



HAL
open science

Impact of surface temperature biases on climate change projections of the South Pacific Convergence Zone

Cyril Dutheil, Margot Bador, Matthieu Lengaigne, Jérôme Lefèvre, Nicolas Jourdain, Jérôme Vialard, Swen Jullien, Alexandre Peltier, Christophe E. Menkès

► To cite this version:

Cyril Dutheil, Margot Bador, Matthieu Lengaigne, Jérôme Lefèvre, Nicolas Jourdain, et al.. Impact of surface temperature biases on climate change projections of the South Pacific Convergence Zone. *Climate Dynamics*, 2019, 53, pp.3197-3219. 10.1007/s00382-019-04692-6 . hal-02401413

HAL Id: hal-02401413

<https://hal.science/hal-02401413>

Submitted on 12 Dec 2019

HAL is a multi-disciplinary open access archive for the deposit and dissemination of scientific research documents, whether they are published or not. The documents may come from teaching and research institutions in France or abroad, or from public or private research centers.

L'archive ouverte pluridisciplinaire **HAL**, est destinée au dépôt et à la diffusion de documents scientifiques de niveau recherche, publiés ou non, émanant des établissements d'enseignement et de recherche français ou étrangers, des laboratoires publics ou privés.

Impact of surface temperature biases on climate change projections of the South Pacific Convergence Zone

Cyril Marcel Dutheil¹, Bador M.², Lengaigne M.³, Lefevre J.¹, Jourdain N.C.⁴, Vialard J.³, Jullien S.⁵, Peltier A.⁶, and Menkes C.¹

¹ IRD (Institut de Recherche pour le Développement)-Sorbonne Universités (UPMC, Université Paris 06)-CNRS-MNHN-IPSL, LOCEAN Laboratory, IRD Nouméa BP A5, 98848 Nouméa cedex, New Caledonia;

² Climate Change Research Centre and ARC Centre of Excellence for Climate Extremes, School of BEES, University of New South Wales, Sydney, New South Wales, Australia.

³ LOCEAN-IPSL, Sorbonne Universités, UPMC, Université Paris 06, CNRS-IRD-MNHN, Paris, France.

⁴ *Univ. Grenoble Alpes / CNRS / IRD / G-INP, IGE, Grenoble, France.*

⁵ Ifremer, Univ. Brest, CNRS, IRD, Laboratoire d'Océanographie Physique et Spatiale (LOPS), IUEM, Plouzané, France

⁶ Météo France, Nouvelle Calédonie

Corresponding author:

C. Dutheil

IRD Nouméa BP A5

98848 Nouméa cedex

New Caledonia

FRANCE

Email: cyril.dutheil@ird.fr

Abstract

The South Pacific Convergence Zone (SPCZ) is poorly represented in global climate models, with trademark biases such as the “double convergence zones” and an equatorial cold tongue that extends too far westward. Such biases limit our confidence in climate change projections for this region. Here, we first show that simulations using a regional atmospheric model capture the SPCZ mean state and interannual variability well over the 1980-2016 period. We then perform climate change experiments by correcting the model surface and lateral boundary conditions on the basis of their projected changes for 2080-2100, obtained from the RCP8.5 seasonalCMIP5 multi-model mean. We also test a statistical method that corrects the Sea Surface Temperature (SST) projected changes on the basis of present-day biases. This projected SST correction strategy yields a considerably larger southwestern Pacific rainfall reduction (~30%) than

35 in CMIP5 simulations (~7%). Sensitivity experiments indicate that the weak CMIP5 response is due to biases in the
36 projected CMIP5 SST, which alter future SST gradients, thereby changing the circulation and mitigating the reduction
37 in south Pacific humidity convergence. While several studies have also shown that CMIP5 models project stronger
38 SPCZ variability, with more frequent zonal SPCZ events, our experiments do not support this increase in the number of
39 zonal SPCZ events. Additional simulations with different physical parameterizations yield very similar results,
40 including the southwestern significant Pacific drying and no increase in the number of zonal SPCZ events.

