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Quantifying atmosphere and ocean origins of North American precipitation variability

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1	Qua	ntifying Atmosphere and Ocean origins of North American Precipitation Variability
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#### 20 Abstract

21 How atmospheric and oceanic processes control North American precipitation variability 22 has been extensively investigated, and yet debates remain. Here we address this question in a 50km-resolution flux-adjusted global climate model. The high spatial resolution and flux 23 24 adjustment greatly improve the model's ability to realistically simulate North American precipitation, the relevant tropical and midlatitude variability and their teleconnections. 25 Comparing two millennium-long simulations with and without an interactive ocean, we find that 26 27 the leading modes of North American precipitation variability on seasonal and longer timescales 28 exhibit nearly identical spatial and spectral characteristics, explained fraction of total variance 29 and associated atmospheric circulation. This finding suggests that these leading modes arise from 30 internal atmospheric dynamics and atmosphere-land coupling. However, in the fully coupled 31 simulation, North American precipitation variability still correlates significantly with tropical 32 ocean variability, consistent with observations and prior literature. We find that tropical ocean variability does not create its own type of atmospheric variability but excites internal 33 34 atmospheric modes of variability in midlatitudes. This oceanic impact on North American 35 precipitation is secondary to atmospheric impacts based on correlation. However, relative to the 36 simulation without an interactive ocean, the fully coupled simulation amplifies precipitation 37 variance over southwest North America (SWNA) during late spring to summer by up to 90%. 38 The amplification is caused by a stronger variability in atmospheric moisture content that is 39 attributed to tropical Pacific sea surface temperature variability. Enhanced atmospheric moisture 40 variations over the tropical Pacific are transported by seasonal mean southwesterly winds into 41 SWNA, resulting in larger precipitation variance.

#### 1. Introduction

Precipitation over North America, a region prone to frequent large-scale droughts 44 45 persisting from a season to multiple decades (e.g., Seager and Ting 2017), has been extensively studied over the past forty years with a goal to assess its predictability and improve predictions. 46 47 Most of the efforts have been focused on the physical understanding of North American 48 precipitation variability. Processes underlying North American (or any regional) precipitation 49 variability can be categorized into two types, external radiative forcing and internal climate 50 dynamics. In this work we focus on the latter, which consists of processes internal to, and 51 interactions among, atmosphere, land, cryosphere and ocean, on seasonal and longer timescales.

52 Precipitation is a meteorological phenomenon and the earliest studies are, as expected, atmosphere-oriented. The most prominent large-scale atmosphere circulation variability that 53 54 have been identified to affect North American precipitation includes the Pacific North America 55 (PNA) pattern (Wallace and Gutzler 1981) and the North Pacific Oscillation (NPO) pattern 56 (Walker 1924). Both patterns represent variations of background mean flow over the North 57 Pacific/America sector. The PNA depicts planetary-scale coherent strengthening or weakening of 58 a trough in the central North Pacific, a ridge over the Rocky Mountains and a trough over 59 southeast North America, while the NPO delineates the meridional movements of the Asian-60 Pacific jet. The PNA and NPO affect North American precipitation by modulating the Pacific 61 storm track and moisture transport into North America (e.g., Leathers et al. 1991; Linkin and 62 Nigam 2008; Liu et al. 2017; Chen et al. 2018; references therein). Although the PNA and NPO 63 are often portrayed as atmosphere circulation variability, they especially the PNA can also be triggered by ocean variability (e.g., Horel and Wallace 1981) and act as a teleconnection to link 64 65 remote ocean impacts to North American precipitation variability (Liu et al. 2017).

66 The tropical ocean's role in North American precipitation variability began to be noticed 67 and understood first by Bjerknes (1966, 1969) and then in the late 1970s and early 1980s, when 68 the now-well-established Rossby wave and teleconnection dynamics were theorized (Hoskins and Karoly 1981) and applied to link the tropical Pacific El Niño-Southern Oscillation (ENSO) 69 70 to meteorological anomalies over North America (Rasmusson and Wallace 1983). Rossby waves 71 excited by ENSO-induced rainfall anomalies in the tropical Pacific propagate poleward and 72 eastward along a great circle of the Earth and generate alternating cyclonic and anticyclonic 73 circulation anomalies that are quasi-stationary and akin to the PNA pattern, with centers of

74 action over the North Pacific, western Canada and the southeastern United States. These 75 circulation anomalies affect the strength and position of the subtropical jet stream over the North 76 Pacific (stronger and further equatorward during warm El Niño events and vice versa during cold 77 La Niña events), which then change the associated storminess, moisture transport and 78 precipitation over North America. This dynamical mechanism has been tested in models (e.g., 79 Palmer and Mansfield 1984, 1986; Suarez 1985) and invoked to explain the observed 80 precipitation variations over North America during major ENSO events (e.g., Ropelewski and 81 Halpert 1986; Trenberth et al. 1988; Palmer and Branković 1989; Trenberth and Branstator 1992; 82 Trenberth and Guillemot 1996; Hoerling and Kumar 2003; Schubert et al. 2009). It has also been 83 shown to operate during non-ENSO seasons such as boreal summer, although the associated 84 atmosphere teleconnection pattern changes with seasonal mean circulation (e.g., Castro et al. 85 2001; and references therein).

86 The ENSO teleconnection mechanism has been expanded by Seager et al. (2003, 2005). They use observations and models to show that ENSO drives a prominent zonally and 87 88 hemispherically symmetric component of precipitation variability in the midlatitudes. El Niño 89 events warm the entire tropical troposphere, which strengthens the subtropical jets in both 90 hemispheres and pulls them equatorward (opposite for La Niña events). These changes in the 91 subtropical jets influence the transient eddy momentum fluxes and the eddy-driven mean 92 meridional circulation, resulting in ascent and more precipitation in the midlatitudes especially 93 over North America (Seager et al. 2005a).

94 On decadal to multidecadal time scales, in the early 2000s, tropical Pacific low-frequency 95 variability was found to affect North American precipitation (e.g., McCabe et al. 2004; Schubert 96 et al. 2004a, b; Seager et al. 2005b). Over the past two decades, tremendous progress has been 97 made, particularly on decadal droughts over North America (see two review papers by Seager 98 and Hoerling 2014; Seager and Ting 2017; and the references therein). The dynamical 99 mechanisms on low-frequency timescales are essentially the same as those on seasonal and 100 interannual timescales as discussed above.

