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The impact of sea surface temperature biases on North American precipitation in a high-resolution climate model

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# The impact of sea surface temperature biases on North American precipitation in a high-resolution climate model

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## ABSTRACT

Positive precipitation biases over western North America have remained a pervasive problem in the current generation of coupled global climate models. These biases are substantially reduced, however, in a version of the Geophysical Fluid Dynamics Laboratory Forecast-oriented Low Ocean Resolution (FLOR) coupled climate model with systematic sea surface temperature (SST) biases artificially corrected through flux adjustment. This study examines how the SST biases in the Atlantic and Pacific Oceans contribute to the North American precipitation biases. Experiments with the FLOR model in which SST biases are removed in the Atlantic and Pacific are carried out to determine the contribution of SST errors in each basin to precipitation statistics over North America. Tropical and North Pacific SST biases have a strong impact on northern North American precipitation, while tropical Atlantic SST biases have a dominant impact on precipitation biases in southern North America, including the western United States. Most notably, negative SST biases in the tropical Atlantic in boreal winter induce an anomalously strong Aleutian low and a southward bias in the North Pacific storm track. In boreal summer, the negative SST biases induce a strengthened North Atlantic Subtropical High and Great Plains low-level jet. Each of these impacts contributes to positive annual mean precipitation biases over western North America. Both North Pacific and North Atlantic SST biases induce SST biases in remote basins through dynamical pathways, so a complete attribution of the effects of SST biases on precipitation must account for both the local and remote impacts.

## 25 **1. Introduction**

26 Prediction of regional precipitation changes, from intraseasonal and seasonal climate  
27 forecasts to projections under global warming, remains a challenge owing to the complexity of  
28 physical processes cutting across a wide range of time and spatial scales. Consequently, state-of-  
29 the-art global climate models (GCMs) encounter persistent errors in simulating the temporal and  
30 spatial variations of precipitation (Dai 2006; Phillips and Gleckler 2006; Liu et al. 2014; Mehran  
31 et al. 2014). Pervasive and well-known biases include an unrealistic double Intertropical  
32 Convergence Zone (Mechoso et al. 1995; Lin 2007), errors in the precipitation diurnal cycle  
33 (Trenberth et al. 2003; Dai and Trenberth 2004), and the excessive production of light precipitation  
34 (Dai 2006; Sun et al. 2006; Wilcox and Donner 2007; Stephens et al. 2010). Regional  
35 climatological precipitation biases also are common. In this study, we focus on precipitation biases  
36 over North America, with emphasis on the tendency for the simulation of excessive precipitation  
37 in western North America (Phillips and Gleckler 2006; Sheffield et al. 2013; Liu et al. 2014;  
38 Mehran et al. 2014; Pascale et al. 2015; Mejia et al. 2018). Approximately 75% of all models  
39 participating in the Coupled Model Intercomparison Project phases 3 and 5 (CMIP3 and CMIP5)  
40 exhibit positive precipitation biases over the western United States (Mejia et al. 2018). This bias  
41 pattern incorporates an excessive amplitude of the annual cycle in the Pacific Northwest and the  
42 failure to capture the transition from a U.S. West Coast precipitation maximum to a Southwest  
43 minimum (Phillips and Gleckler 2006). The errors in southwestern North American precipitation  
44 relate, in part, to errors in the simulation of the North American monsoon system (NAMS), which  
45 features a peak in precipitation from July through September. GCMs typically simulate excessive  
46 NAMS precipitation amounts and season length, with both an early onset and late retreat (Geil et  
47 al. 2013; Sheffield et al. 2013).

48 Numerous sources likely share responsibility for the regional precipitation biases over  
49 North America, including coarse representations of topography and errors in subgrid-scale model  
50 parameterizations, like those of cloud microphysics and atmospheric convection. Common and  
51 persistent patterns of sea surface temperature (SST) biases also may play an important role by  
52 modifying the large-scale circulation and moisture transports. These common SST bias patterns  
53 include an excessively cold and westward extended Pacific cold tongue (Mechoso et al. 1995; Li  
54 and Xie 2014), warm SST biases in eastern tropical and subtropical oceans (Large and  
55 Danabasoglu 2006; Richter 2015; Zuidema et al. 2016), and cold SST biases in the North Atlantic  
56 and extratropical North Pacific (Wang et al. 2014; Zhang and Zhao 2014). Multiple reasons for  
57 these common SST biases have been suggested, including errors in alongshore winds and resulting  
58 ocean upwelling, misrepresented stratocumulus cloud decks and shortwave radiation fluxes, errors  
59 in ocean eddy mixing and vertical ocean temperature gradients (Xu et al. 2014; Richter 2015;  
60 Zuidema et al. 2016), and insufficient heat transport by the Atlantic meridional overturning  
61 circulation (AMOC) (Wang et al. 2014; Zhang and Zhao 2014). Some SST biases may improve  
62 with increasing oceanic and atmospheric resolution, but many of these biases still persist even as  
63 resolution is increased to eddy-permitting and eddy-resolving scales (Delworth et al. 2012;  
64 Kirtman et al., 2012; Griffies et al. 2015; Wittenberg et al. 2018; Vecchi et al. 2019; Adcroft et al.  
65 2019; Held et al. 2019). Attribution of regional SST biases is complicated by the potentially strong  
66 inter-basin links, as regional SST biases can induce biases in remote basins through atmospheric  
67 and oceanic pathways (Xu et al. 2014; Wang et al. 2014; Zhang et al. 2014; Zhang and Zhao 2014;  
68 Zuidema et al. 2016).

69 Although it is widely acknowledged that such SST biases can have important impacts on  
70 the simulation of atmospheric circulation and precipitation, few studies have provided a

71 comprehensive analysis of how common SST bias patterns affect the biases in other climatological  
72 features, including precipitation simulation. Several recent studies have demonstrated that Atlantic  
73 and Pacific SST biases can have far-reaching impacts on temperature, precipitation, and  
74 atmospheric circulation (Large and Danabasoglu 2006; Zhang et al. 2014; Zhang and Zhou 2014;  
75 Xu et al. 2014; Zuidema et al. 2016), although the analysis of these SST bias effects was limited.  
76 Keeley et al. (2012) performed a more targeted analysis of the effect of common North Atlantic  
77 SST biases on North Atlantic and European climate, concluding that the extratropical North  
78 Atlantic SST bias is a major cause of atmospheric circulation biases in the region. Zhang and Zhao  
79 (2014) also demonstrated that North Atlantic SST biases may induce large-scale circulation  
80 anomalies that project onto the northern annular mode, which then induce upstream climate  
81 anomalies, including SST biases in the North Pacific.

82         The changes in atmospheric circulation and moisture induced by SST biases also have the  
83 potential to affect the simulation of precipitation over North America. Recently, Mejia et al. (2018)  
84 performed a regional climate model study to demonstrate that typical SST biases offshore  
85 California and the Baja California Peninsula can explain a substantial fraction of the precipitation  
86 biases in the western United States. In the present study, we take a larger-scale perspective and  
87 investigate the impacts of these biases on North American climatological seasonal precipitation  
88 through the analysis of simulations from a high-resolution GCM, focusing on the impacts of both  
89 the Atlantic and Pacific SST biases and the interactions between the two basins. Approximate  
90 removal of the SST biases over the globe and in selected Atlantic and Pacific regions results in  
91 marked improvements in the simulation of North American precipitation, especially with respect  
92 to the strong zonal contrast between the western and eastern U.S. Emerging themes in this study  
93 include a dominant influence of Atlantic SST biases on the simulation of precipitation over the

94 U.S. and, as discussed briefly above, strong inter-basin links, whereby SST biases in the Pacific  
95 Ocean induce SST and atmospheric biases in the Atlantic, and vice versa.

96

## 97 **2. Data and Methodology**

98

### 99 *a. Observational data*

100

101 We analyze observational data primarily for the purpose of evaluating model biases,  
102 assessed for the 1951-2010 period. We estimate the observed climatological precipitation with the  
103 University of Delaware (UD) product (Willmott and Matsuura 2001), a gridded dataset at 0.5°  
104 resolution derived from station precipitation data. We assess the sensitivity of our analysis to  
105 observation precipitation dataset by performing the same calculations with Global Precipitation  
106 Climatology Centre (Schneider et al. 2014) and the Precipitation Construction over Land (Chen et  
107 al. 2002) datasets. All conclusions are insensitive to precipitation dataset, and so all results with  
108 these latter two datasets are relegated to the Supplemental Material. The climatological SST is  
109 derived from the monthly Hadley Centre Sea Ice and Sea Surface Temperature (HadISST) dataset  
110 (Rayner et al. 2003). For the storm track analyses, we use daily 500 hPa geopotential height and  
111 monthly mean 200 hPa zonal wind from the National Centers for Environmental Prediction-  
112 National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al. 1996) for the  
113 1976-2005 period. The 1976-2005 period is selected for comparison with the climate model  
114 control simulation with 1990 levels of radiative forcing, in contrast with the SST and precipitation  
115 bias analysis that measures against a simulation with historical levels of radiative forcing.

116



117 *b. FLOR Model and Experiments*

118

119         The GCM simulations analyzed in this study are generated by the Geophysical Fluid  
120 Dynamics Laboratory (GFDL) Forecast-oriented Low Ocean Resolution (FLOR) model (Vecchi  
121 et al. 2014; Wittenberg et al. 2018), a version of the Coupled Model version 2.5 (CM2.5; Delworth  
122 et al. 2012) that retains high resolution in the atmosphere and land components (approximately  
123 50km x 50km horizontal resolution) but has lower resolution in the ocean and sea ice components  
124 (horizontal grid spacing of 1°, telescoping to 0.33° meridional spacing near the equator).  
125 Quantities are exchanged between components conservatively, by first averaging from the  
126 transmitting component's grid onto an "exchange grid" (which is the refined "overlay" of the two  
127 participating components' grids), and then onto the receiving component's grid (Balaji et al. 2006).  
128 The high atmospheric and land resolution has yielded benefits in problems ranging from  
129 subseasonal (e.g., Xiang et al. 2014, 2019; Jiang et al. 2018) to seasonal prediction (Vecchi et al.  
130 2014; Jia et al. 2015; Yang et al. 2015; Murakami et al. 2016; Kapnick et al. 2018) and to  
131 anthropogenic climate change (Jia et al. 2016; van der Wiel et al. 2016; Pascale et al. 2017, 2018;  
132 Yang et al. 2018; Vecchi et al. 2019), although high atmospheric resolution is not a panacea (e.g.,  
133 Kapnick et al. 2018). The benefit to computational efficiency from the lower ocean and sea ice  
134 resolution allows us to carry out an extensive array of experiments.

135         We compare the climatological precipitation characteristics in two versions of FLOR, the  
136 standard free-running version and a version for which flux adjustments are applied to bring the  
137 model's climatological SST in alignment with observations (FLOR-FA). Specifically, the flux  
138 adjustment entails modifications to the model's momentum, enthalpy, and freshwater fluxes from  
139 the atmosphere to the ocean in order to remove most of the difference between the model and

140 observational estimates of climatological SST and surface wind stress for the 1979-2012 period.  
141 Additional details on the flux adjustment procedure are found in Vecchi et al. (2014). Figure 1  
142 illustrates the annual climatological precipitation over North America in FLOR, FLOR-FA, and  
143 observations, whereas Figure 2 illustrates the annual climatological SST biases in FLOR and  
144 FLOR-FA (similar SST bias patterns are found for individual seasons). All climatology and bias  
145 calculations are based on a simulation with historical estimates of radiative forcing for the 1951-  
146 2010 period. Consistent with the common biases discussed in the introduction (cf., Fig. 6 of  
147 Pascale et al. 2015), FLOR (Fig. 1b) fails to capture the amplitude of the observed (Fig. 1a) zonal  
148 gradient in climatological precipitation and simulates excessive precipitation over western North  
149 America. The climatological SST in FLOR also exhibits many of the biases discussed in the  
150 introduction (cf., Fig. 1 of Richter 2015): strong negative biases in the extratropical Pacific and  
151 Atlantic Oceans, an excessively cold Pacific cold tongue, and positive SST biases in eastern  
152 tropical and subtropical regions near continents (Fig. 2a). FLOR-FA, in contrast, performs better  
153 in simulating the sharp east-west precipitation gradient and reduces the western North American  
154 precipitation bias (Fig. 1c) (this improvement is quantified in Section 3). This distinction in  
155 climatological precipitation between FLOR and FLOR-FA holds for both cold and warm seasons  
156 and in all observational datasets analyzed (Fig. S1). As discussed above, FLOR-FA – by  
157 construction - also greatly reduces the SST biases (Fig. 2b), although the SST biases are not  
158 eliminated, particularly in extratropical regions where the biases are strongest.

159 In addition to the historical radiative forcing simulations, we also conduct a set of  
160 simulations with fixed radiative forcing to probe the physical processes that connect regional SST  
161 biases to global precipitation biases, as outlined in Table 1. First, we analyze years 101-200 from  
162 200-yr control simulations (to avoid any issues with model spin-up) with radiative forcing held

163 fixed at 1990 values (CTL and FA for the standard and flux-adjusted simulations, respectively).  
164 The climatological differences in precipitation, SST, and atmospheric circulation between CTL  
165 and FA are very similar to the differences in the historical forcing simulations. In order to  
166 determine the roles of individual basin SST biases in the simulation of North American  
167 precipitation, we next analyze a set of 100-yr SST nudging simulations with FLOR. In these  
168 simulations, we nudge the SSTs over individual basins to the total time varying values in FA (FA  
169 climatology plus FA anomalies) with a five-day restoring timescale. This procedure nearly  
170 eliminates the SST differences with FA over individual basins while allowing free  
171 ocean/atmosphere coupling in regions where SSTs are not restored. By allowing full coupling  
172 outside the restoring regions, we can capture the influence of SST biases in one region on the SST  
173 biases in remote regions, as discussed in the introduction. Because FA has much smaller SST  
174 biases than CTL (Fig. 2), the SST restoring experiments essentially isolate the influence of SST  
175 biases in individual basins on the simulated climate.