41

42 **Keywords:** Regional climate models, South Pacific Convergence Zone, Precipitation, Sea surface temperature

43

44 **1. Introduction**

45

46 The South Pacific Convergence Zone is the largest convective area of the southern hemisphere in austral
47 summer (Kiladis et al., 1989; Fig. 1a) and plays a major role in earth's climate (Vincent, 1994; Van der Wiel et al.
48 2016a). The SPCZ is the main provider of rainfall for a vast majority of South Pacific island nations, where local
49 hydrological conditions are very sensitive to its displacements (e.g. Alexander et al., 2006; Gouriou and Delcroix, 2002;
50 Murphy et al., 2015). The location of this rain belt not only strongly constrains the regional hydrological cycle but is
51 also the breeding ground of South Pacific tropical cyclones (Vincent et al., 2011). The SPCZ exhibits a diagonal
52 orientation along a north-west/south-east axis, stretching from the Solomon Islands to the central subtropical Pacific
53 (thick line in Figure 1a; Kiladis et al., 1989; Vincent, 1994), which has been attributed to the equatorward propagation
54 and the diagonal orientation of eastward propagating Rossby wave trains (Matthews 2012; van der Kiel et al. 2015).
55 The climatological characteristics of the SPCZ has been related to the underlying sea surface temperature (SST)
56 gradients (Van der Wiel et al. 2016b). The SPCZ interannual variability affects the frequency of droughts, floods, and
57 cyclones experienced by vulnerable island countries (Vincent, 1994; Alexander et al., 2006; Kirono et al., 2011;
58 Vincent et al., 2011). These variations are strongly tied to the El Niño/Southern Oscillation (ENSO), the most energetic
59 climate mode at interannual timescales. El Niño (resp. La Niña) events tend to induce a north-eastward (resp. south-
60 westward) displacement of the SPCZ (e.g Folland, 2002; Juillet-Leclerc et al., 2006; Kidwell et al., 2016; Trenberth,
61 1976), see also Fig. 4). This picture is much more radical during “extreme” El Niño events such as in 1982-1983, 1997-
62 1998 and 2015-2016, when the SPCZ undergoes an equatorward swing of up to ten degrees of latitude and collapses to
63 a more zonally oriented structure (e.g. Vincent et al., 2011; Cai et al., 2012). The unusual SPCZ shift and re-orientation
64 during these extreme El Niños likely arise from the non-linear rainfall response to ENSO-related sea surface
65 temperature (SST) signals (Chung et al., 2014). These “extreme” events have outsized impacts relative to weaker El
66 Niños, such as massive drought, food shortage (Kirono et al., 2011) and a considerable reorganisation of cyclogenesis
67 (Vincent et al., 2011), with an unusual occurrence of cyclones in French Polynesia.

68 Global warming also has the potential to influence the SPCZ rainfall location, intensity or variability,
69 which would impact the water resources of many South Pacific islands. The detection of such changes in observations
70 remains uncertain because of the strong aliasing by natural decadal/multi-decadal variability over the past 40 years
71 (Deser et al., 2010; McGree et al., 2014; Meng et al., 2012; Solomon and Newman, 2012; Tokinaga et al., 2012; Vecchi
72 et al., 2008). The assessment of the tropical Pacific response to global warming hence heavily relies on climate
73 projections from the Coupled Model Intercomparison Project (CMIP; Meehl et al., 2007; Taylor et al., 2012). These

74 projections for the late 21st century generally depict a robust tropical Pacific SST response to greenhouse warming (see
75 Fig. 1d, and Fig. 13a in Li et al., (2016)). This response includes a larger warming rate along the equator than in the off-
76 equatorial central basin (e.g. Cai et al., 2015; Collins et al., 2010; Widlansky et al., 2013) and a robust precipitation
77 increase in the equatorial Pacific (Fig. 1d, e.g. Watanabe et al., 2014; Li et al., 2016). This warming pattern also drives a
78 drying of the southeast SPCZ margin in CMIP5 models, attributed to the anomalous transport of dry subtropical air into
79 the SPCZ region (Widlansky et al., 2013), in response to increased SST meridional gradients in the southeast tropical
80 Pacific. Within the core of the SPCZ region, west of the dateline, two mechanisms compete in driving the rainfall
81 changes (Widlansky et al., 2013): the thermodynamical “wet gets wetter” mechanism where increased atmospheric
82 water vapour content under global warming results in increased rainfall in the SPCZ convective region, and the
83 dynamical “warmest gets wetter” mechanism where the comparatively weaker warming in the SPCZ region compared
84 to that in the equatorial region results in anomalous moisture divergence and hence decreased rainfall. This results in a
85 general tendency for the SPCZ core to dry for moderate tropical warming (~1-2°C), when increased transport of moist
86 air toward the equatorial warming dominates the local increase of evaporation, and to moisten for stronger tropical
87 warming (~3°C), when local evaporation, which increases the atmospheric moisture, dominates the dynamical changes
88 (Widlansky et al., 2013). Finally, despite the absence of consensus on how ENSO-driven SST variability may change in
89 the future (Collins et al., 2010; Power et al., 2013; Cai et al., 2015), CMIP5 model projections over the 21st century do
90 indicate a consistent intensification of ENSO-related precipitation in the central/eastern equatorial Pacific because
91 rainfall responds nonlinearly to absolute SST (Power et al., 2013). These climate model projections also display an
92 increased occurrence of “extreme” El Niño and zonal SPCZ events projected over the 21st century because of a
93 meridional SST gradient reduction in the South Pacific (Cai et al., 2012, 2014) allowing the SPCZ to shift to the north
94 more easily. As we will see, the regional simulations in our study do not support the increased frequency of zonal
95 SPCZ in CMIP3-5 projections, described by Cai et al. (2012).