In addition to the tropical Pacific, tropical North Atlantic variability also influences North
American precipitation on seasonal to multidecadal timescales (e.g., Enfield et al. 2001; Sutton
and Hodson 2005, 2007; Wang et al. 2008; Kushnir et al. 2010; Ruprich-Robert et al. 2016,
2018; Johnson et al. 2020). The associated mechanisms are insensitive to timescales (as for the

tropical Pacific), but seasonally dependent (Kushnir et al. 2010). In warm seasons, variations in 105 106 tropical North Atlantic sea surface temperature (SST) change the strength of the North Atlantic 107 subtropical high (warmer SSTs lead to a weaker high), which influences the associated 108 northward moisture transport on its west flank from the Gulf of Mexico and thus precipitation 109 over North America (mainly the United States and northern Mexico). In cold seasons, the above 110 warm-season mechanism still operates; in addition, tropical North Atlantic SST variations also 111 modify the convection over the equatorial Pacific, which then affects North American 112 precipitation in a way similar to the ENSO teleconnection but with a weaker amplitude. Overall, 113 the impacts of tropical North Atlantic on North American precipitation are considerably weaker 114 than those from the tropical Pacific.

115 Despite the significant progress on the physical understanding of North American 116 precipitation variability, debates still remain. For example, the argument that the 1988 summer 117 North American drought was caused by the La Niña event that year (Trenberth et al. 1988; 118 Palmer and Branković 1989) has been refuted by others arguing a fundamental role for remote 119 diabatic heating over the western North Pacific as well as local land feedbacks and atmosphere 120 dynamics (Lyon and Dole 1995; Liu et al. 1998; Chen and Newman 1998; Bates et al. 2001, Seager and Hoerling 2014). The attribution of the infamous 1930s Dust Bowl to the persistent 121 122 cold tropical Pacific and warm tropical North Atlantic (Schubert et al. 2004a, b; Seager et al. 123 2005b; Herweijer et al. 2006) has been augmented by studies arguing for an indispensable role 124 for human-induced land changes and dust aerosol feedbacks in the persistence, intensification 125 and spatial expansion of the drought (Cook et al. 2009, 2011) and a fundamental role for 126 atmosphere random processes in the onset of the drought (Hoerling et al. 2009). At the core of 127 these debates is a lack of agreement on the relative contributions to North American precipitation 128 of the tropical oceans and internal atmosphere dynamics-processes internal or intrinsic to the 129 atmosphere and independent of other components of the earth system.

The role of internal atmosphere dynamics in North American precipitation variability on
seasonal and longer time scales has been emphasized by a few studies (e.g., Hoerling and Kumar
1997; Hoerling et al. 2009; Seager and Hoerling 2014; Stevenson et al. 2014; Seager and Ting
2017; Kumar and Chen 2020). On seasonal to interannual time scales, North American
precipitation anomalies during El Niño events differ significantly from event to event
(Rasmusson and Wallace 1983), and most of these differences have been attributed to internal

atmosphere variability (Hoerling and Kumar 1997). On decadal time scales, Stevenson et al. 136 137 (2017) explore the causes of North American megadroughts by comparing a fully coupled 138 simulation with an atmosphere-only simulation driven by climatological SSTs from the coupled 139 simulation. They find similar intensity and frequency of megadroughts between the two simulations and suggest that the atmosphere alone (still coupled with the land) is able to generate 140 141 North American megadroughts. A recent review paper by Seager and Ting (2017) conclude that 142 the ocean's role in North American decadal droughts is influential, but not paramount, with 143 atmosphere internal variability explaining up to three quarters of the low-frequency precipitation 144 variability. In all of these studies except Stevenson et al., the role of internal atmosphere 145 dynamics is not explicitly assessed, but inferred as a residual between the total precipitation variations and the ocean contributions that are explicitly assessed (often by averaging ensembles 146 147 of atmosphere-only simulations driven by observed monthly SSTs). Even in Stevenson et al., internal atmosphere dynamics are only assessed as a whole in terms of the statistics of North 148 149 American megadroughts. It remains unclear how exactly internal atmosphere dynamics work 150 together with ocean variability to generate North American precipitation variability, especially 151 on seasonal and longer timescales.

152 In this work, we aim to fill this gap by (1) investigating the specific role of internal 153 atmosphere dynamics in North American precipitation variability on seasonal and longer time 154 scales and (2) assessing how atmospheric and oceanic processes play out in the context of North 155 American precipitation variability, that is, how different or comparable the atmosphere-induced 156 and ocean-induced precipitation variability are. We will not only demonstrate the fundamental 157 role of internal atmospheric dynamics and coupling with land in shaping the spatial and spectral 158 characteristics of the leading modes of variability in North American precipitation, but also 159 present a new physical mechanism whereby the tropical oceans affect North American 160 precipitation variability.

161

#### **162 2. Methods**

163 2.1 Model and experiments

In order to cleanly separate the contributions of atmospheric and oceanic processes,
analyzing observations alone is not enough and we have to use climate models. The model in this
work is the Forecast-oriented Low Ocean Resolution (FLOR) flux-adjusted model developed at

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the Geophysical Fluid Dynamics Laboratory (Vecchi et al. 2014). FLOR has a high horizontal 167 168 resolution of approximately 50km for the atmosphere (AM2.5) and land (LM3) components and 169 a horizontal resolution of 1° (telescoping to 0.333° near the equator) for the ocean (MOM4) and 170 sea ice (Sea Ice Simulator) components. It has a standard freely coupled version and a fluxadjusted version. The flux-adjustment in FLOR imposes fixed anomalous enthalpy, momentum 171 172 and fresh water fluxes to the ocean to correct biases in the climatological seasonal cycle of SST, 173 sea surface salinity and wind stress (against 1979-2012 observations). We use the flux-adjusted 174 FLOR because of its much-improved mean climate and variability (see the next section).

175 Experiments are summarized in Table 1. We first introduce two long control simulations, 176 one conducted with the fully coupled FLOR for 3500 years and the other with the uncoupled 177 AM2.5 for 1000 years (note in AM2.5, the atmosphere is still coupled with the land). For 178 simplicity, these control simulations are referred to as FLOR and AM2.5, respectively. Both simulations are driven by preindustrial level atmospheric composition and radiative forcing. In 179 180 AM2.5, the boundary forcing is prescribed as the climatological annual cycle of monthly mean 181 SST and sea ice (concentration and thickness) derived from the last 1000 years of FLOR. This 182 1000-year FLOR simulation is compared with AM2.5. FLOR and AM2.5 share the same 183 radiative forcing and the same climatology at the surface, but differ in that the ocean (including 184 sea ice) is fully active in FLOR but is replaced with an imposed climatological seasonal cycle of 185 SST and sea ice in AM2.5. Precipitation variability in AM2.5 (excluding seasonal cycle) thus 186 arises purely from internal atmosphere processes and coupling with land, while that in FLOR 187 also includes contributions from the ocean (specifically, SST and sea ice variability). He et al. 188 (2018a) showed that the ocean and atmosphere induced precipitation variability is linearly 189 additive and that variability internal to the atmosphere (and land) can be cleanly extracted using 190 atmosphere-only simulations with prescribed climatological SSTs. Analyzing and comparing the 191 millennium-long fully coupled and atmosphere-only simulations allows for a physically clean 192 and statistically robust assessment of the relative contributions of internal atmosphere processes 193 and coupling with land versus the ocean in North American precipitation variability.