176 We focus on distinguishing the influence of SST biases in the North Pacific and North  
177 Atlantic Oceans in four distinct regions (Fig. 3). In the simulation designated as TPNP, we restore  
178 total SSTs in the tropical and extratropical North Pacific basin (15°S - 60°N, 120°E to South and  
179 North American West Coast, TPNP domain hereafter) to FA values. Similarly, in the simulation  
180 designated as TANA, we restore SSTs in the tropical and extratropical North Atlantic basin (15°S  
181 - 60°N, South and North American East Coast to African and European West Coast, TANA  
182 domain hereafter) to FA values. Beyond the edges of these domains away from the coastlines, we  
183 apply a 10° buffer within which the restoring is linearly reduced to zero. To distinguish the role  
184 of tropical versus extratropical SST biases, we conduct two additional experiments in which the

185 restoring is only applied to the tropics (15°S - 15°N) in the Pacific and Atlantic Oceans; we  
186 designate these experiments as TP and TA, respectively.

187 We conduct two additional experiments with a reduced length of 50 years to investigate  
188 the roles of local and non-locally induced SST biases. Climatological precipitation and  
189 atmospheric circulation differences between experiments exhibit only small differences when  
190 comparing 50-yr and 100-yr averages (not shown), and so we conclude that 50-yr simulations are  
191 sufficient for the purposes of this study. Because SST biases in one basin can impact the biases in  
192 remote regions, we wish to distinguish the influence of the local versus the remotely forced SST  
193 biases. In the experiment designated as TPNP<sub>iso</sub> (where “iso” stands for “isolated”), we restore  
194 SSTs in the TPNP domain to the FA values, just as in TPNP, but we also restore the TANA domain  
195 SSTs to CTL values. Therefore, the climatological SST differences between TPNP<sub>iso</sub> and CTL are  
196 confined to the tropical and extratropical North Pacific domain, and climatological SSTs are nearly  
197 identical between TPNP<sub>iso</sub> and CTL in all other ocean basins. Similarly, in the experiment  
198 designated as TANA<sub>iso</sub> we restore TANA domain SSTs to those of FA while also restoring the  
199 TPNP SSTs to CTL values. The TPNP<sub>iso</sub> and TANA<sub>iso</sub> experiments allow us to decompose the  
200 total effect of basin SST biases (CTL minus experiment) into locally and remotely forced  
201 components:

$$202 \quad \text{CTL} - \text{TPNP} = (\text{CTL} - \text{TPNP}_{\text{iso}}) + (\text{TPNP}_{\text{iso}} - \text{TPNP}) \quad (1)$$

$$203 \quad \text{CTL} - \text{TANA} = (\text{CTL} - \text{TANA}_{\text{iso}}) + (\text{TANA}_{\text{iso}} - \text{TANA}). \quad (2)$$

204 The left-hand side represents the total effect and the two terms on the righthand side represent the  
205 locally and remotely forced SST effects, respectively.

206

207 *c. Diagnostic Analyses*

208

209 To diagnose the impacts of FLOR’s SST biases on its atmospheric circulation and North  
210 American precipitation, we calculate composite differences between the simulations described  
211 above. To keep the analysis as simple as possible while also illustrating seasonality in the  
212 response, we subdivide the calendar into two six-month seasons, a cold (October – March) and  
213 warm season (April – September). Except for the historical bias calculations, differences express  
214 how CTL compares with the experiment of interest and are calculated as CTL minus the  
215 experiment. To calculate differences in the storm tracks, we identify the storm tracks by the  
216 variance of the high-pass filtered 500 hPa geopotential height (z500) fields, where we use a  
217 Butterworth filter to retain z500 variance with periods less than eight days.

218 To provide further insight into how the circulation and moisture changes induced by SST  
219 biases impact climatological precipitation, we analyze the moisture budget differences between  
220 the experiments. The climatological precipitation budget (e.g. Seager and Henderson 2013) can  
221 be approximated by

$$222 \quad \bar{P} = -\frac{1}{\rho_w g} \nabla \cdot \int_0^{p_s} (\bar{\mathbf{u}}\bar{q} + \overline{\mathbf{u}'q'}) dp + \bar{E}, \quad (3)$$

223 where  $P$  is the precipitation,  $\rho_w$  is the density of water,  $g$  is the gravitational acceleration,  $p_s$  is the  
224 surface pressure,  $\mathbf{u}$  is the horizontal wind vector,  $q$  is the specific humidity, and  $E$  is the surface  
225 evaporation. Double overbars represent climatological seasonal means, and primes represent  
226 deviations from the monthly means, which are at daily resolution in this study. Products of  
227 monthly anomalies are neglected, as the monthly transient eddy convergence term is small over  
228 the domain of interest (not shown). The two terms within the integral represent the effects of  
229 moisture convergence from the climatological flow and the submonthly transient eddy moisture  
230 flux convergence, respectively.

231 As discussed in Seager and Henderson (2013), the moisture budget calculations are quite  
232 sensitive to the horizontal, vertical, and temporal resolution of the archived data, which typically  
233 are stored in a standard grid that is distinct from the model's native grid. In the FLOR experiments,  
234 the monthly data are saved at 17 standard vertical levels, but the relevant daily data are only  
235 available at three vertical levels (surface, 850 hPa, and 500 hPa). The poor vertical resolution of  
236 the higher-frequency data means that the transient eddy moisture flux convergence calculations  
237 are not reliable. Nevertheless, we evaluated whether the estimates from (3) are accurate enough  
238 to provide some insight about the differences in seasonal mean precipitation between the  
239 experiments. Figure 4 shows the seasonal CTL minus FA precipitation differences and the  
240 corresponding differences estimated by (3). The fields in Fig. 4 are smoothed through 20 iterations  
241 of two-dimensional convolution with a 3 x 3 kernel, which especially reduces error in the  
242 decomposition by (3) over regions of strongly varying topography. The actual and derived  
243 precipitation climatology differences in Fig. 4 agree rather well over the Pacific, North America,  
244 and Atlantic regions, indicating that the resolution of the archived data is enough to capture general  
245 features in the precipitation budget differences. For the entire Northern Hemisphere, the pattern  
246 correlations between the actual precipitation climatology difference pattern and that derived from  
247 (1) are 0.89 in October-March and 0.92 in April-September, supporting the reliability of the  
248 moisture budget decomposition in capturing the overall spatial differences. The quantitative  
249 differences, however, are large enough that caution must be made to avoid overextending the  
250 interpretations.

251 We further subdivide the mean flow convergence component of the precipitation  
252 differences into dynamic and thermodynamic components. Specifically, we decompose the  
253 climatological precipitation differences between experiments as

254 
$$\delta\bar{P} = -\frac{1}{\rho_w g} \nabla \cdot \int_0^{p_s} ([\delta\bar{\mathbf{u}}]\bar{q} + \bar{\mathbf{u}}[\delta\bar{q}] + \delta\overline{\mathbf{u}'q'}) dp + \delta\bar{E}, \quad (4)$$

255 where the  $[\delta\bar{\mathbf{u}}][\delta\bar{q}]$  term has been neglected because it is much smaller than the other terms. The  
 256 first term on the right-hand side of (4) represents the impact of the change in climatological  
 257 circulation, holding the climatological specific humidity constant. We call this term the circulation  
 258 bias term. The second term on the right-hand side of (4), the humidity bias term, captures the  
 259 impact of the change in climatological specific humidity, holding the climatological mean flow  
 260 constant. These two terms indicate whether the removal of SST biases impacts precipitation more  
 261 strongly through changes in specific humidity that accompany SST changes (thermodynamics) or  
 262 through impacts of SSTs on the atmospheric circulation, which then impacts precipitation patterns  
 263 (dynamics).

264

### 265 **3. Results**

266 The seasonal North American precipitation biases in the historical FLOR and FLOR-FA  
 267 simulations, presented as a percentage relative to the observed climatology, are illustrated in Fig.  
 268 5. Consistent with Fig. 1, the reduction of SST biases in FLOR-FA reduces or eliminates the  
 269 precipitation biases over portions of western North America. In the extended winter, flux  
 270 adjustment reduces precipitation biases over a large portion of North America, although the wet  
 271 bias persists in FLOR-FA (Fig. 5c). Observational errors in the precipitation climatology,  
 272 however, are clear in the cold season, as a bias discontinuity is apparent at the United States-  
 273 Canada border due to differences in precipitation collection technology leading to improved  
 274 precision in Canada (Adam and Lettenmaier 2003). In the warm season, the bias reduction is even  
 275 stronger, especially over regions most strongly affected by the NAMS and over the Rockies. This

276 finding is consistent with recent work that found superior performance of FLOR-FA in simulating  
277 the NAMS (Pascale et al. 2017, 2018). We note, however, that FLOR-FA does exacerbate the dry  
278 bias over the south-central U.S. in both seasons. Overall, flux adjustment in FLOR reduces the  
279 precipitation climatology root-mean-square error over the U.S. region (25-50°N, 60-130°W) by  
280 18.3% in October-March and by 43.4% in April-September. We find nearly identical results when  
281 using the other observed precipitation datasets (Figs. S2 and S3).

282

### 283 *a. TPNP and TANA simulation results*

284

285         Next, we analyze the TPNP and TANA simulation results to attribute in a general sense  
286 the importance of Pacific and Atlantic SST biases for the FLOR/FLOR-FA climatological  
287 precipitation differences. We begin by analyzing climatological differences in precipitation and  
288 atmospheric circulation between the 100-yr CTL and each of the TPNP and TANA simulations  
289 (designated as  $\delta P_{\text{TPNP}}$  and  $\delta P_{\text{TANA}}$  for the TPNP and TANA precipitation differences, respectively).  
290 A comparison of these plots with the corresponding CTL minus FA plots reveals the degree to  
291 which SST biases in the individual basins can explain the differences in the total SST bias-related  
292 precipitation differences over North America.

293         In Fig. 6 we focus on differences in precipitation, sea level pressure (SLP), and 925 hPa  
294 wind. The 925 hPa wind corresponds closely with the Caribbean and Great Plains low-level jets,  
295 which have a strong impact on the warm season hydroclimate of the central United States (e.g.,  
296 Krishnamurthy et al. 2015) and are impacted by coupled model SST biases (e.g., Krishnamurthy  
297 et al. 2015, 2019). Consistent with the analysis presented earlier, the CTL simulation produces  
298 much wetter conditions over southern North America, especially over the southwestern region,



299 than FA in both the cold and warm seasons (Fig. 6a,b). Figure 6 also reveals that the wetter North  
300 America is accompanied by wetter conditions in the equatorial Atlantic and Pacific Oceans, a much  
301 deeper wintertime Aleutian low, a weaker summertime North Pacific High and continental low in  
302 the North American monsoon region, and a stronger western portion of the summertime North  
303 Atlantic Subtropical High (NASH). Fig. 6 presents the precipitation differences as fractional  
304 differences relative to the CTL simulation, but the largest absolute differences (shown in Fig. S4)  
305 occur in the deep tropics, a region where the differences in convective heating can induce large  
306 differences in the extratropical circulation. The cold season composite differences bear a close  
307 resemblance to the composites associated with strong El Niño episodes (e.g. Johnson and Kosaka  
308 2016), suggesting a role for tropically forced changes in the large-scale circulation and Pacific  
309 storm track, which we explore later. Overall, Figs. 6a and b are consistent with large SST-induced  
310 differences in atmospheric circulation that result in stronger imports of atmospheric moisture into  
311 southern North America in FLOR relative to FLOR-FA.

312 The remainder of Fig. 6 illustrates how much of the CTL/FA differences can be explained  
313 by tropical and extratropical North Pacific (Fig. 6c,d) or North Atlantic (Fig. 6e,f) differences.  
314 The overall impression is that both TPNP and TANA SST biases, primarily negative (Fig. 2),  
315 contribute to drier conditions in northern North America and wetter conditions in southern North  
316 America. Surprisingly, the TANA SST differences appear to have a dominant influence on the  
317 southern North America precipitation and even the North Pacific atmospheric circulation  
318 differences in both seasons. In the cold season, both the TANA and TPNP SST biases induce a  
319 strengthened Aleutian low, but the Aleutian low response to TANA SST biases is stronger (Fig.  
320 6e). Even more surprisingly, the TANA SST biases induce stronger positive fractional

321 precipitation biases in the equatorial Pacific Ocean than the direct response to Pacific SST biases,  
322 at least in the extended winter season.

323 We quantify the impacts of North Atlantic and North Pacific SST biases on the  
324 climatological CTL/FA precipitation differences in Fig. 7. Specifically, we calculate the  
325 percentage of  $\delta P_{FA}$  that can be attributed to  $\delta P_{TPNP}$  and  $\delta P_{TANA}$ . We mask out regions where the  
326 CTL/FA precipitation differences are less than 10% of the CTL climatology to focus on regions  
327 where the differences are large. The results of Fig. 7 confirm the visual impression of Fig. 6 in  
328 that both Atlantic and Pacific SST biases are important for the continental U.S. during the extended  
329 winter, that tropical and/or extratropical North Atlantic (North Pacific) SST biases dominate the  
330 impacts over southern (northern) North America, and that the Atlantic SST biases are particularly  
331 important over the continental U.S. during the extended summer (Fig. 7d). We note, however, that  
332 we should not expect the total impact of North Atlantic and North Pacific SST biases to be a linear  
333 superposition of the TPNP and TANA simulation results because the Pacific and Atlantic SST  
334 biases affect the SST biases in remote ocean basins, as discussed in Section 3c.

