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Mechanism of Future Spring Drying in the Southwest U.S. in CMIP5 Models*

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40 ABSTRACT

41

42 The net surface water budget, precipitation minus evaporation (P-E), shows a clear
43 seasonal cycle in the American Southwest with net gain of surface water (positive P-E) in
44 the cold half of the year (October to March) and net loss of water (negative P-E) in the
45 warm half (April – September), with June and July being the driest months of the year.

46 There is a significant shift of the summer drying toward earlier in the year under a CO₂
47 warming scenario, resulting in substantial spring drying (MAM) of the American
48 Southwest, from the near-term future to the end of the current Century with gradually
49 increasing magnitude. While the spring drying has been identified in previous studies, its
50 mechanism has not been fully addressed. Using moisture budget analysis, we found that
51 the drying is mainly due to decreased mean moisture convergence, partially compensated
52 by the increase in transient eddy moisture flux convergence. The decreased mean moisture
53 convergence is further separated into components due to changes in circulation (dynamic
54 changes) and changes in atmospheric moisture content (thermodynamic changes). The
55 drying is found to be dominated by the thermodynamic driven changes in column averaged
56 moisture convergence, due mainly to increased dry zonal advection caused by the
57 climatological land-ocean thermal contrast, rather than by the well-known “dry get drier”
58 mechanism. Furthermore, the enhanced dry advection in the warming climate is dominated
59 by the robust zonal mean atmospheric warming, leading to equally robust spring drying in
60 Southwest US.

61 **1. Introduction**

62 There is some agreement in previous studies that the Southwest United States
63 (SWUS), a region stretching from the Southern Plains to the Pacific coast between 25 and
64 45N latitudes, will likely become drier in the greenhouse warming future (e.g., Seager et
65 al., 2007, 2013; Seager and Vecchi, 2010; Scheff and Frierson, 2012). While these model-
66 based projections echo the recent severe droughts in the Southwest, there is uncertainty as
67 to the relative roles of radiative forcing and natural variability in driving recent
68 precipitation history, although the latter appears dominant (e.g., Seager et al. 2015,
69 Delworth et al. 2015, Prein et al., 2016). By comparison, there is widespread confidence
70 that warming of the southwest, which creates a tendency to reduce soil moisture and
71 streamflow, is ongoing and driven by climate change (e.g. Williams et al. 2015; Cook et
72 al., 2014; 2015; Diffenbaugh et al., 2015). Given the growing demands for water in the
73 region due to increasing population and economic growth, water resource management is
74 expected to become increasingly challenging if recent trends continue and/or model
75 projections are correct.

76 The future change in surface water availability is season dependent, as most of these
77 areas have a net gain of surface water (precipitation minus evaporation, $P - E$) in the cold
78 half of the year (October to March), and a net loss of water in the warm half (April –
79 September) (Seager et al., 2014). Any seasonal shift of this pattern will add to the
80 complexity of the water resource challenges. In addition, increasing surface temperature
81 due to greenhouse warming will likely reduce snow pack and cause early melting, thus
82 reducing the natural storage of surface water for summer usage (e.g., Mote, 2006; Pierce
83 et al., 2008; Luce et al., 2014).

84 Seager et al. (2014) provided a detailed account of present day and near-term future
85 changes in the hydrological cycle over North America using the moisture budget approach
86 by separating into the warm and cold seasons using the European Centre for Medium
87 Range Weather Forecasts ERA-Interim Reanalysis (ERA-I, Dee et al., 2011) and CMIP5
88 models' historical and future scenario (Representative Path Way 8.5, RCP85) simulations
89 (Taylor et al., 2012). They found that during the winter half year, the models project drying
90 of the Southwest due mainly to the reduction in mean moisture convergence. However,
91 the exact mechanisms and the full seasonal cycle of the Southwest drying trend as projected
92 in the model were not examined, nor whether this trend amplifies over time. Using a finer
93 resolution regional climate model, Gao et al. (2014) examined seasonal changes of P – E
94 for the end of the 21st Century as compared to the present climate and found a robust spring
95 drying in the southwestern U.S.. However, the physical mechanisms for this pronounced
96 spring drying were also not clearly identified.

97 Unlike over the oceans, where changes in P – E are dominated by the so-called wet-
98 get-wetter and dry-get-drier mechanism (e.g. Held and Soden, 2006) as a consequence of
99 increasing atmospheric water vapor content in a warming climate, the continental
100 hydroclimate change is more complex. For example, Boos (2012) and Byrne & O'Gorman
101 (2015) found that changes in zonal temperature gradient, thus the associated atmospheric
102 water vapor gradient, can be an important factor in P - E changes in the last glacial
103 maximum and future warming climate, respectively. These studies, however, do not
104 address specifically the SWUS region, or factors influencing the seasonal cycle of the P –
105 E changes.

106 These previous studies led us to examine the multimodel CMIP5 future projections
107 of the surface water balance and their seasonal change over the southwest in this study,
108 focusing on the mechanisms of the changes and the development over time from the near
109 future (2021-2040) to the end of the 21st Century. The rest of the paper is organized as
110 follows. Section 2 presents the data and methods used in this study, followed by a
111 discussion of the mechanisms for the climatological seasonal cycle of moisture budget in
112 the Southwest U.S. in section 3. Section 4 provides the detailed mechanisms of the change
113 in seasonal moisture budget and the spring drying, followed by a summary in section 5.

114

115 **2. Data and Methods**

116 We used the same 22 CMIP5 models (table 1) as in Seager et al. (2014) that have
117 the available 6-hourly data for calculating transient eddy moisture fluxes necessary for the
118 moisture budget analysis. These 22 models provide historical simulations with both
119 anthropogenic and natural radiative forcings for the historical period and future projections
120 with RCP8.5. In this study, we focus on the period 1979-2005 as the present-day base
121 period, and the future changes (from 2021 to 2100) in hydroclimate and moisture budget
122 are with respect to that reference period. In order to validate the present day CMIP5
123 simulations, we used the ERA-I Reanalysis (Dee et al, 2011) for the same period, 1979 –
124 2005, for direct comparisons.

125 Moisture budget analyses were performed for both the ERA-I and CMIP5 present
126 and future simulations as in Seager and Henderson (2013). Briefly, the column-integrated
127 moisture budget for a steady state atmosphere can be expressed in pressure coordinates as
128 follows:

129
$$P - E = -\frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \vec{u} q dp \quad (1)$$

130 where P represents precipitation, E evaporation/evapotranspiration, g is the gravitational
 131 constant, ρ_w is water density, p is pressure and p_s its surface value, q is specific humidity
 132 and \vec{u} the horizontal wind vector. When averaging over a month, the column-integrated
 133 total moisture convergence (right side of Eq. 1) can be expressed as the sum of the monthly
 134 mean moisture convergence plus the sub-monthly transient eddy moisture convergence, as
 135 follows:

136
$$\bar{P} - \bar{E} = -\frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \bar{\vec{u}} \bar{q} dp - \frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \overline{\vec{u}' q'} dp \quad (2)$$

137 where bar represents monthly mean and prime daily deviation from the monthly mean. The
 138 first term on the right-hand side of Eq. (2) can be further separated into three terms, relating
 139 to mean moisture advection and mass divergence as well as a boundary term as follows:

140
$$\nabla \cdot \int_0^{p_s} \bar{\vec{u}} \bar{q} dp = \int_0^{p_s} \bar{\vec{u}} \cdot \nabla \bar{q} dp + \int_0^{p_s} \bar{q} \nabla \cdot \bar{\vec{u}} dp + \bar{q}_s \bar{\vec{u}}_s \cdot \nabla p_s \quad (3)$$

141 where \bar{q}_s and $\bar{\vec{u}}_s$ represent the surface specific humidity and vector horizontal wind,
 142 respectively. The boundary term arises from the surface pressure gradient and can be large
 143 around mountains and represents in some sense moisture convergence and divergence at
 144 the surface due to the topography.