194 The remaining simulations are summarized in Table 1 and will be introduced later. We 195 first examine the performance of the FLOR model in simulating North American precipitation 196 and the associated climate variability in the tropics and midlatitudes, in order to build confidence in using the model for investigating North American precipitation variability and the underlyingatmospheric and oceanic processes.

199 2.2 Model performance

200 Precipitation biases over North America, especially over the western mountains, are a 201 common problem in the current generation of climate models (Mejia et al. 2018). The FLOR 202 model, however, substantially reduces these biases owing partly to its high resolution and flux 203 adjustment. A recent paper by Johnson et al. (2020) has explained that the flux adjustment in 204 FLOR reduces SST biases in the tropical Pacific and Atlantic, which improves the climatology 205 of North American precipitation through atmosphere teleconnection mechanisms (including the 206 ENSO teleconnection and those proposed by Kushnir et al. (2010)). Here we highlight the 207 precipitation biases that remain in the flux-adjusted FLOR, against two observational products at 208 0.5° resolution, the Climatic Research Unit timeseries (CRU) datasets version 3.24.01 (Harris et 209 al. 2014) and the Global Precipitation Climatology Centre (GPCC) version 2018 (Schneider et al. 210 2011). Fig. 1b shows that the flux-adjusted FLOR still has significant biases over most of North 211 America, with too much precipitation over the Pacific Northwest and the monsoon region and 212 too little precipitation over central to southeast North America. These positive biases are simulated throughout the year while the negative biases occur mainly during early summer to fall 213 (Fig. 17 in appendix). Compared to the climatology, precipitation variability is better simulated 214 215 in FLOR (Fig. 1c and Fig. 18 in appendix, indicated by more stippling). As will be shown later in 216 the Results section, FLOR realistically simulates the leading modes of North American 217 precipitation variability on seasonal and longer (up to the sampling length) time scales.

218 Tropical climate variability is assessed by comparing FLOR with the Met Office Hadley 219 Centre's sea ice and sea surface temperature dataset (HadISST) version 1.1 (Rayner 2003). 220 FLOR realistically simulates the spatial pattern and amplitude of tropical SST variability, 221 although the ENSO variability is excessively strong in FLOR (Fig. 2a-b). This bias over the 222 Niño3.4 occurs during February to June, while during the rest of the year FLOR simulates a very 223 realistic Niño3.4 variability (Fig. 2c). FLOR also realistically simulates the ENSO spectrum, 224 with a broad interannual peak consistent with observations (Fig. 2d) but seemingly too weak decadal variability. 225

To evaluate how FLOR simulates the ENSO teleconnection and impacts on North
American precipitation, we compute the correlation coefficients between the Niño3.4 index and

228 SST, land precipitation and 200mb geopotential height (GHT200) during December to February 229 (DJF). Observational fields include SST from HadISST, precipitation from CRU and 230 geopotential height from the NOAA-CIRES-DOE Twentieth Century Reanalysis (20CR) version 231 2 (Compo et al. 2011) during 1901-2012. Fig. 3 shows that FLOR simulates the observed global 232 ENSO teleconnection very well, indicated by the similar Rossby wave trains emanating from the 233 tropical Pacific. Over the Pacific/North America sector, FLOR realistically simulates the 234 observed PNA-like upper tropospheric circulation (GHT200) associated with ENSO (Fig. 3a vs 235 3c). In particular, observed correlations between North American precipitation and ENSO are 236 realistically captured in FLOR, with a pattern of positive/negative correlations over 237 southern/northern North America, respectively (Fig. 3b vs 3d). We note that the correlations 238 appear a bit weaker in FLOR than in observations, which can be partly attributed to the 239 uncertainty in the observed teleconnection patterns due to the insufficient sampling lengths (112 240 years, (Deser et al. 2017)).

241 To evaluate FLOR's performance in simulating midlatitude variability over the 242 Pacific/North America sector (20°N-80°N, 100°E-50°W), we conduct an empirical orthogonal 243 function (EOF) decomposition of monthly mean sea level pressure (SLP) variability and 244 compare the leading modes between FLOR and the 20CR dataset (Fig. 4, top two rows). Clearly, 245 FLOR faithfully simulates the top three EOF modes in terms of the spatial pattern and the 246 explained variance. Accounting for about 24% of total variance in both FLOR and 20CR, the 247 first EOF mode (EOF1) features three centers of action over the Aleutian Low, the Arctic Ocean 248 and the Gulf Stream, respectively and likely reflects the surface expression of the PNA (which in 249 literature is defined by middle to upper tropospheric geopotential heights). Note that EOF1 is 250 nearly identical to the ENSO teleconnection pattern in SLP (black contours in Fig. 4 upper left 251 two panels), suggesting that EOF1 is related to the ENSO variability. However, this EOF mode 252 does not own its existence to ENSO, because it also appears as the first EOF mode in AM2.5 253 without the interactive varying ocean (Fig. 4 bottom left) and explains a similar amount of 254 monthly SLP variance, 22.7% in AM2.5 vs 24.2% in FLOR. This result supports the familiar 255 concept that ENSO affects the extratropics by projecting onto internal modes of atmospheric 256 variability (e.g., Simmons et al. 1983; Palmer and Mansfield 1984; Geisler et al. 1985; Palmer 257 1993; Lau and Nath 1994; Saravanan 1998; Hoerling and Kumar 2002; Barsugli and 258 Sardeshmukh 2002; Dai et al. 2017; Henderson et al. 2020). A similar story holds true for the

second and third EOF modes, with EOF2 being the North Pacific Oscillation (Walker 1924) and
EOF3 being an east-west seesaw pattern over the Aleutian Islands. Both are realistically
simulated in FLOR.

In summary, as shown by Johnson et al. (2020), the flux-adjusted high-resolution FLOR model substantially improves the simulation of North American precipitation climatology compared to its standard version and other peer models. Here we demonstrate that FLOR also realistically simulates climate variability both in the tropical oceans and midlatitudes as well as the ENSO teleconnection. Next, we move on to the focus of this work—North American precipitation variability on seasonal and longer time scales, evaluate how it is simulated in FLOR and investigate how it is shaped by atmospheric and oceanic processes.