96 While the representation of tropical Pacific climate has slightly improved in CMIP5 relative to CMIP3
97 models (Bellenger et al., 2014; Grose et al., 2014), these models still exhibit long-standing biases. They indeed
98 generally produce an excessive cold tongue that extends too far westward (e.g., Li and Xie, 2014; Li et al., 2015;
99 Mechoso et al., 1995; Zheng et al., 2012; Fig. 1c) and a “double ITCZ” bias (Brown et al., 2011; Grose et al., 2014;
100 Irving et al., 2011; Niznik et al., 2015; Perkins, 2011), i.e. a tendency for the SPCZ to be too zonal and to extend too far
101 eastward (Fig. 1 b,c). The southeast SPCZ margin drying projected by CMIP models thus occurs in a region where
102 these models do not correctly simulate the SPCZ (Fig. 1a,b). The precipitation increase in the equatorial western Pacific
103 (Fig. 1d) also occurs in a region where most CMIP models exhibit a cold bias (Brown et al., 2013; Grose et al., 2014;
104 Fig. 1c). Since many aspects of regional climate change depend upon the present-day climatology (e.g., Held and
105 Soden, 2006; Huang et al., 2013; Matsueda and Palmer, 2011; Xie et al., 2015), these model biases may impair the
106 reliability of projected future precipitation changes in the SPCZ region. Indeed, Li et al. (2016) showed that the lower
107 this cold bias, the largest the projected drying in the western Pacific. This clearly indicates an influence of present-day
108 SST biases on projected precipitation in CMIP5 models (see Fig 12. from Li et al., 2016).

109 A possible strategy to reduce the influence of CMIP biases on climate change projections is to force an
110 atmosphere-only model with bias-corrected SSTs (e.g. Knutson et al., 2008; Nguyen et al., 2012), i.e. to add the global
111 warming CMIP5 SST pattern (ΔSST_{GW}) to the present-day observed mean state. Several authors have used this strategy
112 to assess the sensitivity of the SPCZ response to CMIP mean state biases (Brown et al., 2015; Evans et al., 2016;
113 Widlansky et al., 2013). Using a global atmospheric model, Widlansky et al. (2013) demonstrated that the biases in

114 CMIP tropical mean SST strongly contribute to the uncertainty of twenty-first-century rainfall projections in the South
115 Pacific, with a possible future drying of the western SPCZ portion when those biases are corrected. Using a bias-
116 adjusted SST strategy in a number of Regional Climate Models (RCMs), Evans et al. (2016) showed that these
117 southwest Pacific regional climate projections were much more sensitive to the choice of the regional climate model
118 than to the CMIP model used to specify the RCM boundary conditions. Overall, the ensemble of simulations performed
119 in their study hardly displayed any consensus for projected SPCZ changes, although a majority of ensemble members
120 projected a drying of the SPCZ, in line with Widlansky et al. (2013).

121 The above bias-correction strategy however does not account for possible biases in the CMIP-projected
122 ΔSST_{GW} . The present day common cold tongue bias in CMIP5 models may for instance influence the projected
123 ΔSST_{GW} . A way to alleviate this issue is to use a simple statistical method based on the relationship between CMIP5
124 present-day biases and projected changes, referred to as “observational constraint” or “emergent constraint” (e.g.
125 Bracegirdle and Stephenson 2012). This approach reduces the impact of the present-day model biases on the projected
126 response in the Arctic (e.g. Bracegirdle and Stephenson, 2013) and Indian Ocean (Li et al., 2016; Parvathi et al., 2017).
127 In the tropical Pacific, Li et al. (2016) applied this “observational constraint” approach to correct CMIP5 projections for
128 the tropical Pacific based on present-day rainfall bias in the equatorial western Pacific. This strategy yielded a stronger
129 ΔSST_{GW} in the eastern than in the western equatorial Pacific, and a strong reduction of the inter-model spread in the
130 model precipitations projections. This method is further described in section 2.2. In the present paper, we will use a
131 similar method to correct CMIP5 SST projections based on the present-day SST bias, and investigate its impact on
132 projected SPCZ changes.