145 When the changes of the moisture budget are needed for two selected periods, we
 146 use δ to represent that change and Eq. (2) can be rewritten as follows:

147
$$\delta(\bar{P} - \bar{E}) = \delta \left(-\frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \bar{\vec{u}} \bar{q} dp \right) + \delta \left(-\frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \overline{\vec{u}' q'} dp \right) \quad (4)$$

148 where δ represents the difference between the two periods, and the long bar represents the
 149 period average. The first term on the right side of Eq. (4) can be further separated into terms
 150 representing changes in mean moisture convergence due to only changes in horizontal wind

151 (dynamic, DYN) and that due to only changes in specific humidity (thermodynamic, TH)
 152 as follows:

$$\begin{aligned}
 153 \quad \delta \left(-\frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \bar{\bar{u}} \bar{\bar{q}} dp \right) &\cong \frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \delta \bar{\bar{u}} \bar{\bar{q}}_p dp + \frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \left(\bar{\bar{u}}_p \right) \delta \bar{\bar{q}} dp \\
 154 &= \delta \bar{\bar{DYN}} + \delta \bar{\bar{TH}} \quad (5)
 \end{aligned}$$

155 where $\delta \bar{\bar{u}} = \bar{\bar{u}}_f - \bar{\bar{u}}_p$ and $\delta \bar{\bar{q}} = \bar{\bar{q}}_f - \bar{\bar{q}}_p$, and subscript p represents past (1979-2005)
 156 monthly mean value and subscript f represent future monthly mean value. Note that the
 157 higher order nonlinear term involving the change in circulation and change in humidity is
 158 found to be negligible and not included in Eq. 5. These various decompositions will be
 159 used in the following to disentangle the role of the various physical processes in
 160 contributing to changes in future hydroclimate.

161

162 **3. Climatological Seasonal Cycle of Moisture Budget in** 163 **Southwest United States**

164 While Seager et al. (2014) investigated many aspects of the North American
 165 moisture budget and their future changes in the winter and summer half years, they did not
 166 address the detailed seasonal cycle of the moisture budget and its change, particularly with
 167 respect to the semi-arid Southwest US (SWUS) region. Changes in seasonal cycle have
 168 important implications as water managers need to adjust to the changes when planning for
 169 water allocations throughout the year. Figure 1 shows the three-month mean seasonal
 170 (DJF, MAM, JJA, and SON) net surface water balance (P-E) using the ERA-I Reanalysis
 171 and CMIP5 multimodel mean (MMM). For the SWUS (depicted by the box), during winter
 172 (DJF) there is net gain of surface water over most of the domain except the southernmost

173 region. For both spring (MAM) and summer (JJA), the SWUS is dominated by net loss of
174 surface water, with stronger drying in the summer. The exception is the North American
175 Monsoon region of surface water gain in summer in the southwest of the domain. By the
176 fall, the drying of the SWUS lessens and turns into net surface wetting in the northern
177 portion. This seasonal cycle is well reproduced by the CMIP5 MMM, except that the
178 climatological spring drying is limited to the southern half of the domain, thus indicating a
179 delay in the seasonal cycle of warm season drying. There is also a net gain of water in the
180 fall season in models across the region, indicating a bias toward generally wetter conditions
181 in the model climatology throughout the year.

182 To better illustrate how P-E changes throughout the season and to understand the
183 mechanisms of the spring and summer drying, Fig. 2 shows the SWUS area average (box
184 shown in Fig. 1) P-E along with the mean and transient moisture flux convergences (top
185 three rows) and the mean moisture advection (fourth row) and mass divergence (fifth row)
186 contributions to the total mean moisture convergence terms, along with the boundary term
187 (sixth row), for both ERA-I Reanalysis and CMIP5 MMM, as a function of month. In the
188 ERA-I Reanalysis, there is a net gain of surface water in the winter half year, from October
189 to March, and net loss of water in the summer half year, from April to September. The
190 peak drying time is in June and peak wet months are in December and January. The positive
191 P-E during the winter half year is mainly due to synoptic storms converging moisture into
192 the region, as indicated by the transient moisture convergence term (third row). The
193 transient moisture flux convergence is offset by the mean moisture divergence out of the
194 region (second row) which is negative throughout the year except in July and August when

195 it is weakly positive. The climatological drying in the warmer half year is caused by mean
196 moisture divergence in spring and transient moisture divergence in summer.

197 Furthermore, the mean moisture divergence is due to both mean mass divergence
198 and moisture advection with the latter dominant. The mean moisture advection term is
199 drying for the majority of the annual cycle and peaks in the late spring/early summer
200 months. In this region of complex topography, the boundary term (sixth row) can be a large
201 wetting factor peaking in summer that offsets the advective drying.

202 The CMIP5 MMM well represents the moisture budget terms and their seasonal
203 cycle in the SWUS region. As shown in Fig. 1, the net surface water budget tends to have
204 a wet bias in the region, causing a wetter winter and slightly less dry summer compared to
205 ERA-I. The wet bias is mainly due to the transient eddy moisture flux convergence being
206 too large (compare Fig. 2f to Fig. 2e). Other than these small discrepancies, the CMIP5
207 MMM reproduces well the main features of the moisture budget seasonal cycle and thus
208 can be used for understanding the future changes in SWUS hydroclimate.

209 The dominant climatological drying contribution from the mean moisture advection
210 (Figs. 2g, h) during the spring and summer is somewhat counterintuitive as one would
211 expect prevailing westerlies in the region to bring moisture from the Pacific Ocean into the
212 SWUS region to its east. Since the mean moisture advection turns out to be the dominant
213 mechanism for the future spring drying as well, it is worthwhile to first explore the physical
214 causes of its climatology. After examination, it turns out that the drying is due mainly to
215 the zonal advection term (the meridional advection is secondary and of opposite sign).
216 Thus, we focus below on the zonal moisture advection term.

217 Figure 3 shows the pressure-longitude vertical cross sections of specific humidity,
218 air temperature, and zonal wind vectors averaged over the latitude span of 32-45N for the
219 four seasons using ERA-I (left) and CMIP5 MMM (right). Since the zonal mean
220 components of q and T do not contribute to the zonal advection, we only show their zonally
221 asymmetric parts in Fig. 3. The specific humidity shows a relatively small east-west
222 gradient during winter but a very strong zonal dipole structure in the summer with smaller
223 q over the coastal region and larger q on top of the mountains and east of the Rockies. Both
224 spring and autumn seasons show similar specific humidity structure as the summer but with
225 smaller peaks over the highlands. The air temperature is influenced by the local topography
226 and land sea contrasts with cooler temperature over the oceans and warmer temperature
227 over land, particularly above the mountains in the summer. Part of the specific humidity
228 zonal dipole can be explained by the zonally asymmetric temperature structure according
229 to the Clausius-Clapeyron equation with uniform relative humidity at each level (not
230 shown), and is thus driven by land-sea thermal contrasts and local topography. However,
231 the zonally asymmetric q and T do not coincide with each other completely, suggesting
232 that there are dynamical processes involved in shaping the q structure.

233 To further understand the climatological zonally asymmetric q structure in the
234 region, we show in Fig. 4 the vertically integrated mean moisture transport for all four
235 seasons based on both ERA-I and CMIP5 MMM, along with the 850 hPa specific humidity.
236 In the winter, the moisture transport along the west coast is dominated by westerlies
237 bringing relatively warm and humid air to the region in both reanalysis and CMIP5 MMM
238 (top row). But from spring to fall, the mean moisture transport is dominated by the along-
239 coast cool and dry advection from the north associated with the Pacific subtropical

240 anticyclone while, further inland, it is dominated by the warm moist air from the Gulf of
241 Mexico associated with the Great Plains Low Level Jet (LLJ, Ting and Wang, 2004, Jiang
242 et al. 2007, Parish and Oolman 2010). These processes create a moisture gradient in the
243 region that is dry over the coastal regions and moist further inland. Any zonal advection
244 of moisture in the region would lead to advective drying from March to October (Figs. 2g,
245 h).

246 The CMIP5 MMM shows very similar features of the zonally asymmetric q and T
247 (right panels of Fig. 3), as well as the moisture transports (right panels of Fig. 4). with the
248 strongest drying due to mean moisture advection occurring in June (Fig. 2h), slightly
249 shifted compared to reanalysis observations. The results here suggest that the mean flow
250 moisture divergence in the SWUS, which dominates the climatological warm season
251 drying in the region, is mainly driven by the zonally asymmetric specific humidity
252 gradients. The specific humidity gradients are a result of land-ocean thermal contrasts,
253 local topography, as well as moisture transport associated with the Pacific subtropical
254 anticyclone and the Great Plains low level jet. The next section examines how the zonal
255 specific humidity gradient and the SWUS drying evolve in the future.

256

257 **4. Changes in Seasonal Cycle of Moisture Budget and the** 258 **Mechanisms of Spring Drying**

259 The future changes in SWUS hydroclimate are explored by examining the four 20-
260 year future periods, starting from 2021-2040 to 2081 -2100. Figure 5a illustrates changes
261 in net surface water balance, $P - E$, from each of the 20-year periods with respect to the
262 recent period, 1979 – 2005. These maps show the general drying trend in the SWUS region

263 throughout the seasonal cycle except January and February when the changes are slightly
264 positive for all future periods. More notable is that the spring season, MAM, consistently
265 shows the strongest drying signal, effectively shifting forward the peak drying season of
266 negative P-E from mid-summer (top row in Fig. 2) towards late spring/early summer (see
267 Fig. 10). The amplitude of the drying also increases steadily from the near-term future to
268 the end of the 21st Century. When separating the future drying into the mean and transient
269 contributions in Fig. 5, it is clear that spring drying is predominantly caused by the mean
270 moisture divergence (Fig. 5b), whereas in the summer, drying by transient eddy moisture
271 divergence (Fig. 5c) is largely cancelled by mean flow wetting leading to little change in
272 P-E. Given the large amplitude of the spring drying, we focus the rest of the paper on the
273 mechanisms responsible.