335 The strength of the impact of North Atlantic SST biases on North Pacific precipitation and  
336 atmospheric circulation, though surprising, is consistent with recent studies that have examined  
337 multi-decadal variability and trends of Atlantic SSTs (Kucharski et al. 2011; McGregor et al. 2014;  
338 Li et al. 2016; Ruprich-Robert et al. 2017). In particular, the climate modeling studies of  
339 McGregor et al. (2014) and Li et al. (2016) demonstrate that the tropical Atlantic warming trend  
340 over the past few decades has the potential to induce negative SST and precipitation trends over  
341 the tropical Pacific via modifications of the Walker circulation and coupled ocean/atmosphere  
342 feedbacks. These changes in the tropical oceans also impact the circulation and precipitation over  
343 the North Pacific and North America (McGregor et al. 2014). Recent studies of Atlantic

344 multidecadal variability reveal consistent results. Anomalously warm conditions in the tropical  
345 Atlantic result in negative precipitation anomalies over the tropical Pacific and an anomalously  
346 weak Aleutian low, which impacts the downstream North American climate (Sutton and Hodson  
347 2007; Kushnir et al. 2010; Ruprich-Robert et al. 2017). The negative tropical Atlantic SST biases  
348 in FLOR result in the expected response (opposite to that seen from warming) seen in Fig. 6; that  
349 is, a stronger Aleutian low.

350         The results in Fig. 6 generally are consistent with the picture presented above and more  
351 generally with the studies of Wang et al. (2007, 2008), which examined the influence of the  
352 Atlantic warm pool on Western Hemisphere climate, albeit with a focus only on the summer  
353 season. In both the FA and TANA response maps, large negative precipitation differences are  
354 present over the tropical Atlantic and northern South America, over and near the regions where  
355 tropical Atlantic SST differences are strongly negative. The reduction of atmospheric convection  
356 in the Atlantic warm pool results in a “Gill response” (Gill 1980) that manifests as positive SLP  
357 differences near and just northwest of the precipitation anomalies (Sutton and Hodson 2007; Wang  
358 et al. 2007; Kushnir et al. 2010). The response, however, is not confined to the tropical Atlantic,  
359 as the atmospheric Rossby and Kelvin wave response spreads the anomalous cooling to the tropical  
360 Pacific, destabilizing the atmosphere and promoting enhanced convection remote from the Atlantic  
361 SST forcing (Sutton and Hodson 2007; Kushnir et al. 2010). Therefore, tropical Atlantic cooling  
362 promotes a dipole of anomalous convection, with suppressed convection over the tropical western  
363 Atlantic and enhanced convection in the central and eastern tropical Pacific.

364         In boreal winter the enhanced tropical Pacific convection resulting from the Atlantic  
365 cooling has the potential to force a Pacific/North American-like (PNA-like) circulation pattern that  
366 features an enhanced Aleutian low (Sutton and Hodson 2007; Ruprich-Robert et al. 2017), as

367 shown in Fig. 6e. The tropical Pacific SST differences also can induce tropical precipitation  
368 differences that induce a strengthened Aleutian low (Fig. 6c), but the response is not as strong,  
369 possibly because the tropical Pacific SST differences are not as large as the tropical Atlantic SST  
370 differences (Fig. 2) and possibly because the tropical Atlantic atmospheric convection anomalies  
371 are well positioned to induce remote coupled ocean-atmosphere feedbacks in the tropical Pacific  
372 basin (Li et al. 2016; Ruprich-Robert et al. 2017). We examine the remote SST impacts of the  
373 Atlantic SST biases in Section 3c.

374         In the summer months, the Atlantic SST differences potentially can exert stronger direct  
375 impacts on North American precipitation (Wang et al. 2007, 2008; Kushnir et al. 2010). Figure 6f  
376 indicates positive SLP differences between CTL and TANA over the western tropical Atlantic and  
377 over southern North America, which indicate a strengthened western portion of the NASH and a  
378 weakened North American monsoon low. This pattern is consistent with the climate model  
379 experiments of Wang et al. (2007, 2008) that demonstrated the role of the Atlantic warm pool in  
380 modifying the strength of the summertime NASH and the Great Plains and Caribbean low-level  
381 jets, which then impacts the northward moisture transport and precipitation in the central U.S.  
382 (Wang et al. 2008).

383         Overall, the results presented in this section suggest both Pacific and Atlantic SST biases  
384 prominently drive North American precipitation biases. We also suggest plausible mechanisms  
385 that are consistent with previous studies that focused primarily on the impacts of Pacific and  
386 Atlantic SST variability. We next examine the roles of tropical and extratropical SST biases, the  
387 inter-basin links among the SST biases, and precipitation budget diagnostics to determine if the  
388 arguments presented above hold up to further scrutiny.

389

390 *b. TP and TA simulation results*

391

392         The arguments regarding the prominent role of Atlantic SST biases on the North Pacific  
393 circulation and North American precipitation suggest that tropical rather than extratropical Atlantic  
394 SST biases play the more crucial role. The reason is that tropical Atlantic SST biases can directly  
395 influence moisture advection into the US, and tropical SST biases can more easily induce upstream  
396 circulation impacts due to the larger length scales of the atmospheric response in the tropics  
397 relative to the extratropics. To investigate this hypothesis, we show the CTL/TP and CTL/TA  
398 seasonal composite differences in circulation and precipitation in Fig. 8. Consistent with  
399 expectations, the tropical Atlantic SST biases appear to dominate the Atlantic SST effects on  
400 circulation and North American precipitation. In both seasons, the TA results are similar to those  
401 of TANA (compare Fig. 8c,d with Fig. 6e,f). The tropical Atlantic precipitation and hemispheric  
402 circulation response in the TA results are slightly stronger than that of the TANA experiment,  
403 indicating that the extratropical Atlantic SST biases act to damp the full Atlantic SST response  
404 slightly, particularly in the extended summer. The reason for this damping requires further  
405 investigation, but it appears that colder North Atlantic sea surface in FLOR can induce a stronger  
406 NASH that increases moisture convergence in the Caribbean Sea, partially offsetting the reduced  
407 moisture and atmospheric instability owing to the colder tropical Atlantic sea surface. The  
408 offsetting influence of the extratropical North Atlantic SSTs is consistent with the GCM  
409 experiments of Okumura et al. (2009), who investigated the mechanisms by which a large  
410 freshwater forcing of the North Atlantic can impact North Pacific climate.

411         Examination of the TP results suggests that for the Pacific, both tropical and extratropical  
412 SST biases contribute to North American precipitation biases in the boreal cold season (compare

413 Fig. 6c with Fig. 8a) but that tropical SST biases play little role in the boreal warm season. The  
414 enhanced subtropical convection in CTL relative to TP in the cold season (Fig. 8a) where CTL  
415 SST biases are positive (Fig. 2) may contribute to the slightly stronger Aleutian low through a  
416 poleward propagating Rossby wave response. The tropical Pacific SST biases, however, are small  
417 relative to the extratropical biases (Fig. 2). The strongly negative SST biases in the central North  
418 Pacific in CTL increase the baroclinicity, which also enhances the North Pacific storm track into  
419 southern North America. We examine storm track changes more closely in Section 3d. Overall,  
420 we find that in the dynamically active boreal cold season, both tropical and extratropical North  
421 Pacific SST biases have a substantial impact on FLOR's simulation of North American  
422 precipitation. This contrasts the interannual variability of North American precipitation, in which  
423 tropical Pacific SSTs are believed to play a much stronger role than extratropical Pacific SST  
424 variability (e.g. Kushnir et al. 2002). A key difference is that FLOR's pattern of mean SST biases  
425 (with strong biases in the extratropics and in the tropical Atlantic), looks quite different from  
426 ENSO SST anomalies, which typically have their strongest signature in the central and eastern  
427 equatorial Pacific.

428

429 *c. TPNP<sub>iso</sub> and TANA<sub>iso</sub> simulation results*

430

431 As discussed above, the SST biases in the North Atlantic and the North Pacific can induce  
432 nonlocal SST biases through atmospheric and oceanic pathways. Therefore, the TANA and TPNP  
433 simulations do not necessarily isolate the impacts of the SST biases in the basins for which the  
434 SSTs have been restored. We illustrate the non-local SST impacts in Fig. 9, which shows the  
435 differences in annual mean climatological SSTs between the CTL and each of the TPNP and

436 TANA simulations. By construction, the SST differences over the North Pacific (North Atlantic)  
437 domains defined in Fig. 3 for the TPNP (TANA) simulation are nearly equal to the CTL/FA  
438 differences. The SST differences in all other ocean basins are remotely forced.

439 The SST differences between the CTL and TANA simulation (Fig. 9) reveal that the  
440 negative tropical and North Atlantic SST biases induce strongly negative SST biases in the  
441 extratropical North Pacific, strongest near 40°N. The North Pacific response to Atlantic SST  
442 forcing is consistent with past North Atlantic “water hosing” experiments (Zhang and Delworth  
443 2005; Okumura et al. 2009) in which the North Atlantic is cooled through a large freshwater input  
444 as well as recent studies on Atlantic multidecadal variability (Zhang and Delworth 2007; Ruprich-  
445 Robert et al. 2017; Johnson et al. 2018) and climate model SST biases (Wang et al. 2014; Zhang  
446 and Zhao 2014). Wang et al. (2014) demonstrate that the strength of AMOC may be a key factor  
447 in the link between North Pacific and North Atlantic SST biases in the models participating in  
448 CMIP5.

449 Similarly, the tropical and North Pacific SST biases remotely force Atlantic SST biases  
450 (Fig. 9a), although the overall impact is not as strong as that of the Atlantic on the Pacific. The  
451 negative SST differences over much of the North Atlantic indicate that the removal of the North  
452 Pacific SST biases in the TPNP simulation also reduces the negative SST biases in portions of the  
453 North Atlantic. In the sub-Arctic North Atlantic, the SST differences are positive, possibly  
454 reflecting a shift of the Gulf Stream or changes in the AMOC and oceanic deep convection.  
455 Although the amplitude of remote Atlantic SST changes (Fig. 9a) is considerably less than that of  
456 the remote Pacific SST changes (Fig. 9b), the North Pacific SST biases induce substantial  
457 decreases in the North Atlantic meridional SST gradient (Fig. 9a) and baroclinicity in vicinity of

458 the North Atlantic storm track, which, as shown in the following section, result in notable increases  
459 in evaporation (Figs. 12 and 13) and a reduced storm track intensity (Fig. 15).

460 To distinguish the roles of local versus remotely forced SST biases, we examine the results  
461 of the TANA<sub>iso</sub> and TPNP<sub>iso</sub> experiments following the decompositions given in (1) and (2). The  
462 decomposition of the Atlantic SST bias effect given by (2) is illustrated in Fig. 10. The top panels  
463 show notably stronger precipitation differences over North America than the bottom panels, which  
464 indicate a dominance of the locally forced Atlantic SST bias effects. In October – March, the  
465 remotely forced effects (Fig. 10c) are consistent with those of the TPNP experiments, indicating  
466 that the North Pacific cooling induced by the negative tropical and North Atlantic SST biases  
467 induces drying over northern North America and wetting over southern North America. The local  
468 and nonlocal Atlantic SST bias effects oppose each other in northern North America but reinforce  
469 each other over southern North America (compare Figs. 10a with 10c). In April – September, the  
470 local and nonlocal effects oppose each other over most of North America, but the local Atlantic  
471 SST effects dominate even more than in the boreal cold season.

472 The decomposition of the Pacific SST bias effect reveals a more complicated picture (Fig.  
473 11), particularly in the boreal cold season. In October – March over southwestern North America,  
474 the local and remote TPNP SST effects reinforce each other, indicating that the negative SST  
475 biases in both ocean basins result in increased precipitation. In other parts of North America, the  
476 two effects tend to oppose each other. Most conspicuously, the negative TPNP SST bias pattern  
477 directly results in positive SLP differences over the North Pacific (Fig. 11a), but the negative SST  
478 differences induced in the tropical and North Atlantic (Fig. 9a) force negative SLP differences  
479 over the North Pacific (Fig. 11c) that overcompensate the positive SLP differences. This



480 cancellation between the direct and indirect effect over the North Pacific explains why the impact  
481 of the North Pacific SST biases on the North Pacific circulation is relatively modest (Fig. 6c).

482         The positive SLP response to the negative SST differences over the North Pacific (Fig.  
483 11a) resembles the direct, linear baroclinic response to extratropical SSTs noted in previous studies  
484 (Peng et al. 1997; Peng and Whitaker 1999; Kushnir et al. 2002). Specifically, the North Pacific  
485 high diminishes in amplitude with height (not shown), consistent with the expected direct response  
486 to shallow cooling. However, the total response to extratropical cooling is strongly mediated by  
487 synoptic eddies, which is highly sensitive to the background flow (Peng et al. 1997). The total  
488 eddy-mediated response to North Pacific cooling in Fig. 11a, with a surface high over the North  
489 Pacific and an upper-level trough extending from the eastern Pacific over much of North America  
490 (not shown) resembles the response to North Pacific SST anomalies with February background  
491 conditions studied in Peng et al. (1997) and Peng and Whitaker (1999). However, those previous  
492 studies showed that the response pattern is quite distinct with January background conditions,  
493 demonstrating that the synoptic eddy-mediated response to North Pacific extratropical SST  
494 anomalies is highly sensitive to the background climatology. Therefore, we urge caution to avoid  
495 over generalizing these results.

496         The remote Pacific SST bias effect over North America is substantial in October – March  
497 (Figs. 11c) and generally consistent with the local Atlantic SST bias effect (Fig. 10a). In the  
498 context of all other simulation results and previous studies noted above, this finding reinforces that  
499 negative tropical Atlantic SST biases in the boreal cold season are effective in inducing an  
500 anomalously deep Aleutian low and anomalously wet conditions over much of southern North  
501 America. Moreover, a substantial portion of the negative tropical Atlantic SST biases are remotely

502 forced by the North Pacific SST biases. In the boreal warm season, however, the remotely forced  
503 effect of North Pacific SST biases over North America is small (Fig. 11d).