269

#### **3. Results**

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3.1 Dominant modes of variability in North American precipitation

272 To examine North American precipitation variability on seasonal and longer time scales, 273 we focus on the dominant modes of variability in monthly precipitation calculated from an EOF 274 decomposition. Note that the mean annual cycle has been removed from precipitation prior to the 275 EOF decomposition. In observations and FLOR, the top two EOF modes are significantly 276 separated from each other (and also other modes) based on the North test (North et al. 1982) and 277 reflect the strong precipitation over the southeastern United States and the Pacific Northwest, 278 respectively (recall Fig. 1a). In Fig. 5, for the first matching EOFs, both the observed and FLOR 279 modes have the largest loading over the southeastern United States and explain about 9.0% and 280 7.8% of their total variance over North America, respectively. In Fig. 6, for the second matching 281 EOFs, the observed and FLOR modes are dominated by two largest loadings along the Pacific 282 Northwest and explains about 6.4% of the variance in observations and 8.8% in FLOR. The 283 pattern correlation between observations and FLOR is 0.77 for the EOF modes in Fig. 5 and 0.87 284 in Fig. 6. Note that the order of the two modes is reversed between observations and FLOR, with 285 the first (second) mode in observations being the second (first) mode in FLOR. The reversed 286 order in FLOR is due to the relatively large biases in regions of strong precipitation, namely, the 287 Pacific Northwest (more precipitation and larger variance, thus EOF 1) and the southeastern 288 United States (less precipitation and weaker variance, thus EOF 2) (recall Fig. 1c). In addition to 289 the spatial pattern, the principal components (PC) associated with these two EOF modes also

share similar behavior between observations and FLOR: their auto-correlation function decreases to almost zero at the 1-month lag (little or no persistence) and their spectrum is basically flat, both of which are characteristic of a white noise process. Overall, FLOR realistically simulates the spatial and spectral characteristics of the observed dominant modes of variability in monthly mean precipitation over North America. The reversed order of the two leading modes in FLOR and observations suggests that the relative contributions of the two leading modes to total precipitation variability are also reversed, which should be kept in mind for the rest of the paper.

297 AM2.5 without the interactive ocean simulates virtually the same top two EOF modes as 298 the fully coupled FLOR (Fig. 5 and 6). For both EOF modes, the spatial structure is nearly 299 identical between AM2.5 and FLOR with a pattern correlation of 0.99. The percentage of total 300 variance explained by each EOF mode is also very similar, 8.5% (AM2.5) vs 8.8% (FLOR) for 301 EOF 1 and 7.7% (AM2.5) vs 7.8% (FLOR) for EOF 2. The associated PCs exhibit little or no 302 persistence and have a white-noise spectrum. These similarities suggest that the spatial and 303 spectral characteristics of the two dominant modes of variability in monthly precipitation over 304 North America are not sensitive to the ocean in the FLOR model and are determined by internal 305 atmospheric dynamics and coupling with land.

306 In addition to monthly precipitation, we have conducted the same EOF decomposition on 307 seasonal mean precipitation and low-pass-filtered monthly precipitation with various cutoff 308 periods (1, 3 and 5 years) and also over various sub-regions of North America. These results are 309 not shown but summarized here. Overall, the spatial pattern of leading EOF modes follows the 310 pattern of strong precipitation (as expected, because strong precipitation implies strong variance 311 and the EOF decomposition by design maximizes the variance explained by the leading modes). 312 The main quantitative difference from the results based on monthly precipitation is in the percentage of explained variance, which increases by various extents depending on the season 313 314 and the sub-region. However, regardless of seasons and sub-regions, the leading EOF modes of 315 FLOR and AM2.5 still share very similar spatial patterns and white-noise behavior. These 316 similarities further suggest that internal atmospheric dynamics and coupling with land dominate 317 the spatial and spectral characteristics of the leading modes of precipitation variability over 318 North America not only on high-frequency time scales (in monthly precipitation) but also on 319 low-frequency (interannual to decadal) time scales. Next, we return to monthly fields and present 320 the physical processes associated with the two leading EOF modes.

321 To examine the large-scale atmosphere circulation and possible SST anomalies 322 associated with the top two EOF modes, we calculate the correlation coefficients between the 323 associated PCs and SLP, GHT200 and surface temperature. For the EOF mode with a strong 324 loading over the southeastern United States (Fig. 5), the associated large-scale atmosphere 325 circulation is very similar over North America (where the EOF analysis is performed, indicated 326 by the magenta box in each panel) among observations, FLOR and AM2.5. In Fig. 7, the large-327 scale pattern of SLP correlation (contours in top row) includes two anticyclones centered around 328 east of the Caribbean and off the Canada-Alaska coast and one cyclone in between over the U.S.. 329 The GHT200 correlation (shading in bottom row) exhibits a similar pattern (but slightly tilted 330 westward), suggesting an equivalent barotropic structure for the associated circulation over 331 North America. The implied geostrophic winds and moisture advection are consistent with the 332 EOF precipitation anomalies. In particular, the moist southerlies from the Gulf of Mexico 333 support the strong precipitation over the southeastern United States, while the wet-dry dipole 334 along the Pacific Northwest (weak in observations but strong in the model simulations) agrees 335 with the wet oceanic southerlies and dry inland northerlies that straddle the anticyclone there.

336 For the correlation with surface temperature, the local patterns over North America are 337 also very similar among observations, FLOR and AM2.5, and largely consistent with the 338 anomalous advection of mean surface temperature (that is, southerlies lead to warming and vice 339 visa). Remotely over the ocean, the pattern of SST correlation is similar between observations 340 and FLOR (AM2.5 has no SST variability by design). In particular, the EOF mode shown in Fig. 341 5 is accompanied with positive SST (also increased GHT200 aloft) anomalies over the tropical 342 Pacific and Indian Ocean, suggesting a role for the tropical oceans. However, the weak 343 magnitude of the correlation implies that the role of the tropical oceans is very limited for the 344 variability in monthly precipitation over North America. AM2.5 without the interactive ocean 345 simulates the associated surface temperature and atmospheric circulation over North America 346 (and actually has opposite sign GHT200 anomalies in the tropics), suggesting that the physical 347 processes underlying the precipitation EOF mode shown in Fig. 5 can arise solely from internal 348 atmosphere dynamics and coupling with land.