274 Figure 6 shows the spatial patterns of the spring drying for the four periods in terms
275 of P-E. The spatial pattern of the drying is robustly similar across the four periods with
276 increasing amplitude toward the future and it is particularly strong in the northern part of
277 the domain, from the California coast to Colorado. In the southern tip of the domain, there
278 is actually a slight wetting trend. To gain further insights into the spring drying
279 mechanisms, we show in Fig. 7 the area-averaged moisture budget changes for the four
280 future periods with respect to the recent period, for P-E, total mean moisture convergence
281 and transient moisture flux convergence (Fig. 7a-c). Consistent with Figs. 5 and 6, there
282 is a dominant spring drying in terms of P-E, and this drying is entirely due to the mean
283 moisture divergence, offset somewhat by the transient eddy moisture convergence and
284 wetting. The changes in mean and transient moisture convergence amplify the
285 corresponding climatological processes as shown in Fig. 2. The mean moisture

286 convergence change is further divided into that due to circulation change (dynamic, DYN)
287 and that due to specific humidity change (thermodynamic, TH) as shown in Eq. 5 (Fig.
288 7d,e). The dynamic term contributes negligibly to spring drying (Fig. 7d) and it is instead
289 almost entirely caused by the thermodynamic contribution due to increases in specific
290 humidity (Fig. 7e). The dominance of the thermodynamic term here may not be surprising.
291 It might be thought that since this is a region of mean mass divergence, a warming-driven
292 increase of moisture in the atmosphere would lead to more moisture divergence and hence
293 drying. However, Figs. 7f and g illustrate that the thermodynamic change is almost entirely
294 due to the climatological wind advecting the anomalous specific humidity gradient, while
295 the climatological mean mass divergence of anomalous moisture is negligible. The
296 dominance of the advection term seems to be consistent with the climatological moisture
297 budget shown in Fig. 2. We next examine further how the moisture gradient changes in the
298 future as the climate warms.

299 Figure 8a shows the vertical cross section of the spring zonally asymmetric specific
300 humidity change between the end of the 21st Century and the current climate from CMIP5
301 MMM. There is an enhanced specific humidity gradient with reduced specific humidity to
302 the west and enhanced humidity to the east of the domain. This causes anomalous dry
303 advection by the climatological westerlies (Figs. 3c and 7f). To understand the causes of
304 the change in specific humidity gradient we separate the humidity change using the
305 Clausius-Clapeyron equation.

306 The specific humidity can be written approximately as: $q = r q_s$, where r is relative
307 humidity (defined as the ratio of actual vapor pressure e and saturation vapor pressure e_s , r

308 = e/e_s) and q_s is the saturation specific humidity, which is only a function of temperature
 309 according to the Clausius-Clapeyron equation:

$$310 \quad e_s = e_0 \exp \left[\frac{L}{R_v} \left(\frac{1}{T_0} - \frac{1}{T} \right) \right], \quad q_s \approx \frac{R_d}{R_v} \left(\frac{e_s}{p - e_s} \right) \quad (6)$$

311 where e_0 represents the e_s value when T is equal to a reference temperature T_0 , L is the
 312 latent heat of vaporization and R_d and R_v are the gas constants for dry air and water vapor,
 313 respectively, and p is the air pressure. Define the specific humidity change as

$$314 \quad \Delta q = q_f - q_p = r_f q_s(T_f) - r_p q_s(T_p) = \Delta r q_s(T_p) + r_p \Delta q_s + \Delta r \Delta q_s$$

$$315 \quad (7)$$

316 where subscripts f and p represents future and past values and $\Delta = ()_f - ()_p$. If we
 317 ignore the nonlinear term in (7), then

$$318 \quad \Delta q \approx \Delta r q_s(T_p) + r_p \Delta q_s \quad (8)$$

319 where Δq_s can be written as:

$$320 \quad \Delta q_s = q_s(T_p + \Delta T) - q_s(T_p)$$

321 Figure 8b shows the calculated Δq according to Eq. (8) with the zonal mean part removed
 322 (Δq^*), which agrees well with the Δq^* based on model outputs in Fig. 8a. If we assume
 323 relative humidity does not change in the future, an assumption, which has been shown to
 324 be a good approximation in both observations (Gaffen and Ross, 1999) and theoretically
 325 (Pierrehumbert et al., 2007), then equation 8 can be approximated by

$$326 \quad \Delta q \approx r_p \Delta q_s = r_p [q_s(T_p + \Delta T) - q_s(T_p)] \quad (9)$$

327 The resulting change in the zonally asymmetric specific humidity is shown in Fig. 8c,
 328 which reproduces well the actual model change but with somewhat larger amplitude. Thus,
 329 the change in air temperature (ΔT) with fixed relative humidity dominates the change in q .

330 The air temperature change in Eq. (9) can be further divided into zonal mean change and
 331 zonally asymmetric change in air temperature as follows:

$$332 \quad \Delta q \approx r_p [q_s(T_p + \langle \Delta T \rangle + \Delta T^*) - q_s(T_p)] \quad (10)$$

333 where angle bracket represents the zonal mean value, and asterisk the zonally asymmetric
 334 component. It is clear from Fig. 8d that the zonally asymmetric q change above 700 hPa
 335 is largely explained by the zonal mean temperature change ($\langle \Delta T \rangle$). Figures 8e and f show
 336 the changes in zonally asymmetric q due to ΔT^* only (by setting $\langle \Delta T \rangle$ to zero in Eq. 10)
 337 and relative humidity only (setting Δq_s to zero in Eq. 8), respectively. The contribution to
 338 the zonally asymmetric q change is relatively minor in both cases compared to that due to
 339 the zonal mean temperature change (Fig. 8c). Zonally uniform temperature change ($\langle \Delta T \rangle$)
 340 leads to zonally asymmetric specific humidity change (Δq^*) because land is warmer than
 341 ocean in the spring and, hence, when adding a uniform temperature increase to both land
 342 and ocean, specific humidity increases more over land than ocean due to the nonlinear
 343 Clausius-Clapeyron relation (Eq. 6). It is, however, very interesting that the specific
 344 humidity change is dominated by the zonal mean temperature change, rather than the
 345 asymmetric warming of the land and ocean in the future, or changes in relative humidity
 346 (Byrne and O’Gorman, 2015).

347 To confirm the change in zonal mean temperature, which led to the enhanced q
 348 gradient, is indeed the dominant cause of the spring drying, we computed the corresponding
 349 change in vertically integrated mean moisture convergence ($\delta \left(-\frac{1}{g\rho_w} \nabla \cdot \int_0^{p_s} \bar{\bar{u}} \bar{q} dp \right)$), due
 350 to each q change as shown in Fig. 8. The results are shown in Fig. 9 for the spring season.
 351 Consistent with Fig. 8, Figs. 9a and b are almost identical, indicating that the calculated
 352 specific humidity using the Clausius-Clapeyron equation reproduces well the CMIP5

353 MMM q. Both Figs. 9 a and b show drying in the SWUS and wetting in the east half of
354 the country, and bear some similarities to the MMM P – E pattern in Fig. 5d. This pattern
355 is largely reproduced when assuming constant relative humidity (c) and when only
356 allowing the zonally symmetric temperature to change (d). In contrast, the contributions
357 to this pattern due to change in the zonally asymmetric temperature (e) and only allowing
358 relative humidity to change (f) are relatively small.

359 The enhanced q gradient is also seen in summer and fall (not shown). However,
360 the climatological wind speed is weaker in those seasons than in spring and the enhanced
361 dry zonal advection is also less explaining the maximum drying of the region in spring.

362 Zonal mean temperature changes under greenhouse warming is a relatively robust
363 feature of the CMIP5 models, thus spring drying in SWUS is also very robust, as can be
364 seen in Figs. 6 and 7. We find it interesting that the robust spring drying under global
365 warming can be explained largely by thermodynamic processes through the zonal mean
366 temperature changes, meaning that the change in atmospheric circulation plays little role
367 in causing the drying. The dominance of thermodynamic processes may not be surprising
368 but this advective mechanism is distinct from the well-known “dry get drier” mechanism.
369 The “dry get drier” mechanism best applies over the oceans to regions of climatological
370 mass and moisture divergence and negative P-E and largely explains the large scale drying
371 over subtropical oceans (Held and Soden, 2006). Over land, there is mean moisture
372 convergence, P-E is positive and a simple application of Held and Soden arguments implies
373 wetting. However, drying over land can still occur due to thermodynamic processes and,
374 in the case of the SWUS, it is enhanced advective drying that is the prime mechanism.