133 Using a similar modelling framework as in Brown et al. (2015) and Evans et al. (2016), we investigate
134 the robustness of the SPCZ response to climate change to i) model physics, ii) present-day SST biases and iii)
135 correcting the CMIP5-projected SST change ΔSST_{GW} on the basis of present-day biases. We specifically address the
136 following questions. What are the expected robust changes in the austral summer SPCZ climatology in response to
137 anthropogenic forcing? How will these changes alter the interannual SPCZ variability? To answer these questions, we
138 use the Weather Research and Forecasting (WRF) regional climate model. The paper is organized as follows. Section 2
139 describes the experimental design. Section 3 provides a thorough evaluation of the SPCZ mean state and variability
140 simulated by our present-day simulation, and we compare our simulation to a simulation where we applied the CMIP5
141 SST bias. Section 4 discusses the future changes in the SPCZ mean state and variability derived from our experimental
142 strategy. We also discuss the sensitivity of our results to the CMIP5 model present-day SST biases and to correcting
143 ΔSST_{GW} based on emergent constraints. Section 5 investigates the sensitivity of our results to the model physics.
144 Finally, Section 6 provides a summary and a discussion of our results against previous studies.

145

146

147 **2. Experimental design and methods**

148

149 **2.1 WRF Regional model configuration**

150

151 The two different model configurations used in this study and detailed in Table 1 are based on the
152 Weather Research and Forecasting version 3.6.1 regional atmospheric model, which solves compressible, non-
153 hydrostatic Euler equations with the Advanced Research WRF (ARW) dynamical solver (Skamarock and Klemp,

154 2008). The simulated domain encompasses the tropical Pacific region [101°E-59°W; 26°N-42°S] using a 105 km
155 horizontal grid spacing and 32 pressure levels. In the following, only a subset of this domain centered on the SPCZ
156 region is shown (Fig. 1a).

157 This first configuration is based on the study of Lefèvre et al., (2016). It includes the Lin et al., (1983)
158 microphysics scheme, the Community Atmosphere Model (Collins et al., 2004) for shortwave and longwave radiation,
159 the UW planetary boundary layer (Bretherton and Park, 2009) with the Monin–Obukhov surface layer parameterization,
160 and the Noah land surface model (Chen and Dudhia, 2001). The parameterization of subgrid-scale convection is taken
161 from the Zhang-McFarlane scheme (Zhang and McFarlane, 1995). This configuration will be referred to as the “ZM”
162 configuration in the following.

163 The second configuration is partly based on Jourdain et al., (2011) and Jullien et al., (2014). The set of
164 physical parameterizations used is completely different from the ZM configuration with Betts-Miller-Janjić
165 parameterization for convection (Janjić, 1994), the WRF single-moment three-class microphysics scheme (Hong et al.,
166 2004), the Rapid Radiation Transfer Model for the longwave and shortwave radiation (Iacono et al., 2008) and the
167 Yonsei University planetary boundary layer (Noh et al., 2003). This configuration will be referred to as the “BMJ”
168 configuration in the following.

169

170 **2.2 Present-day and climate change experiments**

171 The surface and lateral atmospheric boundary conditions were generated from the multi-model mean
172 (MMM) of an ensemble of 31 CMIP5 climate models (see Table ; Taylor et al. 2012). For each model, we consider one
173 simulation covering the historical period and the 21st century under the Representative Concentration Pathway 8.5
174 (RCP8.5) of aerosols and greenhouse gases (GHG) scenario, for which the radiative forcing reaches 8.5 W.m⁻² by 2100
175 (van Vuuren et al. 2011). When several members were available for a given CMIP5 model, only the first available
176 ensemble member (referred to r1i1p1 in the CMIP5 naming convention, or r2i1p1 if not available) was considered, to
177 avoid giving more weight to some models than others in the MMM. The CMIP5 fields used to build the boundary
178 conditions and the initial state are listed in Table 2. Table 3 provides a summary of the performed simulations, which
179 are further detailed in the following. It should be noted that the 31 models are not really independent in the sense that
180 many of them share duplicated codes and common history (Herger et al., 2018; Knutti et al., 2013).