Similar conclusions are found for the EOF mode shown in Fig. 6 reflecting strong
 precipitation anomalies over the Pacific coast of North America. In Fig. 8, the associated
 atmosphere circulation over North America features a strong equivalent-barotropic anticyclone

352 over the Pacific Northwest in observations, FLOR and AM2.5. To the west of the anticyclone, 353 the associated geostrophic southerlies bring warm moist air from the ocean to British Columbia 354 and Alaska. To the south of the anticyclone geostrophic easterlies bring cold dry air from inland 355 to the US west coast. This results in the dipole pattern in both surface temperature (Fig. 8, 356 shading in top row) and precipitation (Fig. 6). Note that the anticyclone is part of an anomalous 357 circulation that reflects changes in the strength of midlatitude stationary eddies (Fig. 8, contours 358 in bottom row). This EOF mode also exhibits some positive but weak correlations with tropical 359 SSTs, mostly in FLOR. We note that the large-scale atmosphere circulation associated with the 360 two leading EOF modes has three centers of actions over the North Pacific/American sector. 361 These centers of action resemble but are shifted relative to those of the PNA (the upper left two 362 panels in Fig. 4). This difference in atmosphere circulation is consistent with the fact that both of 363 the leading EOF modes in precipitation are different from the precipitation pattern associated 364 with the PNA (not shown but is nearly identical to the ENSO teleconnection pattern in Fig. 3).

365 Overall, despite the reversed order of the two leading EOF modes, FLOR faithfully reproduces the spatial and spectral features of the observed dominant variability in monthly 366 367 precipitation over North America and the associated large-scale atmosphere circulation and 368 surface temperature. Comparing the fully coupled FLOR and AM2.5 without the interactive 369 ocean, we find that their dominant modes of monthly (and seasonal) precipitation variability over 370 North America have nearly identical spatial structures, explained fraction of total variance, 371 spectral behavior (no persistence and white-noise spectrum) and the associated large-scale 372 atmosphere circulation and local surface temperature anomalies. These similarities suggest that 373 the dominant variability in North American precipitation arises from internal atmosphere 374 dynamics and coupling with land, and does not require SST variability to exist. However, SST 375 variability, particularly in the tropics, still affects North American precipitation, as demonstrated 376 in the literature and indicated here by the positive correlation with tropical SSTs in FLOR and 377 observations. The weak magnitude of correlation suggests that the impacts of the tropical oceans 378 are very limited. The results that the ocean does not change the spatial and spectral 379 characteristics of North American precipitation leading modes of variability suggest that SST 380 variability does not create its own modes of atmosphere variability but rather excites modes of 381 variability that already exist in the atmosphere. This interpretation is also supported by the 382 results in Fig. 4, where the leading mode of midlatitude variability in AM2.5 without the

383 interactive ocean is nearly identical to the ENSO teleconnection pattern in the fully coupled 384 FLOR and observations. Note that these results are consistent with the familiar concept that SST 385 variability affects the atmosphere by projecting onto internal modes of atmospheric variability. 386 We emphasize here that SST variability does not significantly change the basic characteristics of 387 the large-scale atmospheric variability and its influence only accounts for a small portion of the 388 total North America precipitation variability. Despite its limited impacts, SST variability is still 389 the main source of long-term predictability for North American precipitation variability beyond 390 the time scales of internal atmosphere dynamics.

3.2 Precipitation variance over SWNA

To quantify the impacts of the ocean on year-to-year variance of precipitation over North America, we compute the fractional change of monthly precipitation standard deviation at each grid point for each calendar month between FLOR and AM2.5 by

$$\frac{STD_{FLOR}-STD_{AM2.5}}{STD_{FLOR}} \times 100\%,$$

396 where  $STD_{FLOR}$  and  $STD_{AM2.5}$  are the year-to-year standard deviation of monthly precipitation 397 evaluated at the same grid point for the same calendar month in FLOR and AM2.5, respectively. 398 In Fig. 9, the impacts of the ocean on precipitation variance are not significant (at 1% level) for 399 most of North America throughout the year. The largest significant fractional increase of 400 precipitation variance due to the ocean occurs over southwest North America (SWNA) during 401 May to about September, with magnitudes up to 90% over parts of SWNA especially the North 402 American monsoon region during June and July. In contrast, during winter (November to 403 February) when ENSO variability peaks (recall Fig. 2c), the fractional increase of precipitation 404 variance is relatively weak and only significant over smaller areas of SWNA.

405 How does the ocean amplify precipitation variance over SWNA during the late spring 406 through summer? To answer this question, we examine the factors that affect precipitation 407 variability, including evaporation and atmospheric moisture convergence. The latter depends on circulation and moisture variability, which we will focus on. Similar to precipitation, we 408 409 compute the fractional change in year-to-year standard deviation of these two factors between 410 FLOR and AM2.5. The metrics used are monthly 500mb pressure velocity (Omega500) and 411 column-integrated water vapor (CWV) content. Over SWNA, no systematic significant changes 412 are found in the standard deviation of monthly Omega500 between FLOR and AM2.5 413 throughout the year (Fig. 10), which suggests that the interactive ocean in FLOR does not

- 414 significantly affect the variance of atmosphere divergent circulation over SWNA. In contrast, the
- 415 standard deviation in CWV is significantly enhanced over SWNA during May to about October
- 416 (Fig. 11), similar to the changes in precipitation standard deviation (Fig. 9). Together these
- 417 results suggest that the amplified precipitation variance over SWNA in FLOR relative to AM2.5
- 418 is caused by the enhanced variability in atmosphere moisture content.

419 How does the interactive ocean enhance CWV variability over SWNA and why is this enhancement mainly during late spring to summer but not winter? Our interpretation is that the 420 421 seasonal mean atmosphere circulation is transporting enhanced CWV anomalies from surrounding oceans into SWNA. This is supported in FLOR by the vertically-integrated fluxes of 422 specific humidity standard deviation by mean winds,  $\int \vec{u} \cdot STD(q')dp$ , where  $\vec{u}$  is climatological 423 424 monthly winds, STD(q') is the standard deviation of monthly specific humidity anomalies and 425 the integral is done for the entire atmosphere. This quantity can also be interpreted as the 426 vertically-integrated mean circulation weighted by moisture variability and thus reflects the 427 mean circulation of the lower troposphere where most of the moisture is. During late spring to 428 early fall (around May to October), CWV variability is significantly enhanced by SST variability 429 over the tropical to midlatitude North Pacific, which extends northeastward into the SWNA 430 region following the mean southwesterly winds (see vectors in Fig. 11). In contrast, during 431 winter (around December to March), the enhanced CWV variability over the northeastern Pacific 432 is confined to the subtropics (presumably owing to the colder atmosphere and smaller moisture capacity in winter) and the mean northwesterly winds are from ocean regions where CWV 433 434 variability appears insensitive to SST variability, both of which are unfavorable to enhance CWV 435 variability over SWNA.