504

505 *d. Precipitation budget diagnostics*

506

507 To gain additional insight into the mechanisms that connect SST biases to North American  
508 precipitation biases, we examine the contributions to the simulations' climatological precipitation  
509 differences determined from equation (4). Specifically, we focus on the circulation bias, humidity  
510 bias, and evaporation terms, as these terms generally are the largest contributors to the  
511 climatological precipitation differences. As we note above we cannot derive accurate estimates of  
512 the contributions by transient eddy convergence owing to insufficient diagnostic output. To shed  
513 light on the possible role of differences in transient eddies, we examine differences in the  
514 climatological storm tracks.

515 We first focus on the climatological differences in October – March (Fig. 12). Overall, the  
516 circulation bias and evaporation terms make the greatest contributions to the precipitation  
517 differences over the North American continent for each pair of experiments. In general, the  
518 humidity bias term tends to oppose the changes from the circulation bias term, but the effects of  
519 the changing circulation dominate over the thermodynamic effects. Both the TPNP (Fig. 12b) and  
520 TANA (Fig. 12c) experiments capture the CTL minus FA circulation bias pattern, with the TANA  
521 differences generally showing stronger magnitudes. These findings indicate that the climatological  
522 circulation changes induced by the North Pacific and especially tropical Atlantic SST biases  
523 dominate the SST-induced differences in wintertime climatological precipitation over North  
524 America.

525           However, there are a number of regions where the circulation bias effects are not the  
526 dominant factor during the extended winter season. The CTL minus FA circulation bias pattern  
527 (Fig. 12a) features negative differences over Baja California, parts of western North America, and  
528 a portion of the southwestern U.S., which contrast the positive precipitation biases in FLOR over  
529 this region. These negative circulation-induced differences are overwhelmed by the effects of  
530 evaporation (Fig. 12g) and, to a lesser extent, the humidity bias term (Fig. 12d). These opposing  
531 influences are captured well in the TPNP experiment (Fig. 12b,e,h). The residual term (bottom  
532 row of Fig. 12) generally features positive differences in southern North America and negative  
533 differences over the northwest coast of North America. This residual likely reflects, in large part,  
534 the omission of the change in transient eddy fluxes from the change in storm tracks, as discussed  
535 below.

536           In contrast with the extended winter, all three terms make sizeable contributions to the  
537 climatological precipitation differences from April – September (Fig. 13). Once again, the  
538 humidity bias term (second row) tends to oppose the effects of the corresponding circulation term  
539 (top row). The combination of the three terms results in a tendency for positive precipitation biases  
540 over most of North America (Fig. 4d), but the dominant term varies regionally. Overall, the TANA  
541 experiment (right column) captures the total difference patterns (left column) rather well for all  
542 three terms, confirming the dominance of the Atlantic SST biases on the SST-related  
543 climatological precipitation biases over North America in the FLOR model.

544           Changes in the storm tracks also modify the transient eddy moisture flux convergence,  
545 thereby also contributing to the climatological precipitation differences. Figure 14 provides the  
546 climatological storm tracks in CTL, FA, and in reanalysis data for both the extended winter and  
547 summer. Compared with CTL (Fig. 14c,f), FA features more northerly displaced North Pacific

548 and North Atlantic storm tracks in both seasons (Fig. 14b,e), with a somewhat weaker storm track  
549 in the western North Pacific in boreal winter. The northward shift of the storm tracks in FA results  
550 in improved agreement with the position of the storm tracks derived from reanalysis data (Fig.  
551 14a,b), although the stronger storm track in CTL more closely matches reanalysis data in the  
552 western North Pacific region. Overall, Fig. 14 indicates that the SST biases in CTL result in a  
553 southward bias in the location of the dominant Northern Hemisphere storm tracks.

554 The differences in the storm tracks between CTL and FA (Fig. 15a,b) clearly show the  
555 southerly displacement over both basins and the stronger North Pacific storm track in CTL. The  
556 winter difference pattern resembles, in many respects, the storm track response to El Niño (e.g.,  
557 Johnson and Kosaka 2016) as well as to the negative phase of the Atlantic Multidecadal Oscillation  
558 (Zhang and Delworth 2007), which suggests that SST biases in both the Pacific and Atlantic may  
559 contribute to these differences. Consistently, both the TPNP (Fig. 15c,d) and TANA experiments  
560 (Fig. 15e,f) produce similar changes to the storm tracks, indicating that both the Atlantic and  
561 Pacific SST biases contribute to the stronger and southerly displaced storm tracks in CTL. The  
562 storm track differences shown in Fig. 15 closely mirror the differences in 200 hPa zonal wind. The  
563 southerly bias in the North Pacific storm track in CTL likely contributes to the wetter conditions  
564 in southwestern and drier conditions in northwestern North America relative to FA, a conclusion  
565 that is corroborated by the estimated transient eddy contributions to the precipitation budgets,  
566 although we choose not to show these results because the insufficient spatial diagnostic outputs  
567 limit the reliability of these estimates (Section 2c).

568

## 569 **4. Discussion and Conclusions**

570

571 In this study, we have examined the role of SST biases on the simulation of North American  
572 climatological precipitation in a global climate model with 50 km atmospheric horizontal  
573 resolution, the GFDL FLOR model. Like many climate models, FLOR simulates excessive  
574 precipitation over much of western North America, leading to a failure to simulate the strong east-  
575 west contrast in climatological precipitation in observations. A flux-adjusted version of FLOR  
576 that greatly reduces SST biases mitigates this deficiency in continental precipitation, indicating  
577 that SST biases are a contributor to these precipitation biases. Previous investigations have  
578 reached similar conclusions regarding the simulation of the NAMS (Liang et al. 2008; Pascale et  
579 al. 2017, 2018), the Great Plains and Caribbean Low-Level Jets (Krishnamurthy et al. 2019), Gulf  
580 of California moisture surges into southwestern North America (Pascale et al. 2016), and western  
581 United States climatological precipitation in general (Mejia et al. 2018), but the present study  
582 focuses on the pathways by which Atlantic and Pacific SST biases contribute to the simulation of  
583 excessive precipitation. The main findings of the study are summarized in schematic diagrams  
584 shown in Fig. 16. Because the SST biases in FLOR share many features common to many or even  
585 most current global climate models (e.g., Wang et al. 2014; Richter 2015), the results presented  
586 here likely apply broadly to a range of climate models.

587 From the analysis presented here, a few general themes emerge. First, relative to the  
588 Pacific, FLOR's Atlantic SST biases make a substantially greater contribution to the excessive  
589 precipitation in the western U.S. and Mexico for both seasons. One reason appears to be  
590 substantially stronger SST biases in the tropical Atlantic relative to the tropical Pacific in FLOR,  
591 given that tropical SST biases are most effective in exciting large-scale circulation responses  
592 owing to their effects on tropical convection and Rossby wave sources. Although the relative  
593 strength of the tropical Pacific SST biases may differ in other climate models, the strong tropical

594 Atlantic SST biases are pervasive in the current generation of global climate models (Li and Xie  
595 2014; Wang et al. 2014). Another factor is the effectiveness of tropical Atlantic SST biases to  
596 induce substantial circulation and moisture anomalies both locally, through changes in the NASH  
597 and associated low-level jets, and nonlocally in the Indian and Pacific Oceans, through  
598 modifications of the Walker circulation. The latter mechanism, which has been corroborated in  
599 several recent studies on Atlantic multidecadal variability, results in a strong link between negative  
600 SST biases in the tropical Atlantic and an anomalously deepened Aleutian low and an associated  
601 southerly shift of the storm tracks, which contribute substantially to the wet bias over western  
602 North America. Overall, these findings suggest that reductions of tropical Atlantic SST biases in  
603 coupled GCMs, which appear to be closely tied to biases in the Atlantic meridional overturning  
604 circulation (Wang et al. 2014), would have substantial benefits for the simulation of precipitation  
605 over the United States and Central America, especially in boreal summer.

606 Another emerging theme is the difficulty in isolating the effects of local SST biases owing  
607 to precipitation responses to remote SST effects, e.g., the response of precipitation to the SST  
608 changes induced in the North Pacific by Atlantic SST biases. Negative SST biases in the Atlantic  
609 induce negative SST biases in the extratropical Pacific, and negative SST biases in the North  
610 Pacific induce negative SST biases in both the tropical and extratropical North Atlantic. Both the  
611 local and remotely forced SST biases appear to have substantial influences on the atmospheric  
612 circulation and North American precipitation. Another apparently important factor for the reduced  
613 impact of the North Pacific SST biases relative to North Atlantic SST biases is the competing  
614 impacts of the local North Pacific and remotely forced SST bias impacts on the North Pacific  
615 atmospheric circulation. Specifically, the North Pacific surface high forced by local negative SST  
616 biases partially offsets the deepened Aleutian low response to the remotely forced negative tropical

617 Atlantic SST biases (Fig. 11a,c). However, these competing effects may be challenging to  
618 disentangle in studies with multi-model ensembles, as previous studies have demonstrated that the  
619 eddy-mediated response to extratropical SST forcing is sensitive to the details of the background  
620 flow.

621 As discussed in Section 1, various processes and modeling deficiencies contribute to the  
622 pervasive SST biases in the current generation of global climate models. As both  
623 parameterizations improve and model resolution increases, we expect that these SST biases  
624 accordingly shall reduce. The findings presented here provide insight into the expected changes  
625 in climatological precipitation over North America as these SST biases are reduced, regardless of  
626 whether the precipitation biases in other models are stronger or weaker than those of FLOR. This  
627 study also suggests that flux adjustment may remain a viable intermediate solution for problems  
628 for which climatological precipitation simulation is critical. For example, the improved simulation  
629 of the North American monsoon in FLOR-FA has enabled new insights into projected changes of  
630 this monsoon system under global warming (Pascale et al. 2017, 2018). These recent studies  
631 illuminate how the climate sensitivity of some facets of the climate system may be affected by  
632 climatological SST biases and how the removal of these biases through flux adjustment can  
633 improve confidence in projected changes. In addition, a set of seasonal hindcasts with FLOR-FA  
634 successfully captured the western U.S. precipitation pattern during the El Niño winter of 2015-16,  
635 a pattern that was generally poorly predicted and atypical of other strong El Niño events (Yang et  
636 al. 2018). This finding raises questions about how the reduction of SST biases may impact seasonal  
637 forecasts of western U.S. precipitation. Future work shall address how SST biases may impact  
638 other aspects of the variability, prediction skill, and projected changes of North American  
639 precipitation, including the tails of precipitation distribution.

640

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918

## 919 **List of Figures**

920 FIG. 1. Climatological (1951-2010) annual precipitation ( $\text{mm d}^{-1}$ ) in (a) observations, (b) FLOR,  
921 and (c) FLOR-FA.

922

923 FIG. 2. Climatological (1951-2010) annual SST biases (K) in (a) FLOR and (b) FLOR-FA.

924

925 FIG. 3. Regions for which SSTs are restored to FA values (dark red) in (a) TPNP, (b) TANA, (c)  
926 TP, and (d) TA experiments. The regions in which the color smoothly transitions from red to blue  
927 indicate the buffer regions where the restoring is relaxed to zero.

928

929 FIG. 4. CTL minus FA climatological precipitation differences ( $\text{mm d}^{-1}$ ) for (a,c) October – March  
930 and (b,d) April – September. The top panels are the actual differences and bottom panels are  
931 derived from the budget decomposition estimated from (1).

932

933 FIG. 5. Climatological precipitation biases (% relative to U. Delaware precipitation) over North  
934 America in (top) FLOR and (bottom) FLOR-FA for (a,c) October – March and (b,d) April –  
935 September. The values in the lower right corner of each panel are the RMSE ( $\text{mm d}^{-1}$ ) in the U.S.  
936 region (red box in a).

937

938 FIG. 6. FIG. 6. Impact of (left) October – March and (right) April – September (top) global,  
939 (middle) tropical and extratropical Pacific, and (bottom) tropical and extratropical Atlantic FLOR  
940 SST biases, including the effects of remotely induced SST biases, as expressed by the (a,b) CTL  
941 minus FA, (c,d) CTL minus TPNP, and (e,f) CTL minus TANA climatological differences in  
942 precipitation (color shading), SLP (contours), and 925 hPa wind (vectors) for (left) October –  
943 March and (right) April – September. Precipitation differences are expressed as percentage  
944 relative to CTL climatology. SLP is contoured at intervals of 1 hPa with red (blue) lines indicating  
945 positive (negative) differences, and the zero contour is omitted. The reference vector for 925 hPa  
946 wind is shown in the bottom right of panel d.

947

948 FIG. 7. Percent of  $\delta P_{\text{FLOR-FA}}$  that can be attributed to (a,b) TPNP and (c,d) TANA domain SST  
949 differences for (left) October – March and (right) April – September. Areas masked in grey  
950 represent regions where the FLOR-CTL minus FLOR-FA precipitation differences are less than  
951 10% of the FLOR-CTL climatological precipitation.

952

953 FIG. 8. Impact of (top) tropical Pacific and (bottom) tropical Atlantic FLOR SST biases, including  
954 the effects of remotely induced SST biases, as expressed by the (a,b) CTL minus TP and (c,d) CTL  
955 minus TA climatological differences in precipitation (color shading), SLP (contours), and 925 hPa

956 wind (vectors) for (left) October – March and (right) April – September. Precipitation differences  
957 are expressed as percentage relative to CTL climatology. SLP is contoured at intervals of 1 hPa  
958 with red (blue) lines indicating positive (negative) differences, and the zero contour is omitted.  
959 The reference vector for 925 hPa wind is shown in the bottom right of panel b.

960

961 FIG. 9. (a) CTL minus TPNA and (b) CTL minus TANA annual mean climatological SST  
962 differences (K).

963

964 FIG. 10. As in Fig. 8 except for the (a,b) locally forced tropical and extratropical North Atlantic  
965 SST effect (CTL minus  $TANA_{iso}$ ) and the (c,d) remotely forced tropical and extratropical North  
966 Atlantic SST effect ( $TANA_{iso}$  minus TANA).