181 A pair of present-day (PD) simulations is performed for each of the two model configurations (ZM and
182 BMJ), from January 1980 to May 2016. For each of the ZM and BMJ configurations (indicated by subscripts in the
183 acronyms when needed), a 36-yr present-day simulation (labelled PD_OBS hereafter) has been designed as a dynamical
184 downscaling of the NCEP2 reanalysis (NCEP2; Kanamitsu et al., 2002). Surface and lateral boundary conditions are
185 taken from the 6-hourly outputs of this reanalysis. A second 36-yr present-day simulation uses the same NCEP2
186 boundary conditions to which the present-day monthly climatological CMIP5 MMM SST bias is added to the observed
187 SST climatology (labelled PD_CMIP hereafter). The comparison of this simulation with CMIP5 MMM and CC_OBS
188 simulation allows us to test the impact of our model setup, and the impact of the historical CMIP5 SST biases, and will
189 be discussed in Section 4. SST biases are only applied in the SPCZ region, within [130°E-290°E; 7°N- 30°S], with a
190 smooth transition to present-day SST near the boundaries of this region. The lateral and sea surface conditions outside
191 this box remain the same as in PD_OBS. In all these present-day simulations, the CO₂ forcing is kept constant at a level
192 representative of the recent period (379ppm).

193 To assess the southwest Pacific response to climate change (CC), we further conduct a set of three future
194 experiments for each model configuration. For the first two experiments (labelled CC_OBS and CC_CMIP), we apply
195 the climatology of surface and lateral boundary future changes on top of the conditions of PD_OBS and PD_CMIP.
196 These future changes are calculated as the difference between the 2080-2100 and 1989-2009 of MMM monthly
197 climatologies (Fig. 1d), following the same methodology as in Knutson et al. (2008) and Walsh (2015). We apply a
198 constant CO₂ forcing representative of the RCP8.5 projections at the end of the 21st century (845ppm). The CMIP5
199 fields used to build the boundary conditions and the initial states are listed in Table 2. This experimental strategy
200 intentionally assumes no change in variability between the present and future climate for two main reasons. First, the
201 reported lack of consensus on changes in amplitude of ENSO-related SST signals in future CMIP model projections
202 (some global climate models showing an increase in the amplitude of ENSO SST variability in the future, others
203 showing a decrease or, no statistically significant changes; see Bellenger et al. 2014) are an incentive to test the null
204 hypothesis, i.e. that the ENSO SST variability will not significantly change. Second, Cai et al. (2012) found a consistent
205 increase in the frequency of zonal SPCZ events despite this lack of consensus and attributed this increase to the changes
206 in the mean meridional SST gradient in the South Pacific. The present experimental framework that only assume a
207 mean state change allow examining if the mechanism suggested by Cai et al. (2012) operate in a more realistic
208 framework (i.e. without CMIP5 mean present-day biases).

209 We apply a constant CO₂ forcing representative of the RCP8.5 projections at the end of the 21st century (845ppm). The
210 CMIP5 fields used to build the boundary conditions and the initial states are listed in Table 2. The third experiment
211 (CC_COR), is similar to CC_OBS experiments, except that we use the CMIP5 SST anomalous pattern (Fig. 1e)
212 statistically corrected following the method of Li et al. (2016), $\Delta SST_{COR_{GW}}$. In order to produce a corrected SST
213 projection $\Delta SST_{COR_{GW}}$, we first assess the statistical relationship between the historical rainfall bias and ΔSST_{GW}
214 (Fig. 2) in the equatorial western Pacific (140°E-170°W; 2°S-2°N). We find a significant correlation of -0.56 (p-value =
215 0.001) between the mean precipitation and ΔSST_{GW} across 31 CMIP5 models (Fig. 2), which is consistent with Li et al.
216 (2016) who found a -0.52 correlation across 24 CMIP5 models (their Figure 6a). This analysis demonstrates that the
217 drier the present-day west Pacific (and the stronger the cold tongue bias, not shown), the stronger the projected SST
218 warming in the equatorial western Pacific. We use this linear relation between the mean western equatorial Pacific
219 present-day precipitation, and ΔSST_{GW} to correct ΔSST_{GW} , as in Li et al. (2016). The corrected CMIP5 MMM
220 projection exhibits a weaker western Pacific warming (Fig. 1e). This will modify SST gradients quite significantly in
221 this region, relative to ΔSST_{GW} .