436 Which parts of the ocean are responsible for the intensification of precipitation variability 437 over SWNA? To answer this question, we conduct additional experiments (Table 1) that are 438 similar to the AM2.5 simulation but with time varying monthly SSTs from FLOR imposed in 439 different oceans during different times (note that these experiments have first been reported in a 440 companion study focusing on synoptic extreme precipitation in June over SWNA (Zhang 2020)). 441 Four experiments are conducted, each 600 years long (corresponding to FLOR 2601-3200). The 442 first one has monthly SSTs prescribed over the entire globe (for all months, hereafter, Globe 1-443 12) to test whether this technique (AM2.5 imposed with monthly SSTs) can reproduce the FLOR 444 precipitation variance over SWNA given that interactive ocean-atmosphere coupling may matter

445 (e.g., Bretherton and Battisti 2000; He et al. 2017, 2018b). The other three have monthly SSTs prescribed between 35°N and 35°S (with a 5° linear buffer zone) in the entire tropics for all 446 447 months (Tropic 1-12), the tropical Pacific for October to May (Pacific 10-5) and the tropical 448 Atlantic for May to October (Atlantic 5-10), respectively. The reason for prescribing SSTs in 449 Pacific 10-5 and Atlantic 5-10 only during the stated months is that SST correlations with the 450 SWNA precipitation are only significant during the stated months over the respective ocean 451 basins (Fig. 19 in appendix). Tropic 1-12 can test the role of SSTs in the tropics versus 452 extratropics, while Pacific 10-5 and Atlantic 5-10 can test the role of SSTs in the tropical 453 Pacific from October to May and in the tropical Atlantic from May to October, respectively. We 454 note that, in Pacific 10-5 and Atlantic 5-10, the May and October SSTs are extrapolated into 455 their neighbor months such that September and June in Pacific 10-5 and April and November in 456 Atlantic 5-10 also have anomalous SST forcing (by default in all AM2.5 simulations, monthly 457 SST values are placed in the middle of each month before interpolation onto the model time 458 step). These four experiments allow us to identify the oceans responsible for the amplification of 459 precipitation variance over SWNA.

460 Compared to AM2.5 without the interactive ocean, Globe 1-12 generally well reproduces the month-to-month variance (i.e., average of the year-to-year variance over twelve calendar 461 months) of both precipitation and CWV over SWNA in FLOR, albeit a bit weaker and confined 462 463 to lower latitudes (Fig. 12a-d). These similarities also hold true for the year-to-year variance (Fig. 20-21 in appendix), which justifies the experimental technique to pin down the oceans 464 responsible for the enhanced precipitation variance over SWNA in FLOR. Tropic 1-12 without 465 466 extratropical SST variability also well reproduces the precipitation and CWV variance over 467 SWNA in FLOR (Fig. 12e-f for month-to-month variance and Fig. 22-23 for year-to-year 468 variance), suggesting that the tropical SST variability is responsible for the enhanced 469 precipitation variance over SWNA in FLOR.

For the total month-to-month variability, Pacific\_10-5 (Fig. 12g-h) is able to enhance the variance in both CWV and precipitation over most of SWNA, while Atlantic\_5-10 (Fig. 12i-j) only enhances the CWV variance locally in the tropical Atlantic but neither the CWV nor the precipitation variance over SWNA (except southern Mexico and northern central America). For the year-to-year variability, Pacific\_10-5 enhances the precipitation (Fig. 13) and CWV (Fig. 14) variance over SWNA mainly during May and June (recall the extrapolation of SST forcing from

- 476 May to June), similar to the enhancement in FLOR, Globe 1-12 and Tropic 1-12. In contrast,
- 477 Atlantic 5-10 does not simulate systematic enhancement of precipitation (Fig. 15) and CWV
- 478 (Fig. 16) variance over SWNA during May to October when the SST forcing is imposed in the
- tropical Atlantic. Together, these results suggest that it is the tropical Pacific that accounts for the
- 480 enhanced precipitation variance over SWNA in FLOR relative to AM2.5.

481 The lack of impacts from the tropical Atlantic on the SWNA precipitation variance seems 482 inconsistent with previous studies (Kushnir et al. 2010; Johnson et al. 2020). In particular, 483 Johnson et al. (2020) have attributed most of the climatological precipitation biases over 484 southern North America in the standard FLOR model to the strong SST biases in the tropical 485 North Atlantic. This paradox can be explained by the different magnitudes of SST biases and 486 variability in the tropical North Atlantic relative to the tropical Pacific. The SST biases in the 487 standard FLOR model are much larger in the tropical North Atlantic (about 1-3K) than the 488 tropical Pacific (less than 1K, see Fig. 2 in Johnson et al.), while the total month-to-month 489 variability in the flux-adjusted FLOR is much stronger in the tropical Pacific than the tropical 490 Atlantic (Fig. 2b). Therefore, our results from Atlantic 5-10 should be interpreted as that SST 491 variability in the tropical Atlantic is relatively weak and unable to significantly affect the 492 variance of atmosphere moisture content and thus precipitation over SWNA. The enhancement 493 of the SWNA precipitation variance in FLOR relative to AM2.5 is mainly caused by the tropical 494 Pacific through the new mechanism proposed in this work.

495 We point out an important distinction between our new mechanism and the classic ENSO 496 teleconnection mechanism. In the ENSO teleconnection mechanism, tropical Pacific SST 497 anomalies drive anomalous atmosphere circulation over SWNA that is capable of inducing 498 anomalous precipitation. Our mechanism does not directly involve changes in atmosphere 499 circulation over SWNA, but only involve enhanced atmosphere moisture content variability, that 500 is, larger atmosphere moisture content anomalies. A new implication from our mechanism is that 501 an interactive ocean is crucial to simulate and predict the amplitude of precipitation variability 502 (i.e., precipitation intensity) over SWNA.

503

504 **4. Summary and Discussion** 

We have quantified the atmospheric and oceanic contributions to North American
precipitation variability on seasonal and longer time scales in the FLOR model. FLOR features a

- 17 -

507 50km resolution in its atmosphere/land components and flux adjustment to correct biases in 508 mean SST; as a result, it greatly reduces the pervasive biases in North American precipitation 509 that have plagued climate models of the same generation. FLOR also realistically simulates 510 tropical and midlatitude climate variability as well as the ENSO teleconnection, all of which are 511 critical for a reliable quantification of the atmospheric and oceanic contributions to North 512 American precipitation variability.

513 Comparing two millennium-long simulations with and without an active ocean (FLOR 514 and AM2.5), we find that the dominant modes of variability in North American monthly 515 precipitation between the two simulations share a nearly identical spatial structure, explained 516 fraction of total variance and white-noise spectra with no persistence. Furthermore, the 517 associated large-scale atmosphere circulation and surface temperature anomalies over North 518 America are also very similar in terms of both spatial pattern and magnitude. These similarities 519 suggest that the dominant modes of North American precipitation variability do not owe their 520 existence to the ocean, but rather arise from internal atmosphere processes and coupling with 521 land.