375

376 **5. Summary**

377 We explored the detailed mechanisms that caused the robust spring drying over
378 SWUS under greenhouse warming as projected by the CMIP5 multimodel mean. While
379 the conventional wisdom may be that the SWUS is located in a region of mean mass
380 divergence, thus the increase of moisture in the atmosphere as a result of warming would
381 lead to more moisture divergence, an application over land of the so-called “dry get drier”
382 mechanism (Held and Soden, 2006), we find that is not the dominant mechanism in this
383 case. In fact, even in the climatological sense, the mean mass divergence is not the
384 dominant mechanism for the region being semi-arid in the first place. The spring and
385 summer SWUS drying, on the other hand, is dominated by the zonal mean advection of
386 drier air into the region due to the strong east-west humidity gradient. Intuitively, one
387 would expect the westerlies to advect moist ocean air into the drier land region, thus
388 causing wetting of the region. However, due to the land-ocean thermal contrasts and the
389 topography of the region, land is warmer than ocean during the spring, summer and fall
390 seasons allowing a maximum in specific humidity in the highland surface region and a
391 specific humidity gradient with increasing moisture inland. In the greenhouse future, when
392 a zonally uniform warming is added to the existing land-ocean thermal contrasts, the
393 anomalous specific humidity gradient intensifies due to the nonlinearity of the Clausius-
394 Clapeyron relationship. With the stronger climatological westerlies in the spring compared
395 to summer and fall, the anomalous mean moisture advection due to the climatological flow
396 advecting the anomalous specific humidity gradient reaches a maximum in the spring,
397 causing robust spring drying in SWUS. The effect increases linearly from the near future

398 (2021-2040) to the end of the 21st Century, and shows extreme robustness across the
399 CMIP5 models.

400 The mechanism here seems to be consistent with Byrne and O’Gorman (2015) in
401 that the horizontal gradients of changes in temperature and relative humidity need to be
402 taken into account to explain the P-E response to warming over land. However, we found
403 that it is not the changes in temperature gradient, but rather the nonlinear response of the
404 specific humidity gradient to the zonal mean warming superimposed on the zonally
405 asymmetric land-ocean thermal contrasts, that dominates the spring drying in the
406 Southwest United States. The contributions from both the change in zonally asymmetric
407 temperature and change in relative humidity are relatively small.

408 There are important implications of the spring drying in the SWUS. Currently the
409 peak drying season is in the summer months, while winter and early spring provide much
410 needed supply of water and water storage to the region. Since CMIP5 tends to have a wet
411 bias in P-E (Fig. 2), we can crudely correct for this by subtracting from the model future
412 P-E for each month the constant annual mean bias value of 0.54 mm/day. The resulting
413 seasonal cycle of P – E for each of the future period is shown in Fig. 10. When the seasonal
414 cycle shifts toward a drier spring, there is much reduced positive P-E in March and
415 substantially increased negative P-E in April, May and June (Fig. 10). The total reduction
416 of surface water of the entire season from March to June will prolong and intensify the dry
417 season. This will adversely impact the spring growing season, potentially increase fire risk,
418 degrade pasturelands, rangelands, and crops and lower spring and summer streamflow.
419 Although spring drying is dominant and has been the focus here, there is also substantial
420 drying in the fall season, as can be seen in Figs. 6 and 10. Coupled with early melt of snow

421 cover due to warming, a shortened winter wet season could substantially reduce the SWUS
422 water supply and storage in the future.

423

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428

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496

497

498 **List of Figures**

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500 CMIP5 MMM (right) averaged for the period 1979-2005 for DJF (a, b), MAM (c, d), JJA
501 (e, f), and SON (g, h). Units are in mm/day and contour interval is 1 mm/day.

502 FIG. 2. The climatological mean (1979-2005) seasonal cycle of the moisture budget terms
503 averaged over the Southwest United States (the outlined region in Fig. 1) for ERA-Interim
504 Reanalysis (left) and CMIP5 MMM (right). (a) and (b): precipitation minus evaporation,
505 (c) and (d): column averaged mean flow moisture convergence (MC), (e) and (f): column
506 averaged sub-monthly transient eddy MC, (g) and (h): column averaged MC due to mean
507 moisture advection, (i) and (j): column averaged MC due to mean flow mass divergence,
508 and (k) and (l): the surface boundary term due to surface pressure gradient.

509 FIG. 3. Longitude-pressure cross sections of the climatological mean (1979 – 2005) zonally
510 asymmetric temperature (black contours), zonally asymmetric specific humidity (green
511 contours) and total zonal wind vectors averaged from 32N - 45N using ERA-Interim
512 Reanalysis (left) and CMIP5 MMM (right) for DJF (a, b), MAM (c, d), JJA (e, f), and SON
513 (g, h). Contour intervals are 0.5oC for temperature and 0.25 g/kg for specific humidity and
514 negative values are dashed.

515 FIG. 4. Vertically integrated mean moisture transport (arrows) and 850 hPa specific
516 humidity (shading) for ERA Interim (left) and CMIP5 MMM (right) for the four seasons
517 averaged for the period 1979-2005. The unit is g/kg for specific humidity and $\text{kg/m}^2 \text{ m/s}$
518 for moisture fluxes. Vector scale is shown at the lower left.

519 FIG. 5. The changes in $P - E$ (a), mean moisture convergence (b), and transient moisture
520 convergence (c) based on CMIP5 MMM's RCP8.5 future scenario simulations for the

521 period 2021-2040, 2041-2060, 2061-2080, and 2081-2099 with respect to the historical
522 simulation averaged from 1979-2005, averaged over the Southwest United States land
523 region from 125W-103W and 25N to 45N.

524 FIG. 6. Changes in P – E for March, April and May seasonal average based on CMIP5
525 MMM RCP8.5 scenario simulations for 2021-2040 (a), 2041-2060 (b), 2061-2080 (c), and
526 2081-2099 (d), with respect to the 1979-2005 historical simulation.

527 FIG. 7. Changes in the various moisture budget terms for March, April and May seasonal
528 average based on CMIP5 MMM RCP8.5 scenario simulations for the four future periods
529 with respect to the 1979-2005 historical simulation for (a) P – E, (b) Mean moisture
530 convergence (MC), (c) transient MC, (d) mean MC due to changes in atmospheric
531 circulation only (DYN), (e) mean MC due to changes in specific humidity only (TH), (f)
532 the part in (e) due to climatological mean flow advecting anomalous specific humidity
533 gradient, and (g) the part in (e) due to climatological mass divergence of the anomalous
534 specific humidity.

535 FIG. 8. Longitude-vertical cross sections of the zonally asymmetric specific humidity
536 change (2075-2099 minus 1979-2005) for (a): CMIP5 MMM, (b): calculated based on
537 Clausius-Clapeyron equation and given the relative humidity and temperature changes, (c):
538 same as (b), but with fixed relative humidity and only allow the temperature to change, (d):
539 same as in (c) but only allow the zonal mean temperature to change, (e): same as in (c) but
540 only allow the zonally asymmetric temperature to change, and (f): same as in (b) but only
541 allow the relative humidity to change. Contour interval is 0.05 g/kg and negative contours
542 are dashed.

543 Figure 9. Changes in vertically integrated mean moisture convergence between averages
544 for the period (2075 - 2099) and (1979 - 2005) calculated using (a): CMIP5 MMM wind,
545 specific humidity and surface pressure, (b) same as (a) except using the specific humidity
546 calculated from Clausius-Clapeyron equation given MMM relative humidity and
547 temperature changes, (c): same as (b), but with fixed relative humidity and only allow the
548 temperature to change, (d): same as in (c) but only allow the zonal mean temperature to
549 change, (e): same as in (c) but only allow the zonally asymmetric temperature to change,
550 and (f): same as in (b) but only allow the relative humidity to change. Contour interval is
551 0.4 mm/day and negative contours are dashed.

552 FIG. 10. Bias-corrected (subtracting a 0.54 mm/day wet bias from each month to correct
553 the annual mean P-E difference between models and ERA-I Reanalysis) seasonal cycle of
554 P – E for five different periods as simulated by the CMIP5 MMM for the US Southwest
555 domain (box shown in Fig. 5). Future simulations are the RCP8.5 scenario, and past
556 simulations use CMIP5 historical forcing.