222 In the rest of the manuscript we will refer to the difference between historical CMIP5 climatological
223 SST, and observed SST as the “*historical CMIP5 SST bias*” and to the difference between ΔSST_{GW} and $\Delta SST_{COR_{GW}}$
224 as the “ *ΔSST_{GW} bias*”. (Strictly speaking this last quantity is not bias but an estimated bias).

225

226 2.3 Analysis methods

227 The SPCZ position at each longitude is taken as the location of the maximum precipitation between the
228 equator and 30°S, where the precipitations are higher than 6mm.day⁻¹. The mean SPCZ slope is calculated as a linear fit
229 to this SPCZ position between 160°E and 140°W. As in Vincent et al. (2011), we define LatW as the SPCZ latitude
230 average between 160°E and 180°E, and latE as the SPCZ latitude averaged between 170°W and 150°W. El Niño (La

231 Niña) events are defined as austral summer seasons (December-January-February; DJF) when the DJF Niño3.4 SST
232 anomaly is greater (lower) than 0.5°C, following NOAA’s index:
233 http://www.cpc.noaa.gov/products/analysis_monitoring/ensostuff/ensoyears.shtml. We classify El Niño events into two
234 classes: extreme El Niño events that encompass the three largest events (1982/1983, 1997/1998, 2015/2016) following
235 Santoso et al. (2017) and moderate El Niño events that gather all the other events.

236 To investigate the relative importance of moisture changes (referred to as “thermodynamical”), and
237 large-scale atmospheric circulation changes (referred to as “dynamical”) in shaping the pattern of moisture divergence
238 changes between future and present, we use the same method as discussed in Widlansky et al. (2013) with a simple
239 decomposition of surface moisture divergence anomalies (the anomaly being the difference between CC_OBS_{ZM} and
240 PD_OBS_{ZM}) into surface divergence of:

241 - projected moisture changes advected by the mean circulation: $\delta TH = \nabla \cdot (\bar{u}q')$ (Eq. 1)

242 - mean moisture advected by the projected circulation changes: $\delta DYN = \nabla \cdot (u'\bar{q})$ (Eq. 2)

243 where u is the horizontal surface wind vector, q is the specific surface humidity, the overbar denotes the values from the
244 present day simulation and primes denote differences between future and present day simulations. The cross terms are
245 not discussed here as δTH and δDYN dominate the balance in the region of interest (not shown).

246

247 **3. Results: Present-day simulations**

248 **3.1 Climatological features**

249 The observed SPCZ persists throughout the year, but reaches its maximum extension during austral
250 summer (DJF). At this time of the year, the SPCZ extends from 2°S on its western edge to 22°S on its eastern edge and
251 exhibits a moderate tilt in its western portion (150°E-160°W) and more southeastwardly tilted portion in its eastern
252 portion (160°W-140°W). The surface winds converge toward the SPCZ (Fig. 3a).

253 The PD_OBS_{ZM} experiment simulates the observed climatological DJF precipitation pattern reasonably
254 well (Fig. 3b), including the southeastward SPCZ tilt. The surface wind intensity and convergence are correctly
255 reproduced, resulting in a realistic moisture convergence (not shown). This simulation however exhibits some biases,
256 including a ~12% underestimation of ITCZ rainfall and a ~30% underestimation of the western equatorial Pacific
257 rainfall. PD_OBS_{ZM} also overestimates precipitation in the core of the SPCZ. Despite these biases, this simulation
258 reproduces the location of the observed SPCZ very accurately (yellow and red curves on Figure 3a), a major asset for
259 the objectives of the present study. In addition, this simulation seems to better reproduce the position of SPCZ than the
260 other studies using regional models (Evans et al., 2016).

261 These precipitation patterns are further compared to those of the NCEP2 reanalysis (Fig. 3c), which is
262 used as boundary conditions in the present study. The NCEP2 reanalysis also captures the main observed climatological
263 precipitation pattern reasonably well. In contrast with PD_OBS_{ZM}, it tends to overestimate ITCZ (~15%) and equatorial
264 western Pacific (~5%) precipitation. The NCEP2 dataset also exhibits a diagonally oriented SPCZ, although this
265 reanalysis appears to overestimate the SPCZ tilt in its western portion relative to PD_OBS_{ZM} and CMAP and
266 underestimate it in its eastern portion. Like PD_OBS_{ZM}, the NCEP2 re-analysis overestimates precipitation in the core
267 of the SPCZ.