522 In the fully coupled FLOR, however, the dominant modes of North American 523 precipitation variability are still significantly correlated with the tropical oceans, especially the 524 tropical Pacific and Indian Oceans. These correlations are consistent with observations and 525 previous studies that have demonstrated a role for the tropical oceans. How the tropical oceans 526 affect extratropical climate variability has been long debated, with some arguing that tropical 527 variability exerts its own unique impacts by driving atmospheric circulation anomalies that are 528 distinct from internal modes of atmospheric variability (e.g., Straus and Shukla 2000) but others 529 arguing that tropical variability merely excites modes of variability that already exist in the 530 atmosphere (e.g., Simmons et al. 1983; Palmer and Mansfield 1984; Geisler et al. 1985; Palmer 531 1993; Lau and Nath 1994; Saravanan 1998; Hoerling and Kumar 2002; Barsugli and 532 Sardeshmukh 2002; Dai et al. 2017; Henderson et al. 2020). Here our results on North American 533 precipitation variability support the latter argument. In particular, the ENSO teleconnection 534 pattern over the Pacific/North America sector in FLOR appears in AM2.5 without the interactive 535 ocean as the leading mode of midlatitude variability. This corroborates the idea that tropical climate variability affects the extratropics by exciting modes of variability that already exist in 536 537 the atmosphere. Taken together, our modeling results suggest that internal atmosphere dynamics

and coupling with land determine the spatial and spectral characteristics of the leading modes of
North American precipitation variability and dictate their explained fraction of total variance,
while the tropical oceans contribute by exciting the same atmosphere dynamical processes.

541 Although the tropical oceans do not change the spatial and spectral characteristics of the 542 leading modes of precipitation variability over the entire North America, they significantly 543 enhance the variance of monthly precipitation over SWNA in FLOR relative to AM2.5. This 544 enhancement in FLOR is seasonal and occurs mainly during late spring through summer when 545 the ENSO variability is weakest (Fig. 2) and the correlation between the tropical Pacific SST and 546 SWNA precipitation is weak or not significant (Fig. 19 in appendix). Examining the two factors 547 critical for land precipitation, we find significant enhancement in the variance of atmosphere 548 moisture content but not of divergent circulation over SWNA. The enhanced variability of 549 atmosphere moisture content over SWNA is traced down to the tropical Pacific, where SST 550 variability amplifies atmosphere moisture content variability, which is then transported into the 551 SWNA region during late spring to summer by the seasonal mean southwesterly winds. This 552 interpretation is different from the classic ENSO teleconnection mechanism not only in that it 553 operates mainly during non-ENSO seasons, but more importantly, because it involves a different 554 pathway of influence from the northeastern tropical Pacific to SWNA that is associated with the 555 mean circulation and moisture anomalies as opposed to the dominance of anomalous circulation 556 in the classic ENSO teleconnection mechanism. Therefore, our interpretation is a new 557 mechanism, in addition to the classic ENSO teleconnection, via which the tropical Pacific Ocean 558 variability affects precipitation over SWNA. Note that this mechanism has been invoked in a 559 companion study (Zhang 2020) to explain how the tropical Pacific intensifies extreme rainfall 560 over parts of SWNA in June. Here we extend the mechanism from synoptic time scales to 561 seasonal and longer time scales.

We point out that the amplification of the SWNA precipitation variability by the tropical oceans has been noted by Seager et al. (2014, see their Fig. 5 and the relevant discussions), but their presumed interpretation is the classic ENSO teleconnection. In addition, projected future amplification in hydroclimate variability (Seager et al. 2011; Pendergrass et al. 2017) has been attributed to global warming-induced increases in *mean* atmosphere moisture content via the thermodynamic Clausius-Clapeyron relationship. Our work differs from those projection studies in that we focus on internal climate variability for a constant radiative forcing with a steady atmosphere moisture content climatology. The variability in atmosphere moisture content (over
the ocean) comes mostly from a dynamical redistribution (i.e., convergence or divergence) of
atmosphere moisture induced by SST variability.

572 SST variability in the tropical Atlantic does not significantly contribute to the amplification of precipitation variance over SWNA (except for southern Mexico and northern 573 574 central America), in contrast to previous studies demonstrating a role for the tropical North 575 Atlantic. This paradox is likely attributed to the weak amplitude of the tropical Atlantic SST 576 variability, which is unable to enhance the variance of atmosphere moisture content over SWNA. 577 We emphasize that the tropical Atlantic SST variability can still affect the precipitation 578 variability over SWNA through other dynamical processes (Kushnir et al. 2010; Johnson et al. 579 2020).

580 The contributions of atmospheric and oceanic processes to North American precipitation 581 variability on seasonal and longer time scales in FLOR are summarized as follows. The spatial 582 and spectral characteristics of the dominant modes of variability in North American monthly 583 (and seasonal) precipitation along with their explained fraction of total variance are controlled by 584 internal atmosphere processes and coupling with land. Ocean variability, mainly from the 585 tropical oceans, contributes to North American precipitation variability in two ways. First, on 586 continental scales, it dynamically excites the same internal atmosphere processes as above (to 587 some extent, as beating a drum excites its normal modes), and thus does not change the spatial 588 characteristics of the leading modes. This impact is secondary compared to internal atmosphere 589 processes and unable to significantly modify the leading modes' white-noise spectrum and 590 explained fraction of total variance. Nonetheless, this ocean impact provides potential long-term 591 predictability for North American precipitation beyond the timescales limited by internal 592 atmosphere processes. Second, on regional scales, SST variability in the tropical Pacific 593 amplifies atmosphere moisture content variability, which during late spring to summer is 594 transported by mean southwesterly winds into the SWNA region and enhances the variance of 595 precipitation over SWNA. This enhancement implies that the tropical Pacific is required for a 596 reliable simulation and prediction of the intensity of precipitation over SWNA.

597

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- 605

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## 822 Table 1. Simulations.

Experiments	Forcing	Length (yr)
FLOR	preindustrial-level atmospheric composition and radiation	3500
	(the same for all experiments), flux adjustment	
AM2.5	driven by climatological annual cycle of SST and sea ice	1000
	from last 1000 years of FLOR	
Globe_1-12	as AM2.5, but driven by monthly varying SSTs from 600	600
	years of FLOR over entire globe and for all months	
Tropic_1-12	as AM2.5, but driven by monthly varying SSTs from 600	600
	years of FLOR over the tropics (35°N and 35°S) and for	
	all months	
Pacific_10-5	as AM2.5, but driven by monthly varying SSTs from 600	600
	years of FLOR over the tropical Pacific (35°N and 35°S)	
	and only from October to May	
Atlantic_5-10	as AM2.5, but driven by monthly varying SSTs from 600	600
	years of FLOR over the tropical Atlantic (35°N and 35°S)	
	and only from May to October	

823 Note that preindustrial radiative forcing is nearly in balance at top of the atmosphere and thus

824 more suitable for millennium-long steady state simulations than present-day conditions.

