 9, if the model mean bias has any guiding value, models with equatorward (poleward) biased W–NPSH latitudes may project a drier (wetter) future for California winter precipitation than how reality will unfold.

This study clearly shows that the magnitude of future projections of California winter precipitation depends on the specific model within the MME (see also Neelin et al. 2013). For example, the magnitude of the projected California precipitation changes under the RCP8.5 scenario can differ by a factor of six depending on the sign and the magnitude of the E–NPSH latitude trends. This implies that planning for adaptation and mitigation of the projected hydrometeorological climate changes at a regional scale, which is often based on MME results, should be considered
with a great deal of caution. Considering the uncertainty in the projected circulation change may be a good starting point for such research.

Acknowledgements

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References


Table 1. Models used in this study. The model acronyms were identified by the Fifth Assessment Report of the Intergovernmental Panel on Climate Change.

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Figure 1. Contour lines indicate the multi-model ensemble (MME) climatology (1959–1999) for December–January–February (DJF) (a) sea level pressure and (b) precipitation from the CMIP5 historical simulation. Shading notes the difference between the MME and observations. Here, HadSLP2 (1959–1999) and GPCP (1979–1999) are used for the observations. Thick solid and dashed purple lines in (a) indicate the locations of North Pacific Subtropical High (NPSH) from MME and observation, respectively. The red box in (b) includes the target area for California precipitation as reported by Neelin et al. (2013).
Figure 2. (a) Scatterplots of the climatological locations of the Eastern NPSH (E–NPSH) and the Western NPSH (W–NPSH). Each number denotes the model number as listed in Table 1. The correlation of the E–NPSH latitudes against the W–NPSH latitudes is indicated in the figure. The filled square and red circle indicate the MME and observation (HadSLP2), respectively. (b) Same as (a) but for E–NPSH latitudes and California precipitation (averaged over the red box area in Figure 1).
Figure 3. Future projections of DJF (a) sea level pressure and (b) precipitation changes over the period of 2007–2099 under the RCP8.5 scenario. The regions in which more than 66% of models (≥24) or 90% of models (≥32) have the same sign as the MME are represented by shading or hatches, respectively. Contours show the model spread (standard deviation) around the MME. The purple line in (a) and the red box in (b) are the same of those in Figure 1.
Figure 4. Same as Figure 2 but for the projected future change in the E–NPSH and W–NPSH latitudes and California precipitation over the period of 2007–2099. The black square denotes the MME from all 36 models. Red and blue circles indicate the MME of the selected models showing positive and negative trends of E–NPSH latitudes, respectively.
Figure 5. (a) Relationship between the projected change in the W–NPSH latitude and in static stability. (b) Same as (a) but for the E–NPSH. (c), (d) Same as (a), (b) but for the zonal wind shear on the $Y$-axis. The static stability and zonal wind shear are averaged over 20°–40°N and 140°E–120°W. Red and blue circles in (b) and (d) indicate the MME of the selected models showing positive and negative trends of E–NPSH latitudes, respectively.
Figure 6. Projected future changes of DJF sea level pressure (contours) and precipitation (shading) for the models of (a) the northward shift of the E-NPSH (E-NPSH-N) and (b) the southward shift (E-NPSH-S). (c) Difference between (a) and (b).
Figure 7. (a) Correlation map between the linear trends of skin temperature (Ts) and E–NPSH latitude. (b), (c) Same as (a) but for sea level pressure and zonal wind at 250 hPa, respectively.
Figure 8. Regression map against the leading principal component of the singular value decomposition (SVD) mode for (a) skin temperature and (b) sea level pressure. (c) Normalized principal components of the leading SVD mode. The X-axis in (c) indicates the model number as listed in Table 1.
Figure 9. Scatterplots showing the relationship between the projected future change in the E–NPSH latitudes and climatological locations of (a) W–NPSH and (b) E–NPSH. Each number denotes the model number as listed in Table 1. The filled square indicates the MME.