967

968 FIG. 11. As in Fig. 8 except for the (a,b) locally forced tropical and extratropical North Pacific  
969 SST effect (CTL minus  $TPNP_{iso}$ ) and the (c,d) remotely forced tropical and extratropical North  
970 Pacific SST effect ( $TPNP_{iso}$  minus TPNP).

971

972 FIG. 12. Contributions to the October – March (left column) CTL minus FA, (middle column)  
973 CTL minus TPNP, and (right column) CTL minus TANA climatological precipitation  
974 differences ( $\text{mm d}^{-1}$ ) attributed to the following terms: (a-c) circulation bias, (d-f) humidity bias,  
975 (g-i) evaporation, and (j-l) the residual, calculated as the actual precipitation difference minus the  
976 sum of the three terms.

977

978 FIG. 13. As in Fig. 12 but for April – September.

979

980 FIG. 14. Climatological storm tracks, as measured by 8-day high-pass filtered 500 hPa geopotential  
981 height variance (shading,  $m^2$ ) and 200 hPa zonal wind (grey contours at an interval of  $10 \text{ ms}^{-1}$ ) for  
982 (left) NCEP/NCAR reanalysis data, (center) FA, and (right) CTL simulations in (a-c) October –  
983 March and (d-f) April – September.

984

985 FIG. 15. Differences in climatological storm tracks, as measured by 8-day high-pass filtered 500  
986 hPa geopotential height variance (shading,  $m^2$ ), and 200 hPa zonal wind (grey contours at an  
987 interval of  $2 \text{ ms}^{-1}$ , zero contour is omitted) for (a,b) CTL minus FA, (c,d) CTL minus TPNP, and  
988 (e,f) CTL minus TANA in (left) October – March and (right) April – September.

989

990 FIG. 16. Schematic showing the dominant impacts of Pacific and Atlantic SST biases on North  
991 American precipitation biases in the boreal cold and warm seasons. In the boreal cold season, (a)  
992 negative SST biases in the extratropical North Pacific promote a strengthened and southerly shifted  
993 storm track, which enhances precipitation in the southwestern United States and suppresses  
994 precipitation in northern Canada. (b) Tropical Atlantic cold SST biases induce circulation changes  
995 throughout the entire tropics resembling the classic Gill model, with a surface anticyclone in the  
996 vicinity of the cold bias and low-level convergence and enhanced precipitation in the equatorial  
997 Pacific. The enhanced tropical Pacific rainfall excites a deepened Aleutian low and enhanced  
998 moisture transport and precipitation in the southern United States. In the boreal warm season, the  
999 effects of (c) North Pacific SST biases are modest, but a weaker northern portion of the North  
1000 Pacific storm track promotes drier conditions in northern North America. (d) The cold Atlantic  
1001 SST biases have a much stronger impact, substantially strengthening the western lobe of the North

1002 Atlantic Subtropical High and weakening the thermal low over southern North America. These  
1003 changes enhance the Great Plains Low Level jet and moisture transport into southwestern North  
1004 America. Because the SST biases in each basin influence the SST biases in the other basin, the  
1005 total SST bias effects are not limited to the direct effects described here.

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1025 Table 1. List of FLOR experiments analyzed in this study.

Experiment name	Description	Duration (years)
CTL	FLOR with 1990 radiative forcings	100
FA	FLOR with 1990 radiative forcings and flux adjustments to correct SST biases	100
TPNP	CTL but with tropical and extratropical North Pacific SSTs restored to FA values	100
TANA	CTL but with tropical and extratropical North Atlantic SSTs restored to FA values	100
TP	CTL but with tropical Pacific SSTs restored to FA values	100
TA	CTL but with tropical Atlantic SSTs restored to FA values	100
TPNP <sub>iso</sub>	As in TPNP but with tropical and extratropical North Atlantic SSTs restored to CTL values	50
TANA <sub>iso</sub>	As in TANA but with tropical and extratropical North Pacific SSTs restored to CTL values	50

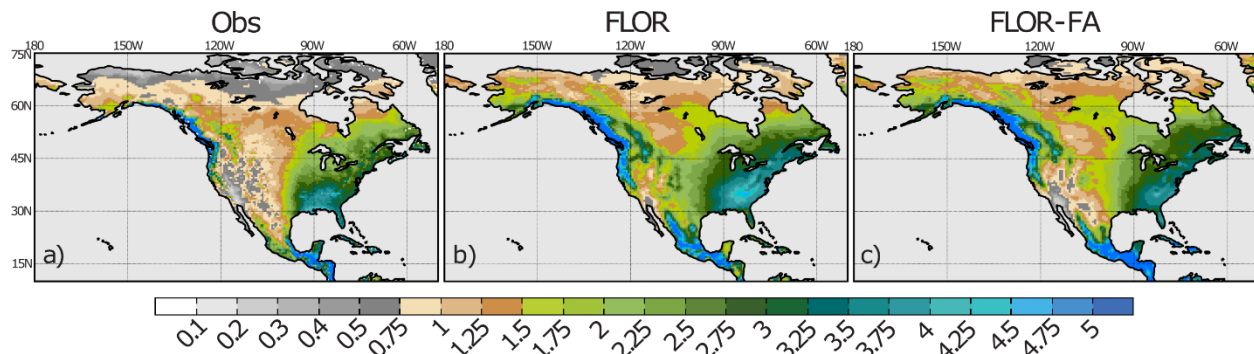
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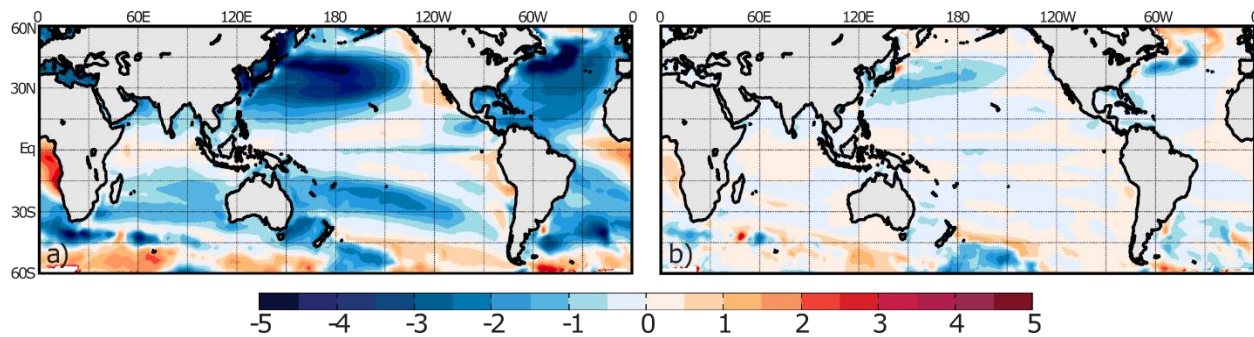
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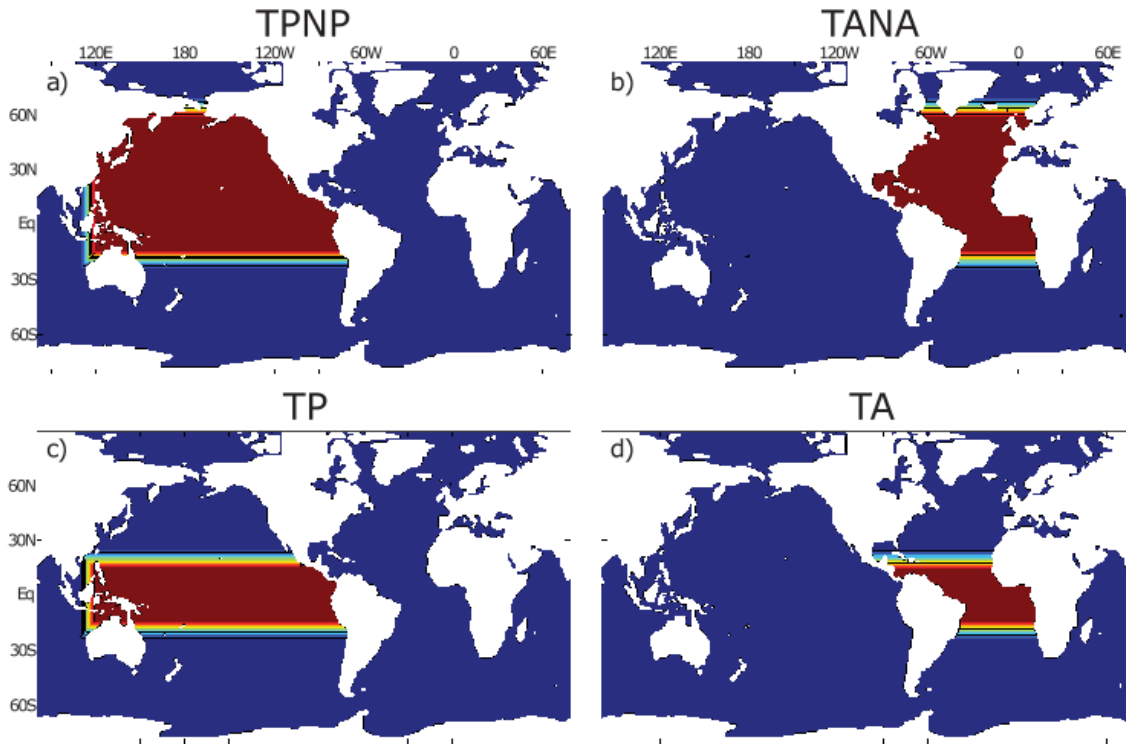
1031  
 1032 FIG. 1. Climatological (1951-2010) annual precipitation ( $\text{mm d}^{-1}$ ) in (a) observations, (b) FLOR,  
 1033 and (c) FLOR-FA.

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 1040 FIG. 2. Climatological (1951-2010) annual SST biases (K) in (a) FLOR and (b) FLOR-FA.

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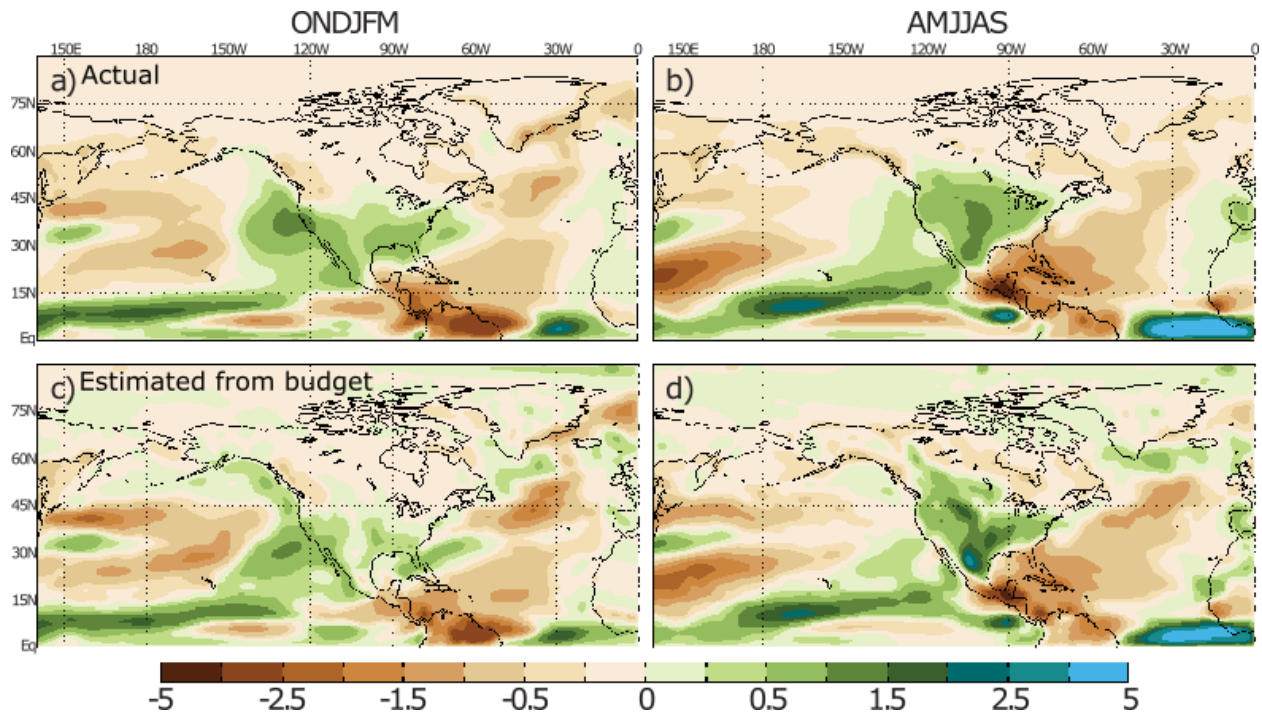
1043

1044 FIG. 3. Regions for which SSTs are restored to FA values (dark red) in (a) TPNP, (b) TANA, (c)  
 1045 TP, and (d) TA experiments. The regions in which the color smoothly transitions from red to blue  
 1046 indicate the buffer regions where the restoring is relaxed to zero.