557

558 TABLE 1. CMIP5 models used in this study, including their originating institutions,
 559 horizontal and vertical resolutions, and ensemble sizes.

Institute	Model	Resolution (lon x lat), level	Ensemble size	
			20thC	rcp85
Beijing Climate Center (BCC)	1. bcc-csm1-1	T42, L26	1	1
	2. bcc-csm1-1-m	T106, L26	1	1
College of Global Change and Earth System Science, Beijing Normal University (BNU)	3. BNU-ESM	T42, L26	1	1
Canadian Centre for Climate Modeling and Analysis (CC-Cma)	4. CanESM2	T63 (1.875° x1.875°), L35	1	1
National Center for Atmospheric Research (NCAR)	5. CCSM4	288x200 (1.25° x0.9°), L26	1	1
Centro Euro-Mediterraneo per I Cambiamenti Climatici (CMCC)	6. CMCC-CM	T159, L31	1	1
Centre National de Recherches Meteorologiques / Centre Européen de Recherche et Formation Avancées en Calcul Scientifique (CNRM-CERFACS)	7. CNRM-CM5	T127(1.4° x1.4°), L31	1	1
Commonwealth Scientific and Industrial Research Organisation in collaboration with the Queensland Climate Change Centre of Excellence (CSIRO-QCCCE)	8. CSIRO-Mk3-6-0	T63(1.875° x1.875°), L18	1	1
Institute of Atmospheric Physics, Chinese Academy of Sciences and Tsinghua University (LASG-CESS)	9. FGOALS-g2	128x60, L26	2	1
Geophysical Fluid Dynamics Laboratory (NOAA GFDL)	10. GFDL-CM3	C48 (2.5° x2.0°), L48	5	1
	11. GFDL-ESM2G	144x90 (2.5° x2.0°), L24	1	1
	12. GFDL-ESM2M	144x90 (2.5° x2.0°), L24	1	1
NASA Goddard Institute for Space Studies (NASA GISS)	13. GISS-E2-H	2.5° x2°, L40	1	1
	14. GISS-E2-R	2.5° x2°, L40	1	1
Institut Pierre-Simon Laplace (IPSL)	15. IPSL-CM5A-LR	3.75° x1.875°, L39	6	3
	16. IPSL-CM5A-MR	2.5° x1.25°, L39	2	1
	17. IPSL-CM5B-LR	96x96 (3.75° x1.875°), L39	1	1
Atmosphere and Ocean Research Institute (The University of Tokyo), National Institute for Environmental Studies, and Japan Agency for Marine-Earth Science and Technology (AORI/NIES/JAMSTEC)	18. MIROC5	T85, L40	5	1
	19. MIROC-ESM	T42, L80	3	1
	20. MIROC-ESM-CHEM	T42, L80	1	1
Meteorological Research Institute (MRI)	21. MRI-CGCM3	TL159 (1.125° x1.125°), L48	1	1
Norwegian Climate Centre (NCC)	22. NorESM1-M	144x96 (2.5° x1.875°), L26	3	1

560

561

ERAInt and CMIP5 P - E, (1979-2005)

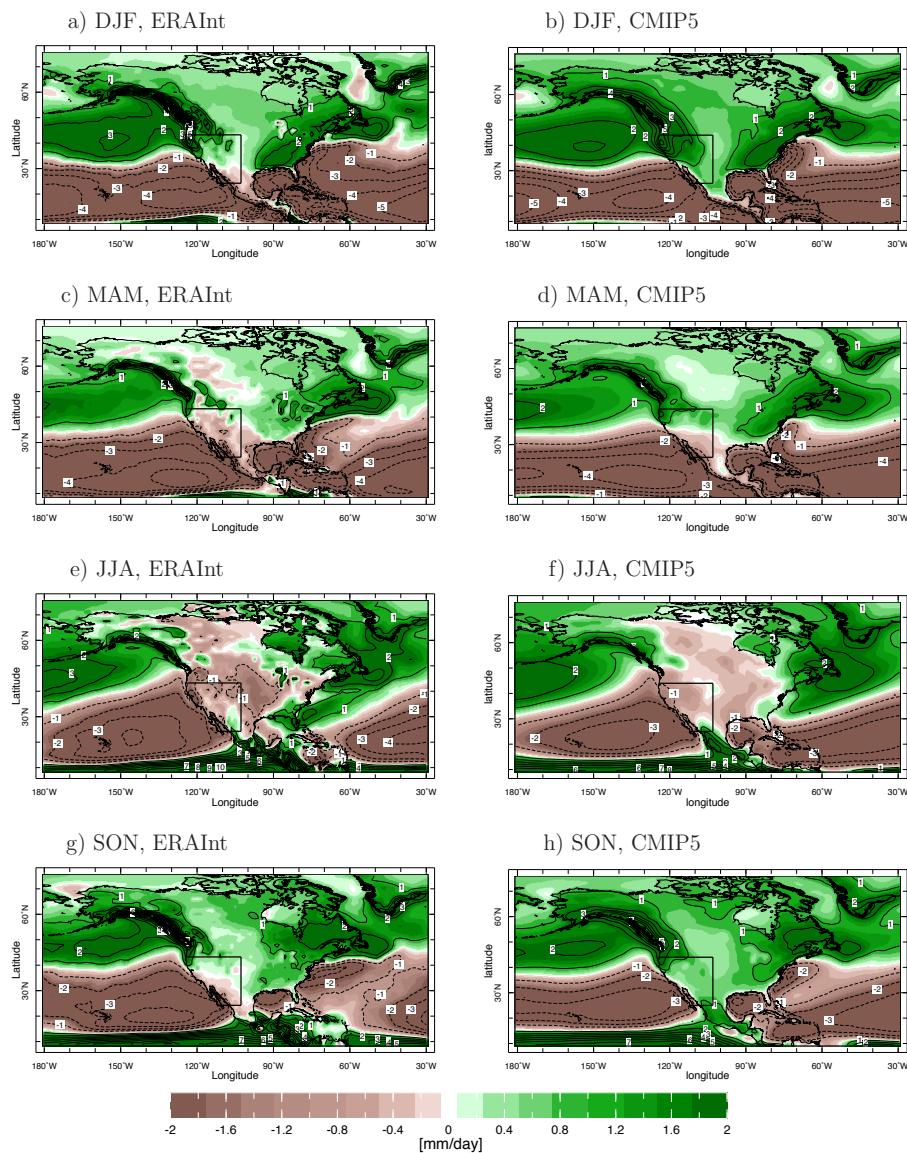


FIG. 1. Precipitation minus evaporation ($P - E$) from ERA-Interim Reanalysis (left) and CMIP5 MMM (right) averaged for the period 1979-2005 for DJF (a, b), MAM (c, d), JJA (e, f), and SON (g, h). Units are in mm/day and contour interval is 1 mm/day.