825

## 826 Figures





829 climatology; FLOR biases (FLOR – CRU) in (b) climatology and (c) standard deviation.

830 Stippling is a measure of insignificance, indicating at least one of CRU and GPCC is inside the

- range of a synthetic 33-member FLOR ensemble, which is constructed by sampling the 1000-
- 832 year FLOR simulation with a 30-year (to mimic 1981-2010) non-overlapping period.
- 833



Fig. 2 Tropical SST variability (°C) in FLOR versus HadISST 1870-2018. SST standard deviation (shading) and climatology (white contours) in HadISST (a) and FLOR (b); monthly standard deviation (c) and spectrum (d) of the Niño3.4 index (SST time series averaged over 170°W-120°W,5°N-5°S). The spectrum is normalized by the variance of the Niño3.4 index. X-axis has a unit of calendar month in (c) and frequency of per month in (d), respectively.



Fig. 3 ENSO teleconnection in observations 1901-2012 (a, b) and FLOR (c, d). Teleconnection
is assessed during DJF as correlation coefficients between the Niño3.4 index and SST (shading),
200mb geopotential height (black contours) in (a, c), land precipitation in (b, d). Here 112 years
are used for observations, while 1000 years are used for FLOR. Random sampling of 112-year
segments from FLOR leads to small changes in correlation pattern and magnitude (not shown).





849 Fig. 4 Midlatitude variability of monthly SLP (hPa) in observations (top row), FLOR (middle 850 row) and AM2.5 (bottom row). The leading three EOF modes are shown with their explained variance inside the parentheses above each panel. Contours (positive solid and negative dashed) 851 852 in the EOF1 panels (left column, top two rows) are the regression coefficients (hPa/K) of 853 monthly SLP anomaly against the Niño3.4 index, a measure of the ENSO teleconnection. 854



Fig. 5 Top row: (a) the first EOF mode in CRU observations and the second EOF mode in (b)
FLOR and (c) AM2.5. The explained variance percentage is denoted above each panel. Pattern
correlation is 0.772 between CRU and FLOR and 0.993 between FLOR and AM2.5. Bottom
row: (d) the associated PC autocorrelation and (e) spectrum.

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Fig. 6 The same as Fig. 5 but for the second EOF mode in observations and the first EOF mode in FLOR and AM2.5. Pattern correlation is 0.869 between CRU and FLOR and 0.996 between 865 FLOR and AM2.5.

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Fig. 7 Large-scale pattern of correlation coefficients between the PCs of the EOF modes in Fig. 5
(reflecting precipitation variability over the southeastern United States) and surface temperature
(shading in top row), SLP (contours in top row) and 200mb geopotential height (shading in
bottom row) in observations (left column, 20CR dataset), FLOR (middle column) and AM2.5
(right column). Contours in the bottom row are the departure of climatological GHT200 from its

zonal mean, denoting climatological stationary eddies. Contour interval is 0.1 for SLP

- 875 correlation and 20m for GHT200, and solid (dashed) contours denote positive (negative) values
- 876 (zeros omitted). The magenta box in each panel denotes the region where the EOF
- 877 decomposition is conducted.
- 878 879



- Fig. 8 The same as Fig. 7 but for the EOF modes shown in Fig. 6 (reflecting precipitation variability over the Pacific Northwest).
- 883



Fig. 9 Fractional change (%) in year-to-year standard deviation of monthly precipitation between
FLOR and AM2.5 as a function of calendar month, defined as (FLOR-AM2.5)/FLOR\*100%.
Gray stippling indicates changes are not significant at the 1% level based on a two-sided Fisher
test. The blue transposed 'L' outlines the SWNA region where the month-to-month variance of
monthly precipitation (Fig. 12a) is significantly enhanced by the ocean in FLOR relative to
AM2.5.







vertically integrated flux of standard deviation of specific humidity by mean circulation (kg/m/s).
A scale vector of 20kg/m/s is shown in the upper right corner of each panel.



Fig. 12 Fractional change (%) in monthly precipitation (top row) and CWV (bottom row) month-to-month standard deviation in FLOR (a-b), Globe\_1-12 (c-d), Tropic\_1-12 (e-f), Pacific\_10-5 (g-h) and Atlantic\_5-10 (i-j) with respect to AM2.5. Gray stippling indicates changes are not significant at the 1% level based on a two-sided Fisher test. The blue 'L' outlines the SWNA region.





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Fig. 15 The same as Fig. 13 but for Atlantic\_5-10.



Fig. 16 The same as Fig. 15 but for CWV in Atlantic\_5-10.

# 923 Appendix





Fig. 17 Biases of monthly precipitation climatology (mm/day) in FLOR compared to the 19802010 CRU and GPCC observations. Shown here is FLOR – CRU. Stippling is measure of
insignificance, indicating either CRU or GPCC climatology is inside the range of a synthetic 33member FLOR ensemble, which is constructed by sampling the 1000-year FLOR simulation
with a 30-year non-overlapping period.

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Fig. 18 Biases in year-to-year standard deviation of monthly precipitation (mm/day) over North
America in FLOR compared to the 1980-2010 CRU and GPCC observations. Stippling is a



- of a synthetic 33-member FLOR ensemble, which is constructed by sampling the 1000-year
- 938 FLOR simulation with a 30-year non-overlapping period.
- 939



- **Fig. 19** Correlation of monthly surface temperature with monthly precipitation averaged over
- southwestern North America (19-40°N, 125-96°W, indicated by a red box in each panel) as a
- 943 function of calendar months in FLOR. Gray stippling denotes that the correlation is not
- 944 significant at 5% level (based on a two-sided student t test). Note that this figure has been
- 945 published in Zhang (2020, Fig. 13).
- 946



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**Fig. 20** Fractional change (%) in year-to-year standard deviation of monthly precipitation in

Globe\_1-12 relative to AM2.5, (Globe\_1-12 - AM2.5)/Globe\_1-12\*100%, as a function of
calendar month. Gray stippling indicates changes are not significant at the 1% level based on a
two-sided Fisher test.

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Fig. 21 The same as Fig. 20 but for CWV in Globe\_1-12.





Fig. 23 The same as Fig. 22 but for CWV in Tropic\_1-12.