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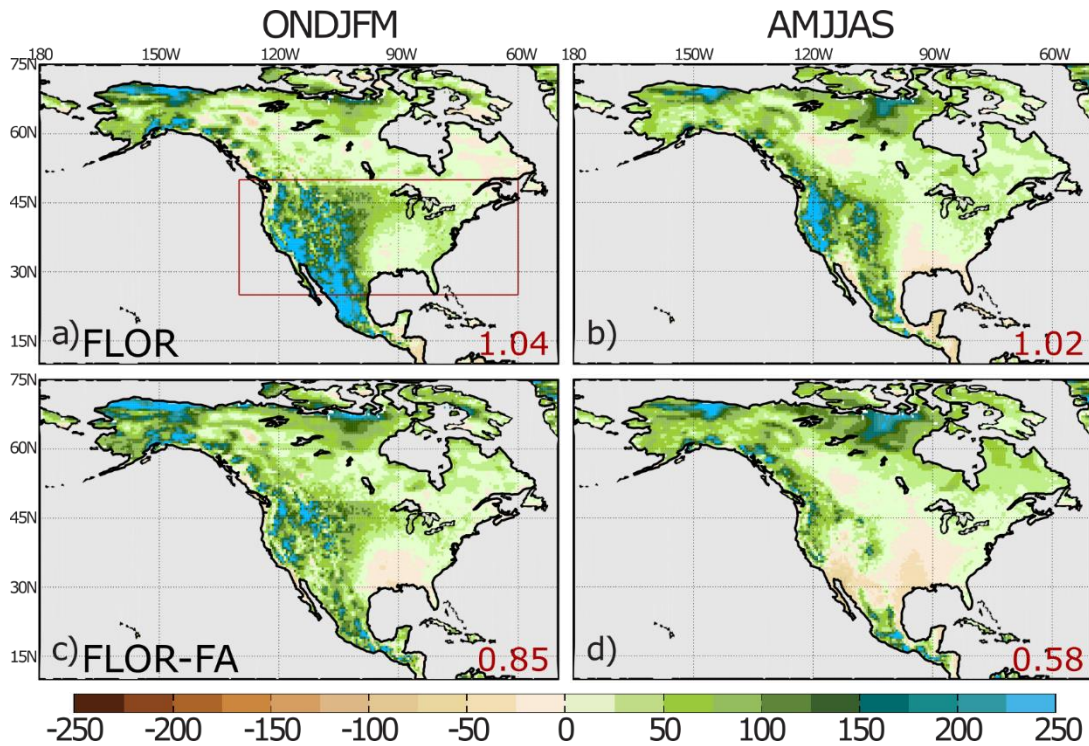
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 1053 and (b,d) April – September. The top panels are the actual differences and bottom panels are  
 1054 derived from the budget decomposition estimated from (1).

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1060 FIG. 5. Climatological precipitation biases (% relative to U. Delaware precipitation) over North  
 1061 America in (top) FLOR and (bottom) FLOR-FA for (a,c) October – March and (b,d) April –  
 1062 September. The values in the lower right corner of each panel are the RMSE ( $\text{mm d}^{-1}$ ) in the U.S.  
 1063 region (red box in a).

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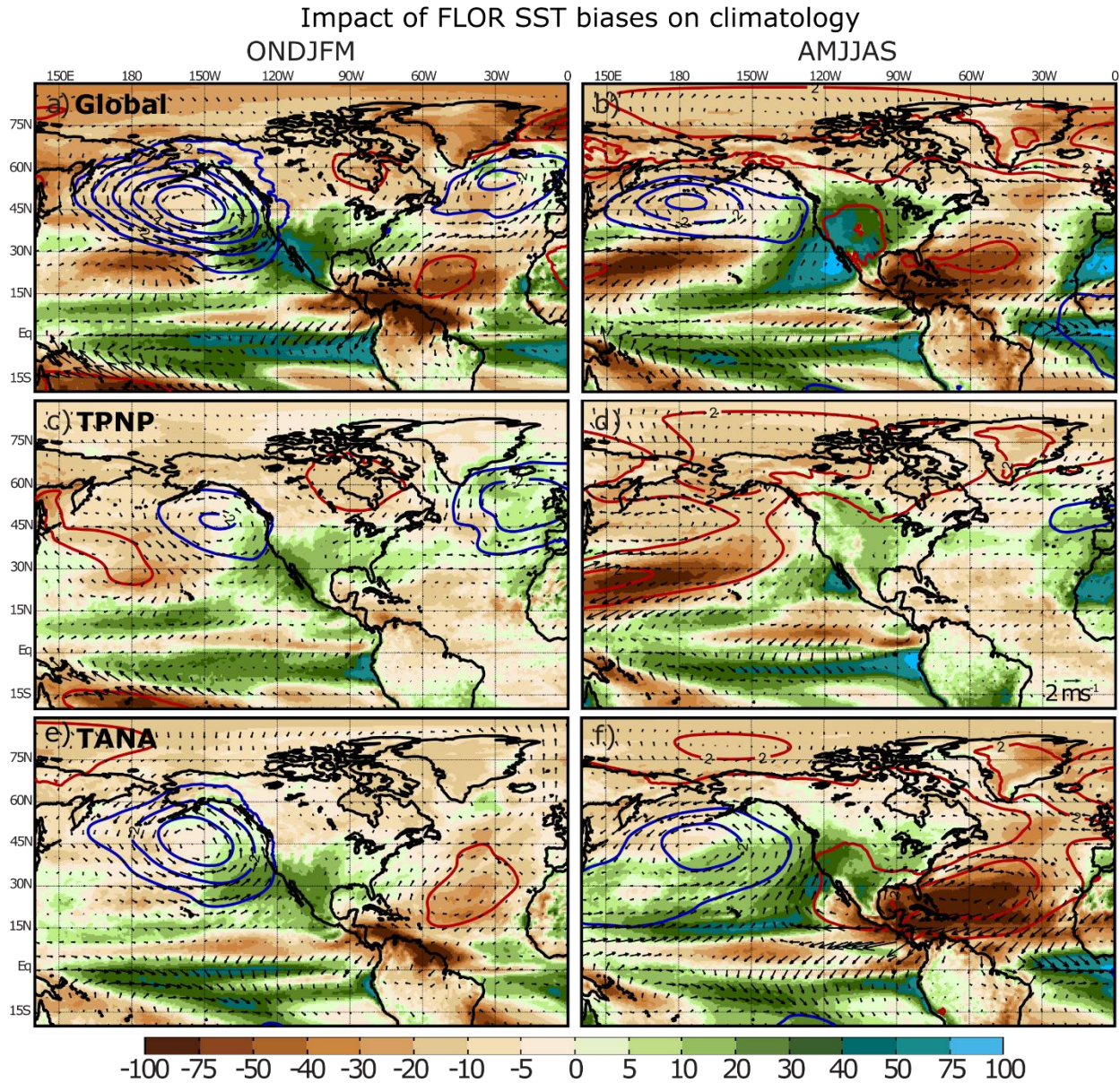
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1072 FIG. 6. Impact of (left) October – March and (right) April – September (top) global, (middle)

1073 tropical and extratropical Pacific, and (bottom) tropical and extratropical Atlantic FLOR SST

1074 biases, including the effects of remotely induced SST biases, as expressed by the (a,b) CTL minus

1075 FA, (c,d) CTL minus TPNP, and (e,f) CTL minus TANA climatological differences in

1076 precipitation (color shading), SLP (contours), and 925 hPa wind (vectors) for (left) October –

1077 March and (right) April – September. Precipitation differences are expressed as percentage

1078 relative to CTL climatology. SLP is contoured at intervals of 1 hPa with red (blue) lines indicating

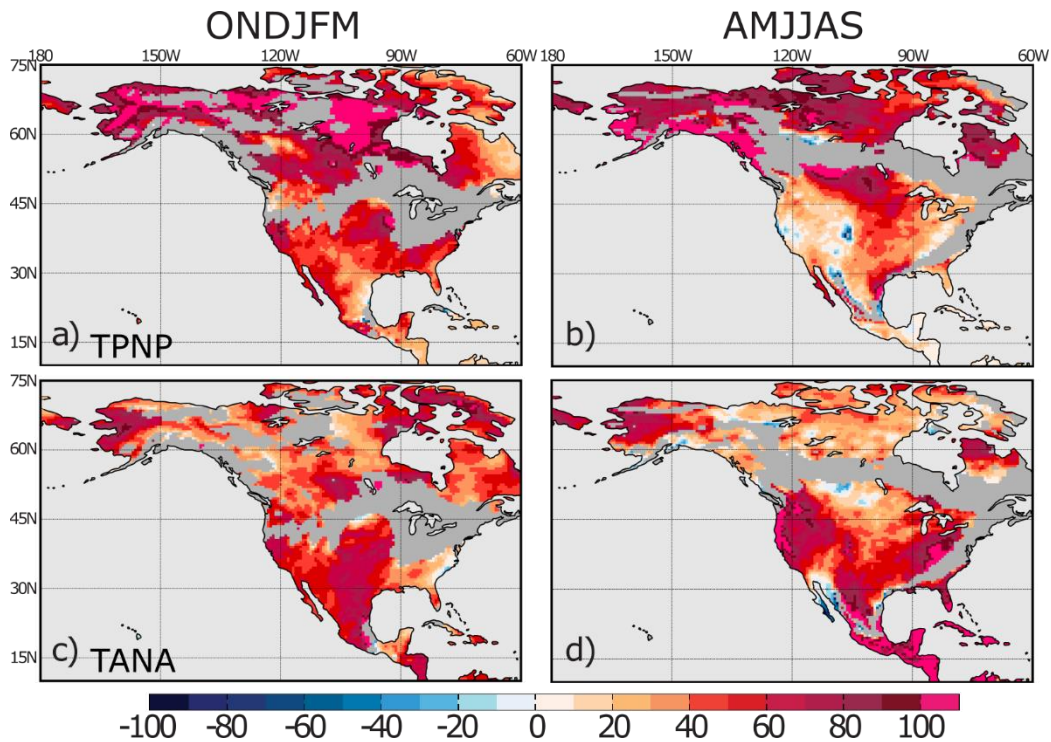
1079 positive (negative) differences, and the zero contour is omitted. The reference vector for 925 hPa  
1080 wind is shown in the bottom right of panel d.

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1087 FIG. 7. Percent of  $\delta P_{\text{FLOR-FA}}$  that can be attributed to (a,b) TPNP and (c,d) TANA domain SST  
1088 differences for (left) October – March and (right) April – September. Areas masked in grey  
1089 represent regions where the FLOR-CTL minus FLOR-FA precipitation differences are less than  
1090 10% of the FLOR-CTL climatological precipitation.

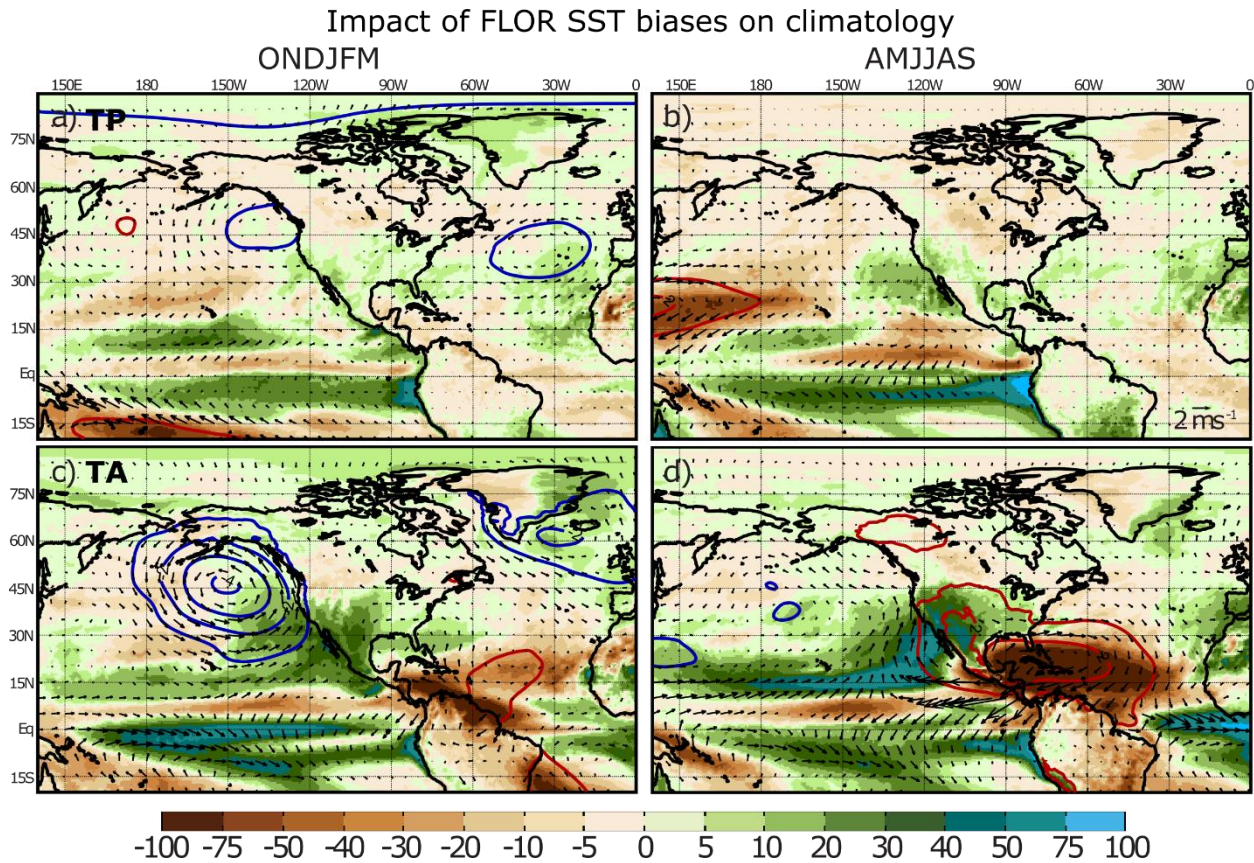
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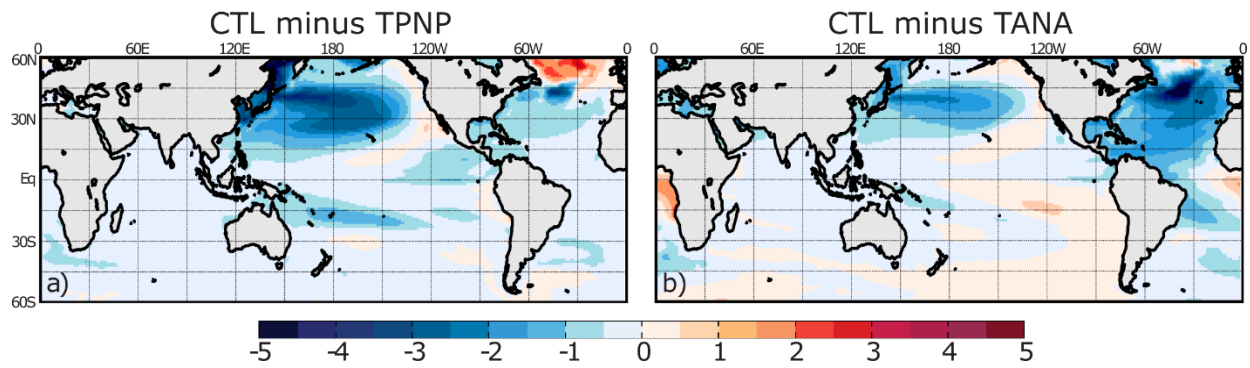
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FIG. 8. Impact of (top) tropical Pacific and (bottom) tropical Atlantic FLOR SST biases, including the effects of remotely induced SST biases, as expressed by the (a,b) CTL minus TP and (c,d) CTL minus TA climatological differences in precipitation (color shading), SLP (contours), and 925 hPa wind (vectors) for (left) October – March and (right) April – September. Precipitation differences are expressed as percentage relative to CTL climatology. SLP is contoured at intervals of 1 hPa with red (blue) lines indicating positive (negative) differences, and the zero contour is omitted. The reference vector for 925 hPa wind is shown in the bottom right of panel b.