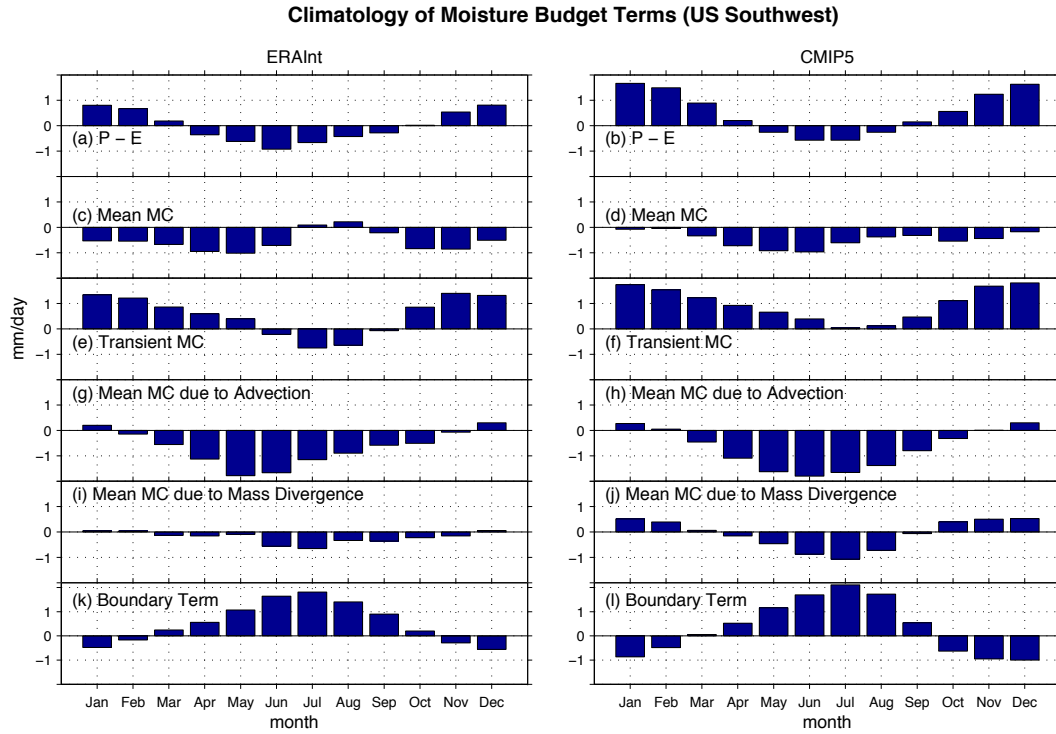


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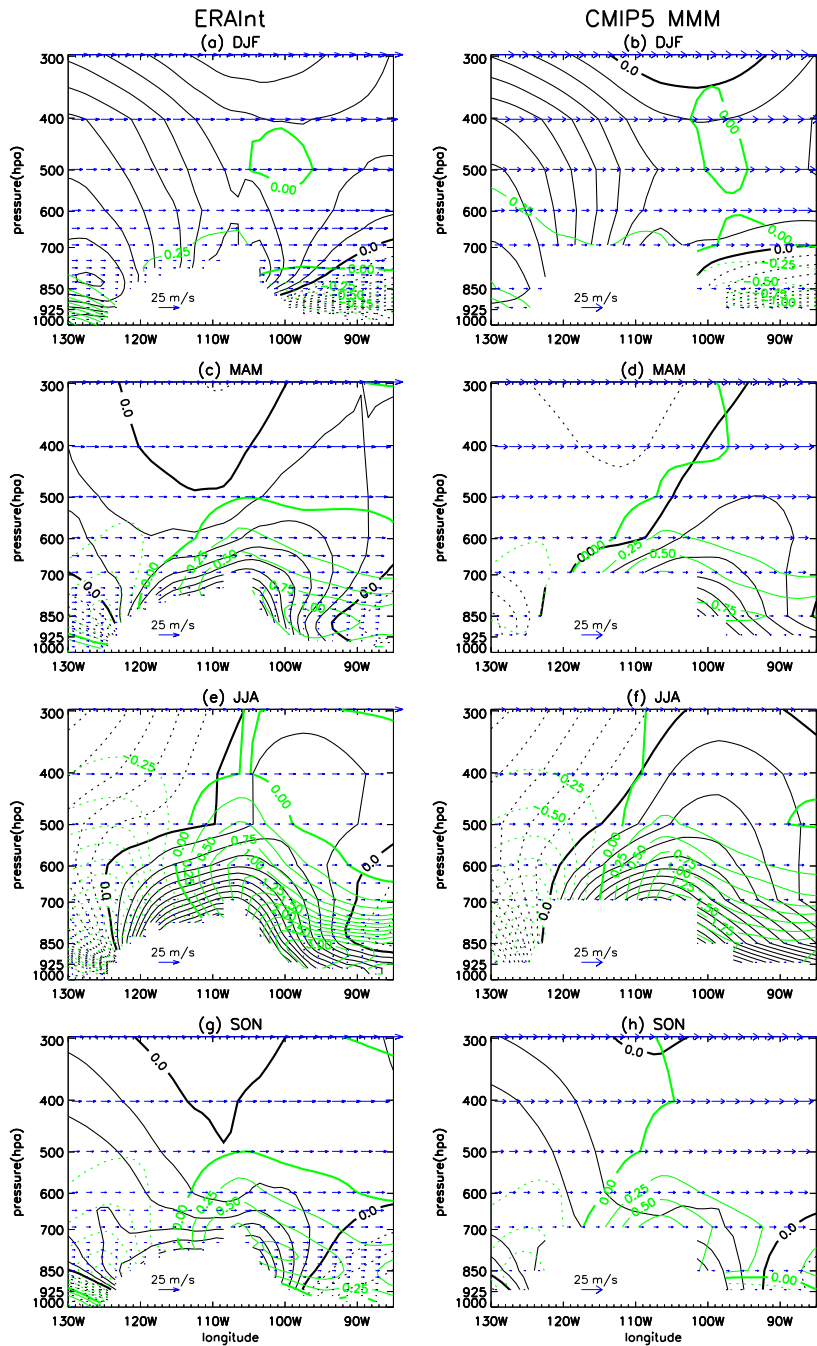


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Vertically Integrated Mean Moisture Transport and 850mb Specific Humidity (1979–2005)

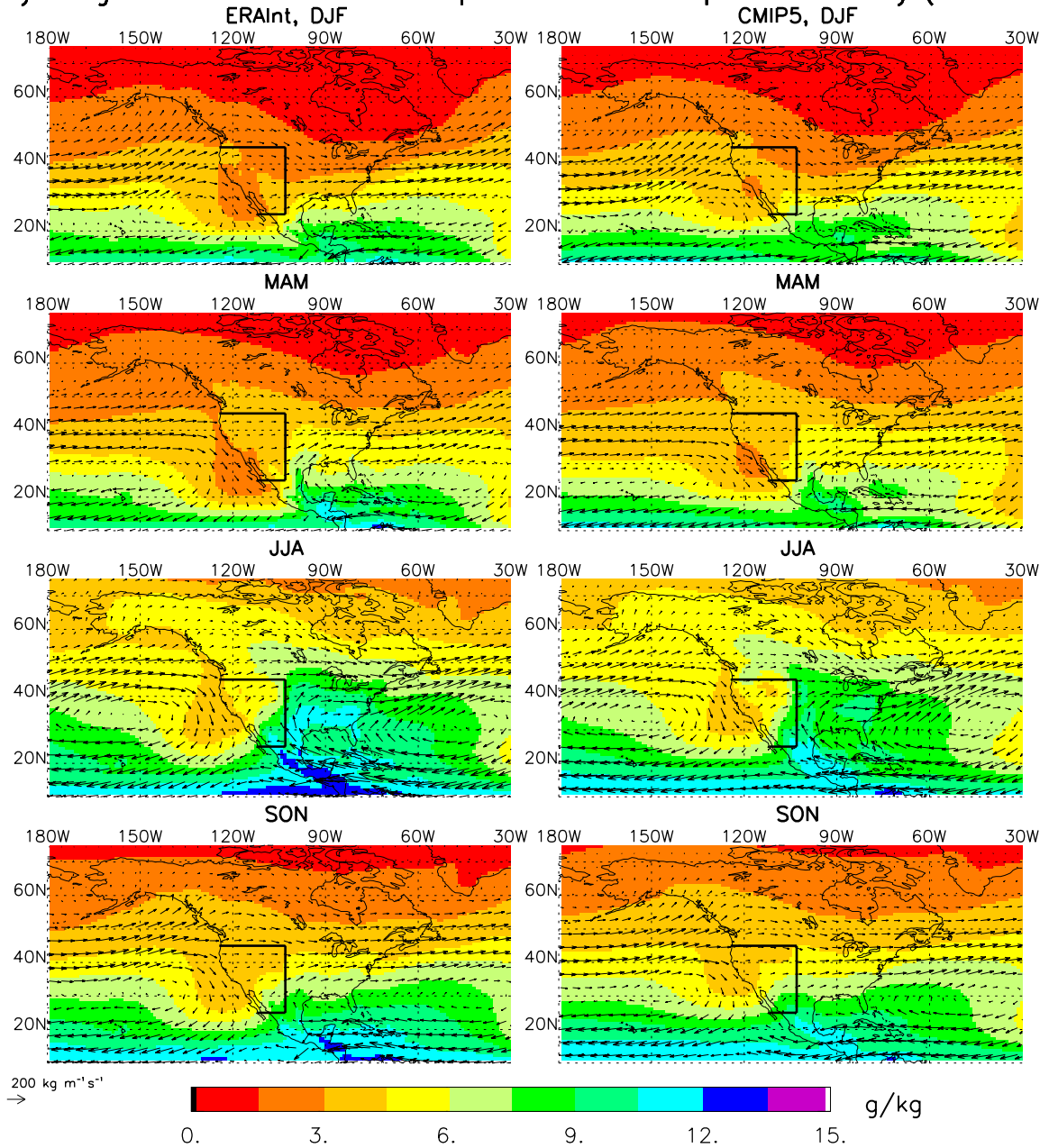


Fig. 4. Vertically integrated mean moisture transport (arrows) and 850 hPa specific humidity (shading) for ERA Interim (left) and CMIP5 MMM (right) for the four seasons averaged for the period 1979-2005. The unit is g/kg for specific humidity and $\text{kg m}^{-2} \text{m/s}$ for moisture fluxes. Vector scale is shown at the lower left.