268 A more quantitative assessment of the SPCZ climatological characteristics simulated by PD_OBS_{ZM} (red
269 circle), NCEP2 (green circle) and the CMIP5 models (boxplot) is provided in Figure 4. Figure 4a clearly illustrates that
270 all CMIP5 models considerably underestimate the slope of the SPCZ as already reported in past literature (e.g. Brown et
271 al., 2012), with the historical CMIP5 MMM exhibiting a four times weaker slope than in observations (0.06°S/°E as
272 compared to 0.26°S/°E for CMAP). This slope varies considerably among individual CMIP5 simulations, ranging from
273 a completely zonal SPCZ to a more realistic diagonally oriented but still underestimated SPCZ slope of up to 0.14°S/°E.
274 PD_OBS_{ZM} captures this SPCZ slope far more accurately (0.22°S/°E). The strong underestimation of the SPCZ slope in
275 historical CMIP simulations largely results from a large northward bias in the location of its eastern portion (Fig. 4b;
276 9°S for the CMIP MMM, 13.5°S for observations). This location is very well captured in PD_OBS_{ZM} (14°S). The
277 location of the western SPCZ portion (Fig. 4c) around 8°S is reasonably well captured by PD_OBS_{ZM}, NCEP2 and
278 CMIP5 simulations. PD_OBS_{ZM} also captures the South Pacific (0-30°S; 150°E-240°E) rainfall pattern accurately (Fig.
279 4d), with a pattern correlation of 0.96, compared to a correlation ranging from 0.5 to 0.9 for CMIP5 and 0.87 for
280 NCEP2. Overall, the PD_OBS_{ZM} simulation outperforms all CMIP5 models and behaves as accurately or slightly better
281 than NCEP2, as far as the SPCZ climatological features are concerned.

282

283 3.2 Interannual variability

284 Figure 5 exhibits a composite of anomalous rainfall and surface winds during the different ENSO
285 phases. Moderate El Niño and La Nina events generally exhibit opposite patterns. Moderate El Niño years (resp. La
286 Nina years) are characterized by modest positive (negative) rainfall anomalies in the vicinity of the dateline, while weak
287 negative (positive) rainfall anomalies prevail in the western equatorial Pacific and south of the climatological SPCZ
288 position in the southwest Pacific (Fig 5a,c). During these moderate El Niño events (resp. La Nina), the SPCZ
289 experiences a 3° northward (southward) shift between 150°E and 150°W. This picture is very different during extreme
290 El Niño events (Fig. 5b): the equatorial rainfall increase is four times larger than during moderate El Niño events and its
291 pattern is considerably extended and shifted eastward, with a maximum near 165°W. The South Pacific rainfall
292 decrease is also about twice as large as during moderate events and extends further east to 140°W. Finally, the area
293 covered by the ITCZ in the western equatorial Pacific also exhibits a large drying that is not present during moderate El
294 Niño events. These stronger and spatially-extended rainfall anomalies associated with extreme El Niño events translate
295 into larger SPCZ displacements: the SPCZ eastern portion shifts northward by up to ten degrees, while its western
296 portion remains close to its climatological position, which produces a more zonally oriented SPCZ from 150°E to
297 150°W.

298 From year to year, the eastern portion of the SPCZ experiences meridional displacements, with the
299 largest northward excursions occurring in austral summer 1982 (~+10°), 1992 (~+7°), 1997 (~+10°) and 2015 (~+8°)
300 (Fig. 6). As discussed in Vincent et al. (2011), these years are indeed characterized by an extreme northward swing of
301 the SPCZ, which ends up with a more zonally oriented structure. PD_OBS_{ZM} and NCEP2 accurately reproduce these
302 SPCZ excursions (0.71 correlation between CMAP and both datasets), despite a 20% overestimation of the variability
303 in NCEP2 (NCEP2 STD=4.99° vs CMAP STD=4.1°). In particular, PD_OBS_{ZM} correctly simulates the SPCZ extreme
304 northward excursions during the 1982, 1992, 1997 and 2015 El Niño events, with a tendency for both PD_OBS_{ZM} and
305 NCEP2 to overestimate the northward excursion of the SPCZ during these events (Fig. 6). The variability of the western
306 portion is also well captured by both NCEP2 and PD_OBS_{ZM}, with a ~0.7 correlation to CMAP (not shown). Overall, in

307 this section, we have demonstrated the capacity of our simulation to reproduce reasonably well the present-day SPCZ
308 mean state (in particular: much better than in CMIP5 models) as well as its interannual variability.