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1107 FIG. 9. (a) CTL minus TPNP and (b) CTL minus TANA annual mean climatological SST  
1108 differences (K).

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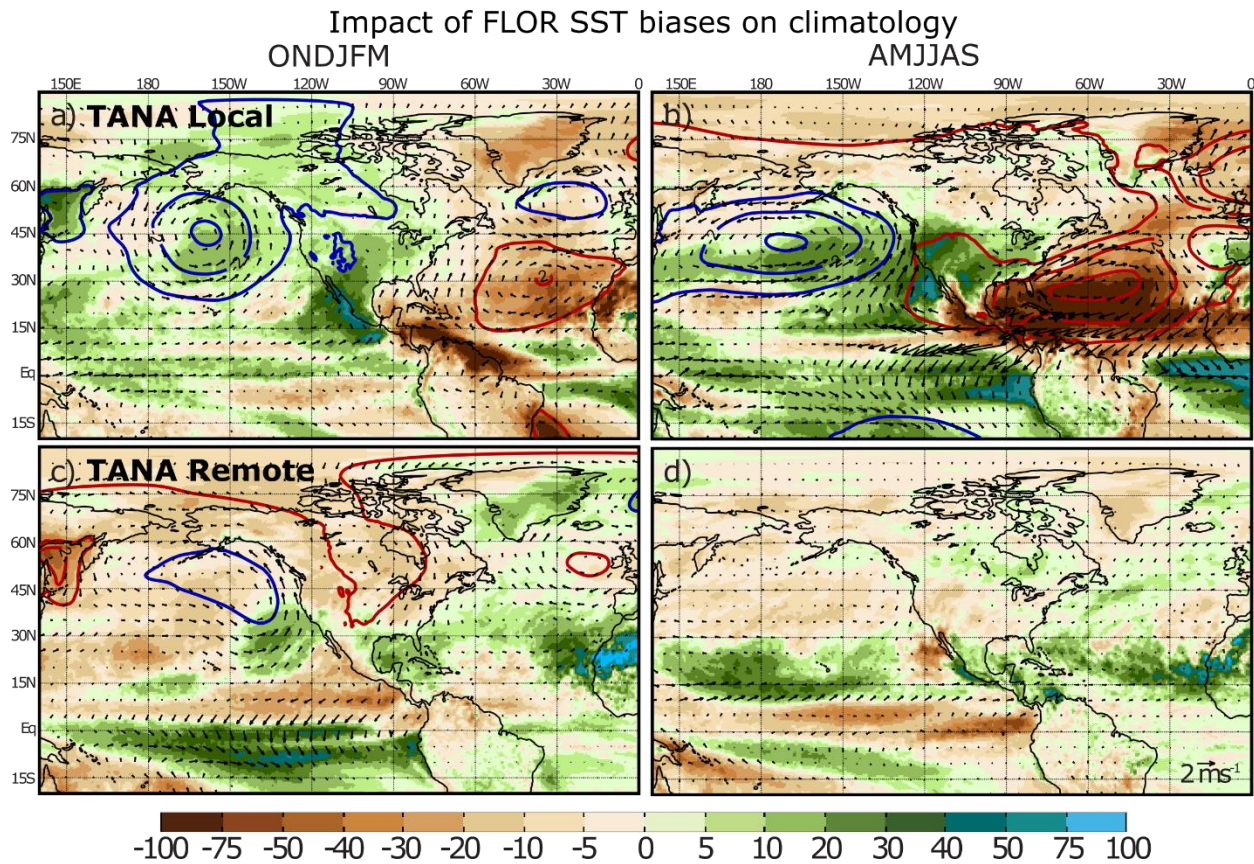
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1117 FIG. 10. As in Fig. 8 except for the (a,b) locally forced tropical and extratropical North Atlantic

1118 SST effect (CTL minus  $TANA_{iso}$ ) and the (c,d) remotely forced tropical and extratropical North

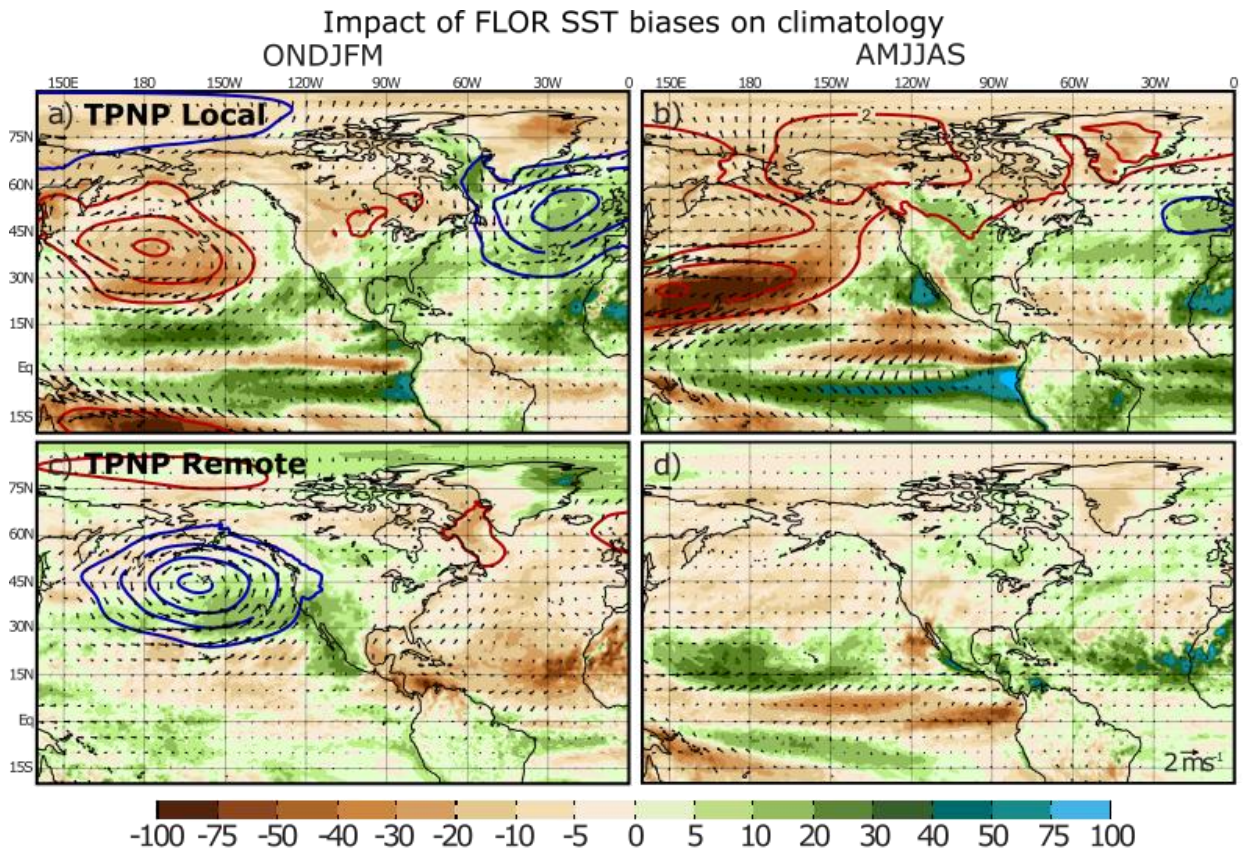
1119 Atlantic SST effect ( $TANA_{iso}$  minus TANA).

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1125 FIG. 11. As in Fig. 8 except for the (a,b) locally forced tropical and extratropical North Pacific

1126 SST effect (CTL minus  $\text{TPNP}_{\text{iso}}$ ) and the (c,d) remotely forced tropical and extratropical North

1127 Pacific SST effect ( $\text{TPNP}_{\text{iso}}$  minus TPNP).

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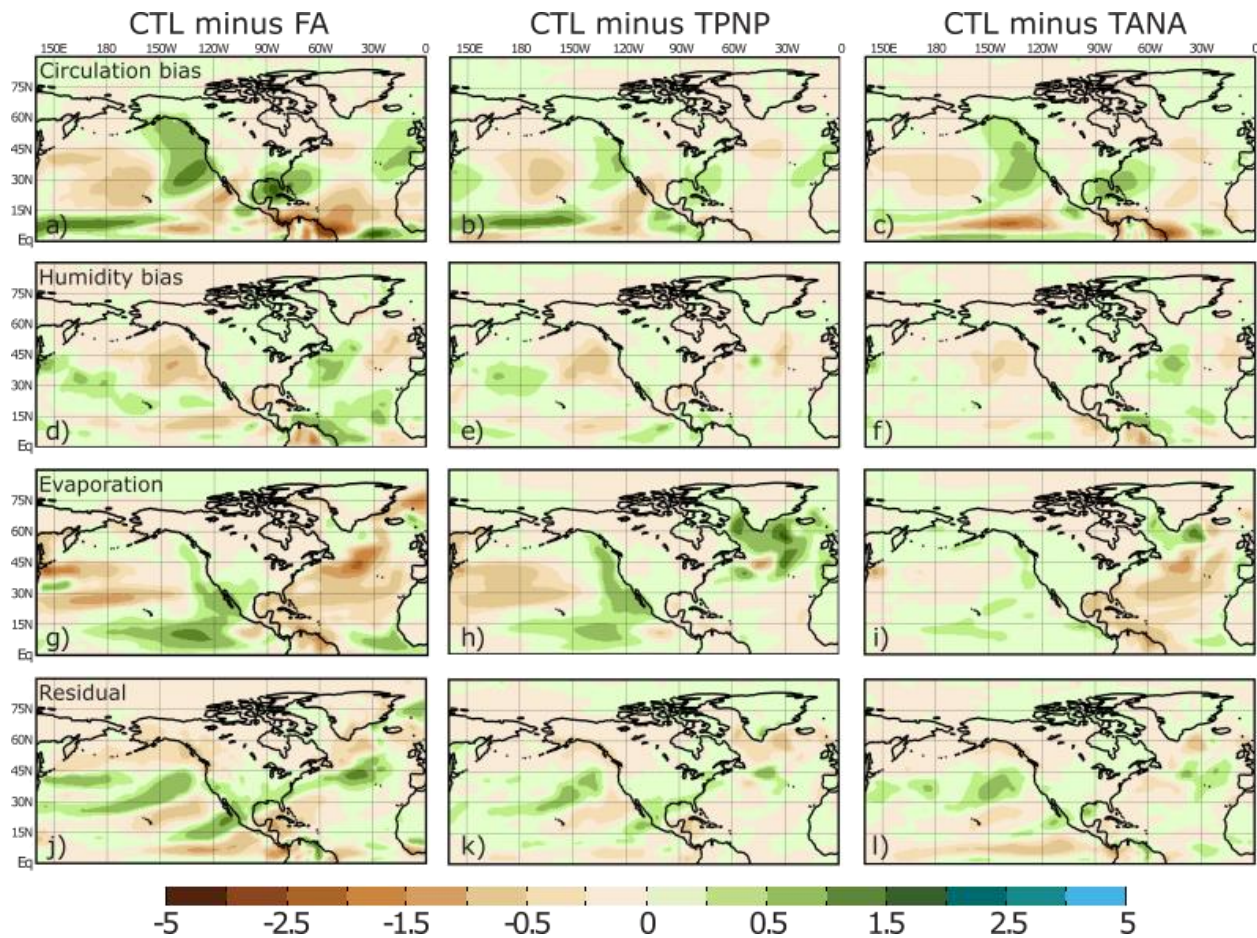
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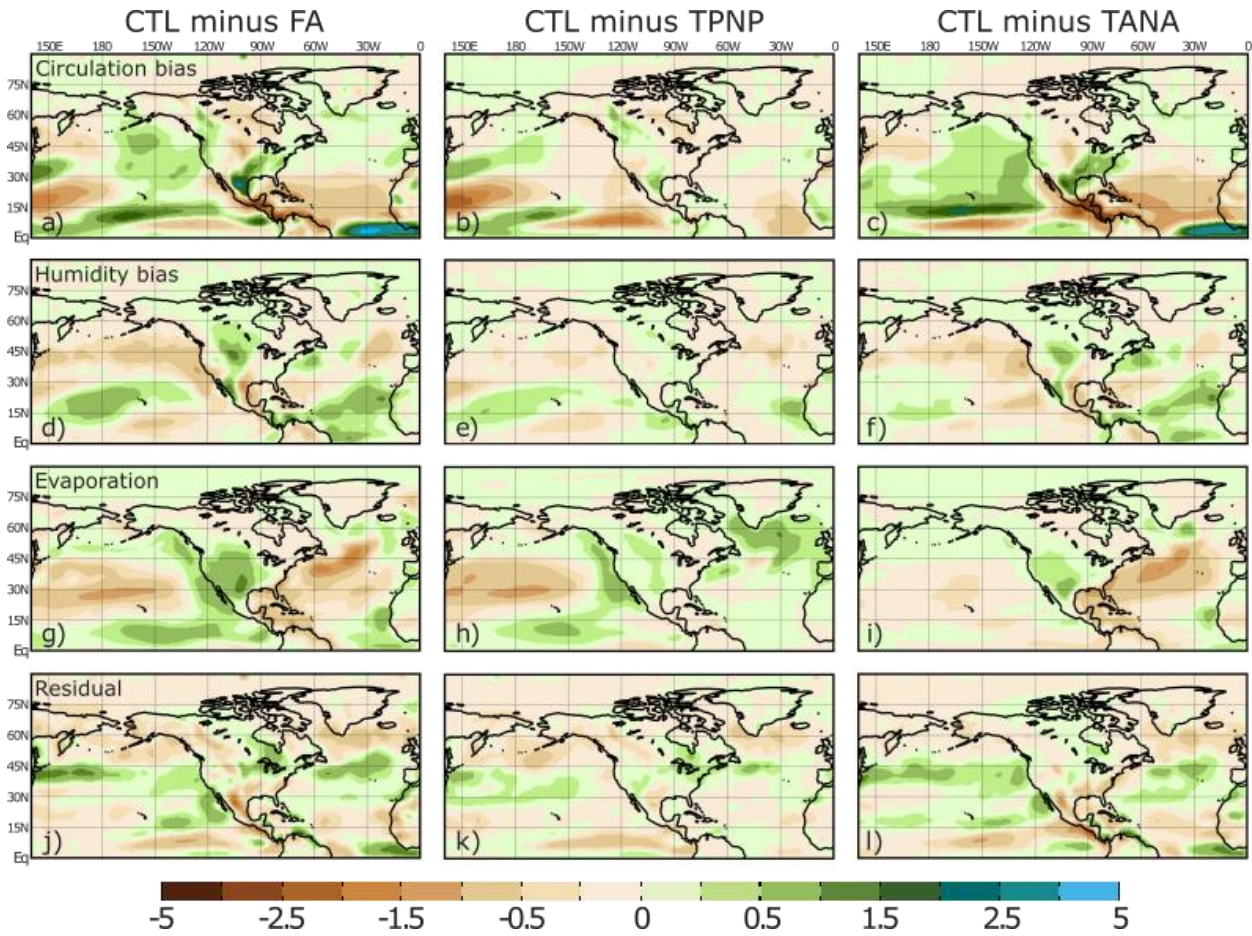