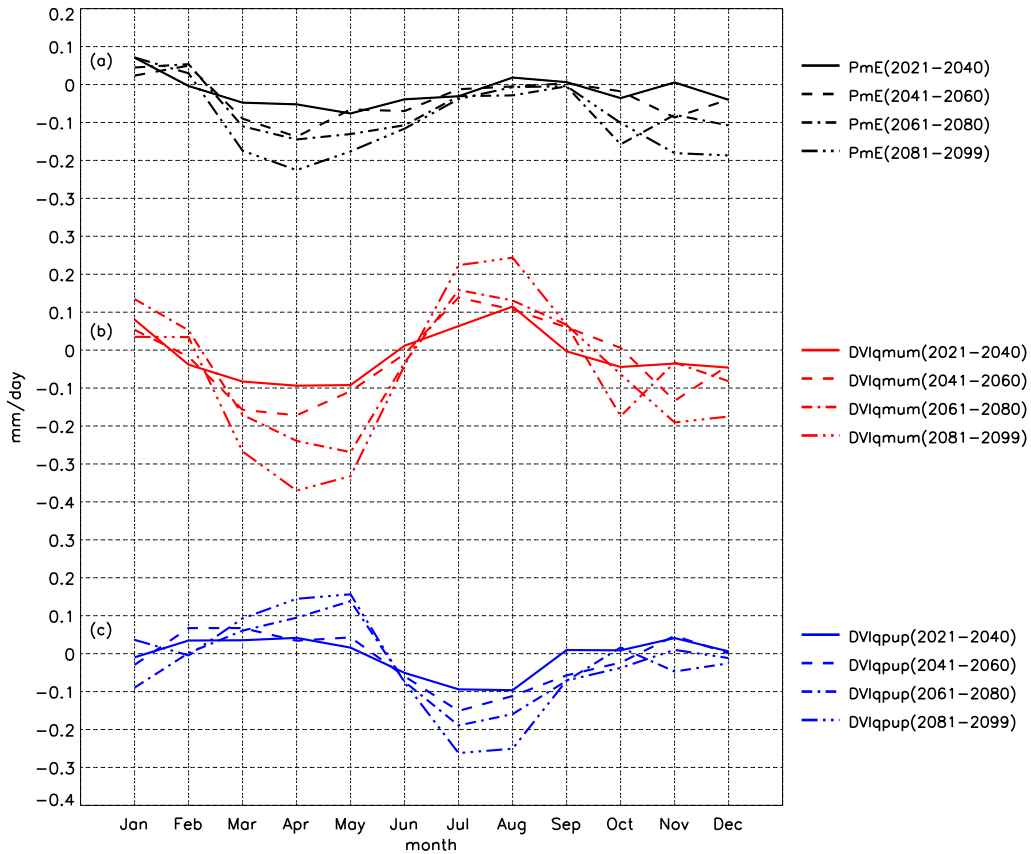


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CMIP5 P - E Change in Spring

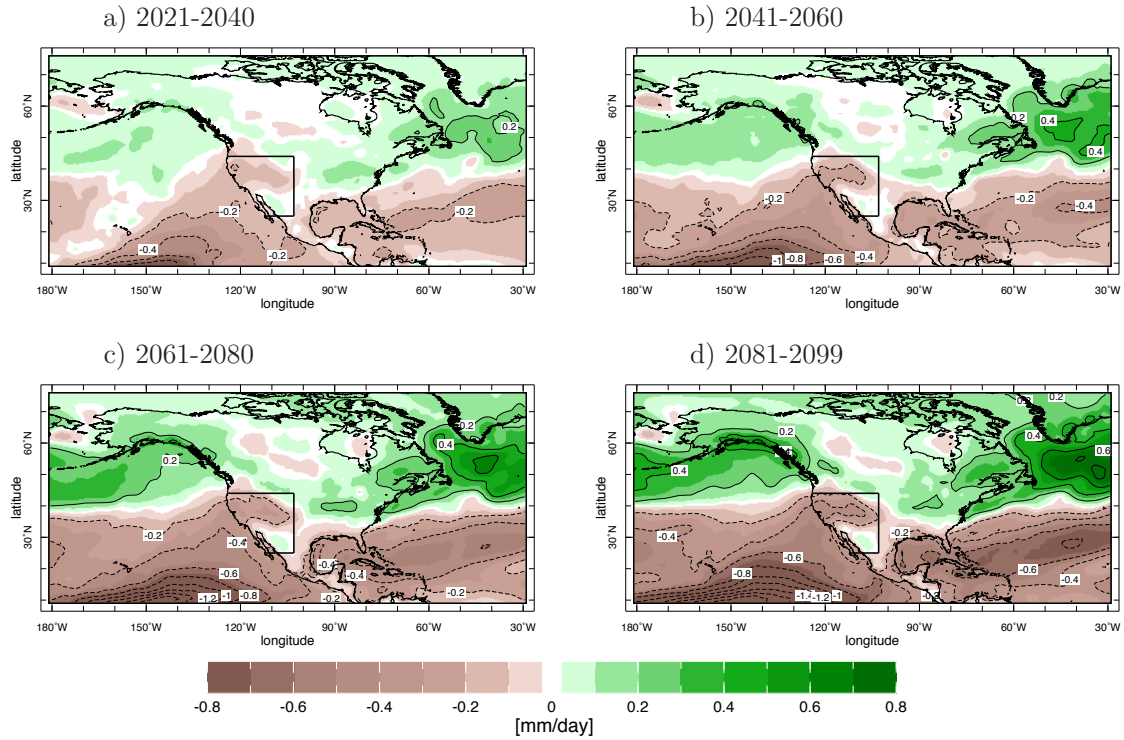


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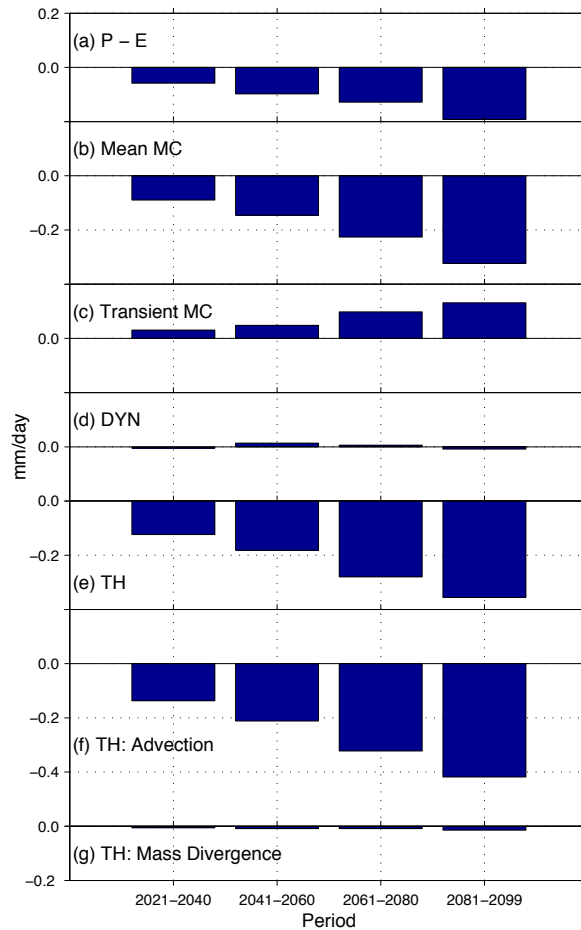


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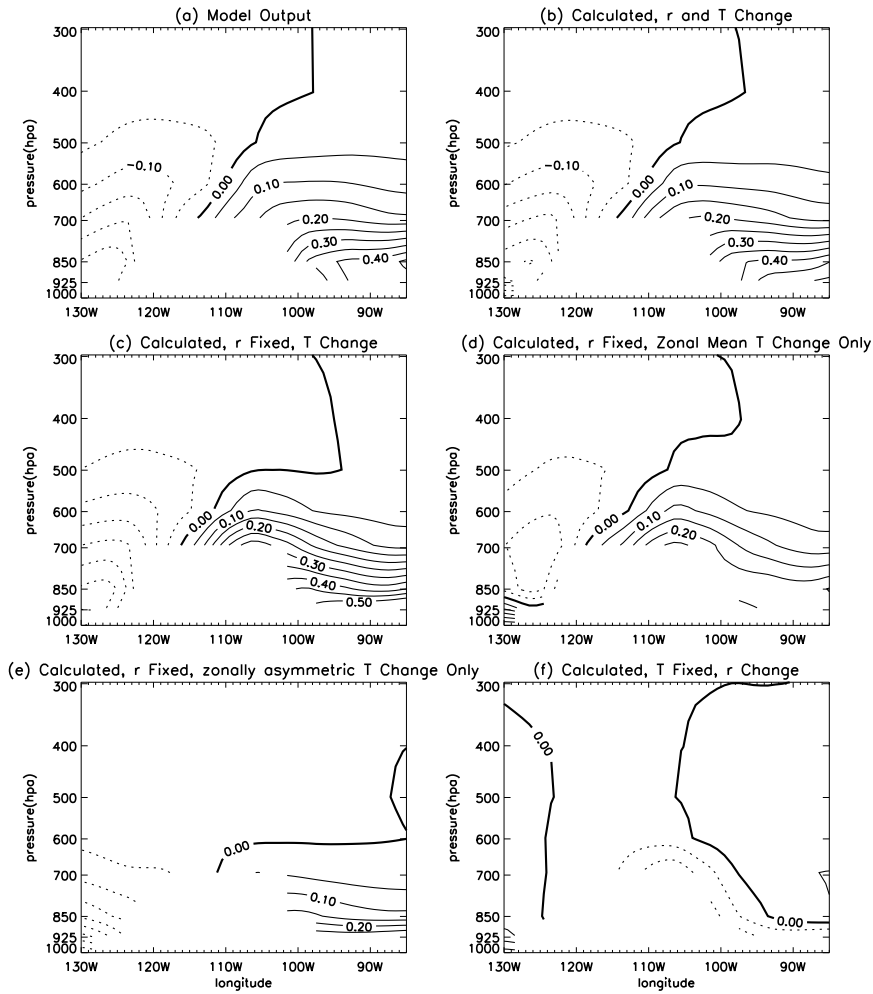


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CMIP5 Vertically Integrated Mean Moisture Convergence Change, (2075-2099) - (1975-2005), MAM

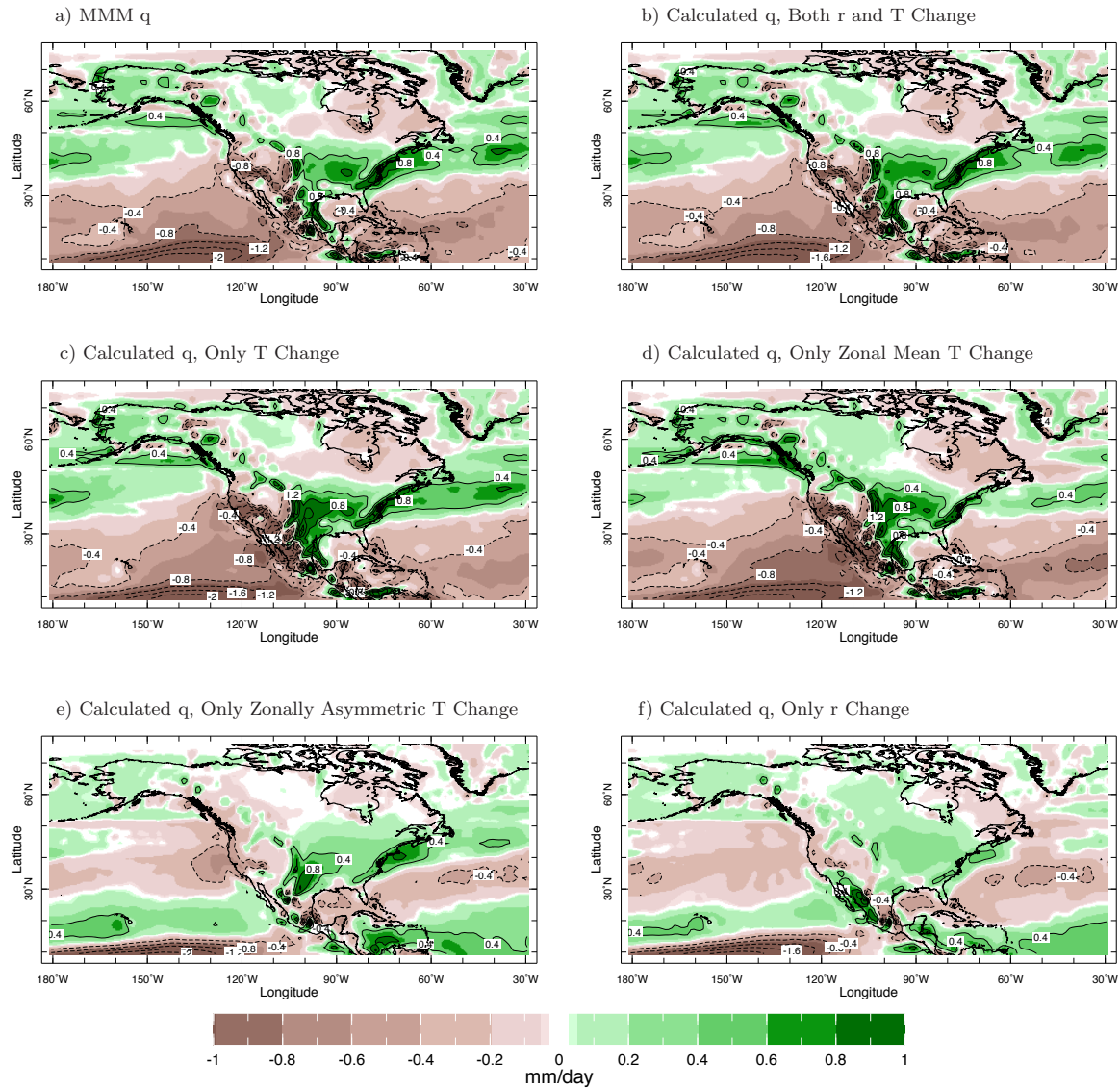


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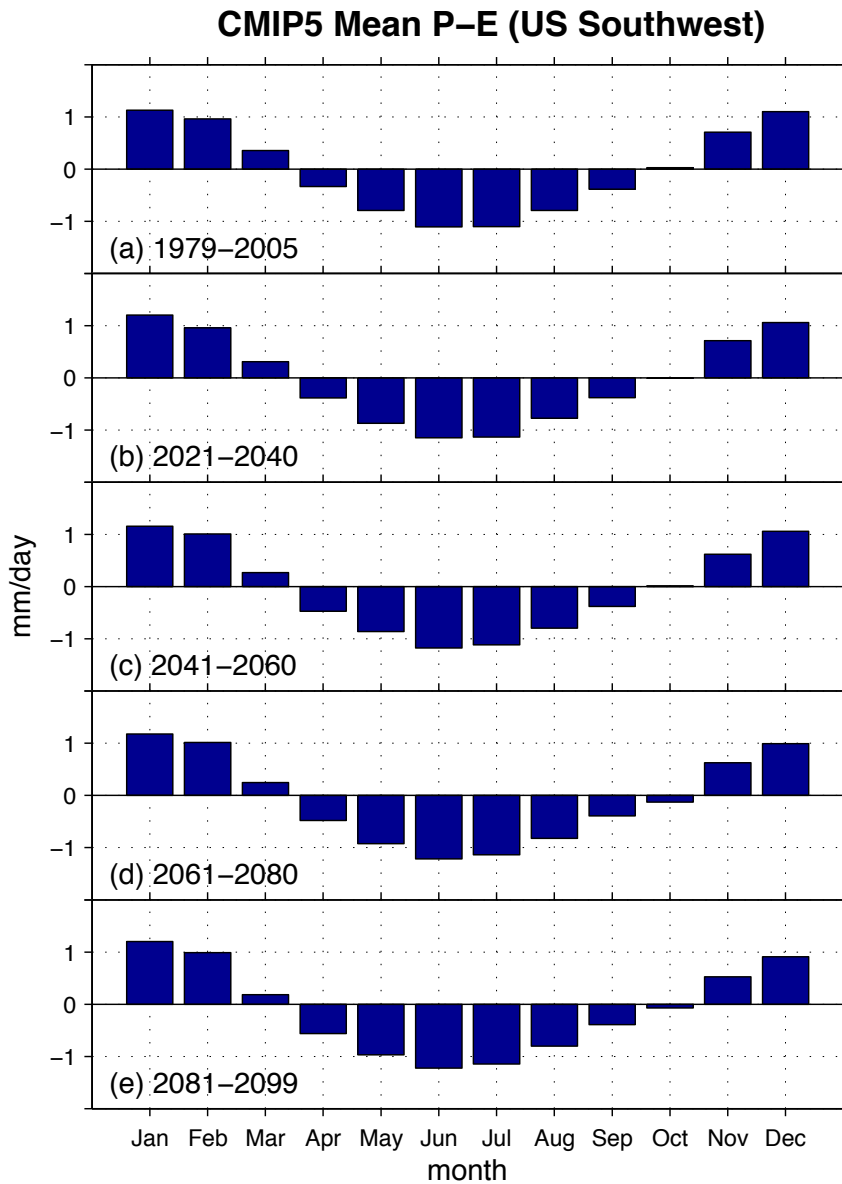


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