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 1136 FIG. 12. Contributions to the October – March (left column) CTL minus FA, (middle column)  
 1137 CTL minus TPNP, and (right column) CTL minus TANA climatological precipitation  
 1138 differences ( $\text{mm d}^{-1}$ ) attributed to the following terms: (a-c) circulation bias, (d-f) humidity bias,  
 1139 (g-i) evaporation, and (j-l) the residual, calculated as the actual precipitation difference minus the  
 1140 sum of the three terms.

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1148 FIG. 13. As in Fig. 12 but for April – September.

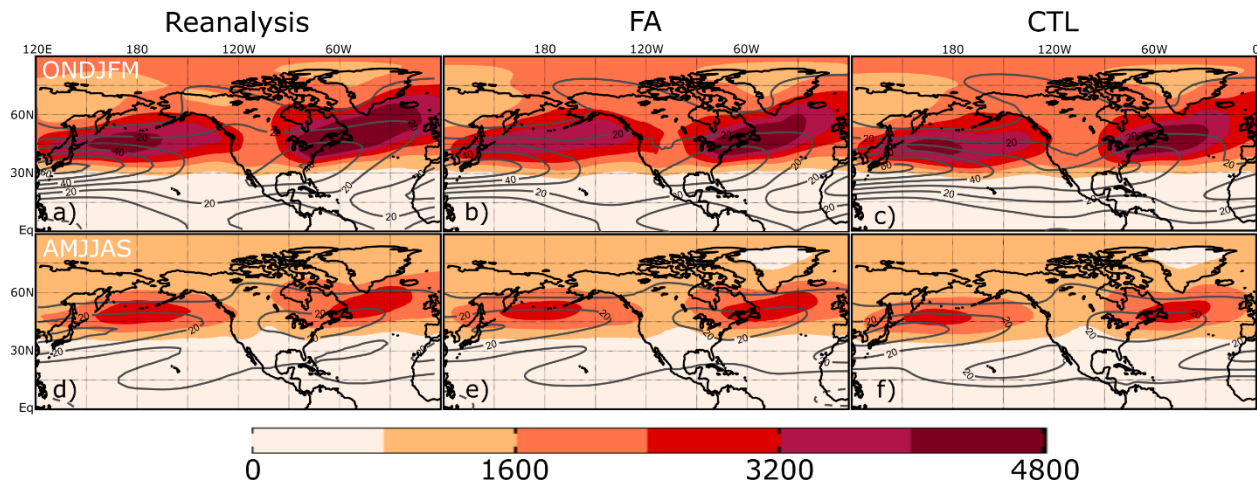
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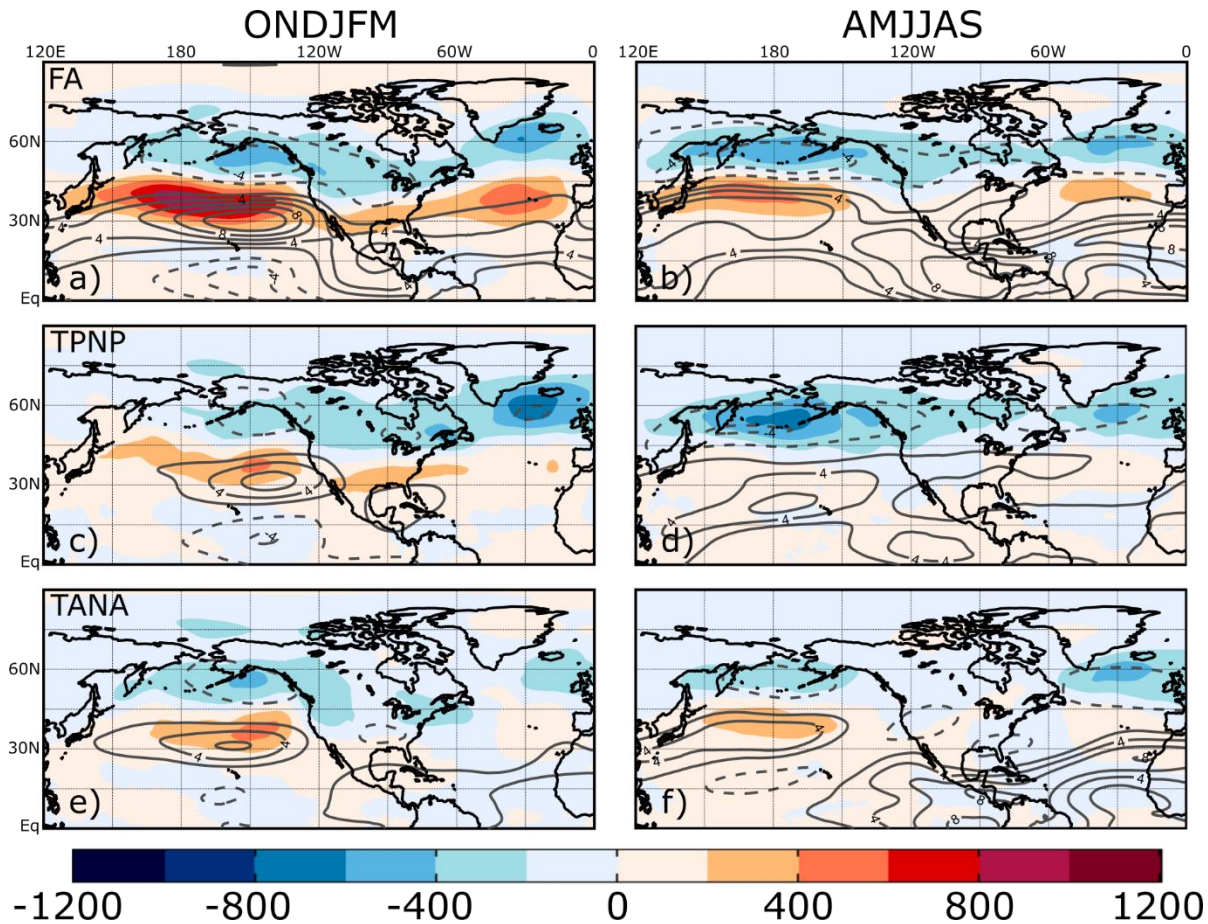
1155 FIG. 14. Climatological storm tracks, as measured by 8-day high-pass filtered 500 hPa geopotential  
 1156 height variance (shading,  $\text{m}^2$ ) and 200 hPa zonal wind (grey contours at an interval of  $10 \text{ ms}^{-1}$ ) for  
 1157 (left) NCEP/NCAR reanalysis data, (center) FA, and (right) CTL simulations in (a-c) October –  
 1158 March and (d-f) April – September.

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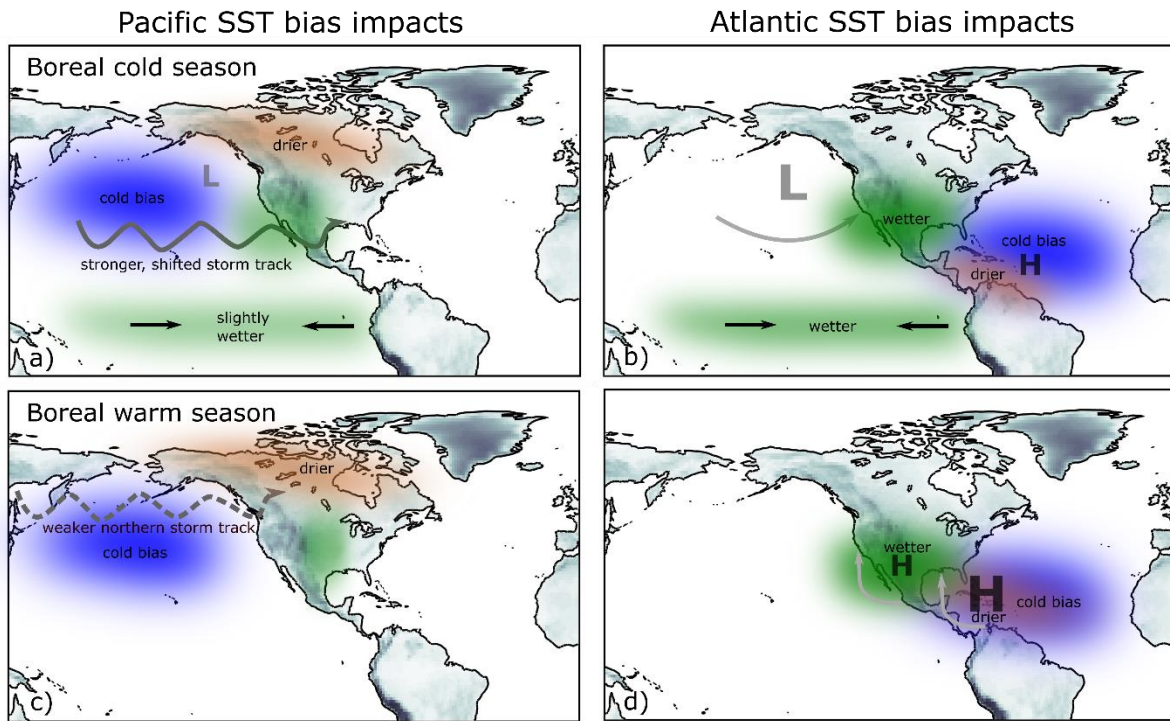




1162 -1200 -800 -400 0 400 800 1200  
 1163 FIG. 15. Differences in climatological storm tracks, as measured by 8-day high-pass filtered 500  
 1164 hPa geopotential height variance (shading,  $m^2$ ), and 200 hPa zonal wind (grey contours at an  
 1165 interval of  $2 \text{ ms}^{-1}$ , zero contour is omitted) for (a,b) CTL minus FA, (c,d) CTL minus TPNP, and  
 1166 (e,f) CTL minus TANA in (left) October – March and (right) April - September.

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1176 FIG. 16. Schematic showing the dominant impacts of Pacific and Atlantic SST biases on North  
1177 American precipitation biases in the boreal cold and warm seasons. In the boreal cold season, (a)  
1178 negative SST biases in the extratropical North Pacific promote a strengthened and southerly shifted  
1179 storm track, which enhances precipitation in the southwestern United States and suppresses  
1180 precipitation in northern Canada. (b) Tropical Atlantic cold SST biases induce circulation changes  
1181 throughout the entire tropics resembling the classic Gill model, with a surface anticyclone in the  
1182 vicinity of the cold bias and low-level convergence and enhanced precipitation in the equatorial  
1183 Pacific. The enhanced tropical Pacific rainfall excites a deepened Aleutian low and enhanced  
1184 moisture transport and precipitation in the southern United States. In the boreal warm season, the  
1185 effects of (c) North Pacific SST biases are modest, but a weaker northern portion of the North  
1186 Pacific storm track promotes drier conditions in northern North America. (d) The cold Atlantic  
1187 SST biases have a much stronger impact, substantially strengthening the western lobe of the North



1188 Atlantic Subtropical High and weakening the thermal low over southern North America. These  
1189 changes enhance the Great Plains low-level jet and moisture transport into southwestern North  
1190 America. Because the SST biases in each basin influence the SST biases in the other basin, the  
1191 total SST bias effects are not limited to the direct effects described here.