309

310 **4. Results: Response to climate change**

311 **4.1 Climatology**

312 In this section, we investigate the climatological SPCZ response to anthropogenic forcing in our 3 sets of
313 experiments, CMIP_{ZM}, OBS_{ZM} and COR_{ZM} and compare it with that in the CMIP5 MMM. We start with a description of
314 the changes in CMIP5 MMM, and then compare them with the CMIP_{ZM} experiments to assess the impact of our model
315 setup. We will then compare CMIP_{ZM} and OBS_{ZM} to evaluate the impact of the historical CMIP5 SST bias. Finally,
316 OBS_{ZM} and COR_{ZM} are used to evaluate the effect of the ΔSST_{GW} bias.

317 The CMIP5 MMM precipitation response to climate change is characterized by a consistent and
318 significant rainfall increase along the equator, reaching up to 3 mm.day⁻¹ west of the dateline (Fig. 7a). These
319 simulations also display a consistent and statistically significant drying of up to 1 mm.day⁻¹ east of 150°W and south of
320 10°S. In the southwest Pacific (10°S-30°S; 160°E-160°W) the rainfall change is not statistically significant, with less
321 than 75% of the models simulating a rainfall increase. Overall, these modest rainfall changes only result in very modest
322 future changes in the climatological SPCZ location in the CMIP5 CC simulations (solid and dotted black lines in Figure
323 7a). The projected wind changes depict an intensification of the trade winds in the southeastern part of the domain,
324 converging towards the equatorial region (Fig. 7a).

325 Let us first explore how our model compares with the CMIP5 MMM. The CC_CMIP5_{ZM} simulation displays a
326 similar significant precipitation increase in the equatorial region (Fig 7b; ~2.5 mm.day⁻¹ in CMIP5_{ZM} experiment
327 compared to ~2.2 mm.day⁻¹ in CMIP5 MMM, on average in the red box Fig. 7d). This significant precipitation increase
328 is confined to the western/central Pacific or outside the 5°N-5°S band at east of 170°W. CC_CMIP5_{ZM} also yields a
329 significant drying in the southeast Pacific, extending up to 175°E, and stronger than in the CMIP5 MMM (~1.0
330 mm.day⁻¹ in CMIP5_{ZM} experiment compared to ~0.4 mm.day⁻¹ in CMIP5 MMM, on average in the green box Fig. 7d).
331 These anomalies do not affect the climatological SPCZ position, which remains similar from 150°E to 160°W (solid
332 and dotted black lines on Figure 7b), but the rainfall decrease east of 170°W leads to a shrinkage of the SPCZ, defined as
333 the region where precipitation is stronger than 6mm.day⁻¹.

334 As expected, correcting the historical CMIP5 SST bias considerably impact the mean simulated rainfall
335 pattern. CC_OBS_{ZM} (Fig. 7c) displays a significant rainfall increase in the equatorial region between 7°S-10°N, much
336 stronger than CMIP5_{ZM} experiments (~3.5 mm.day⁻¹ in OBS_{ZM} experiment compared to ~2.5 mm.day⁻¹ in CMIP5_{ZM}
337 experiment, on average in the red box Fig. 6d), reaching a maximum in the western/central Pacific, near 5°N (Fig. 7c).
338 The CC_OBS_{ZM} simulation also displays a significant rainfall decrease (~0.9 mm.day⁻¹ on average in the green box Fig.
339 7d), similar to the CMIP5_{ZM} experiment, south of 7°S and north of the SPCZ climatological position extending from the
340 eastern Pacific to 180°. This response represents a 15% rainfall decrease within the [5°S-15°S; 180°-150°W] region
341 compared to the present-day climatology. These large-scale and intense anomalies affect the climatological SPCZ
342 position in its eastern portion (solid and dotted black lines on Figure 7c), with a southward shift of 4° of the maximum
343 precipitation.

344 Let us finally examine the effect of correcting the ΔSST_{GW} bias. CC_COR_{ZM} (Fig. 7d) yields a weaker increase

345 of precipitation in the western/central equatorial region compared to CC_OBS_{ZM} (~3 mm.day⁻¹, on average in the red
346 box Fig. 7d). The area of significant increase located north of the equator is confined in the western Pacific in
347 CC_COR_{ZM} while it extends throughout the Pacific in CC_OBS_{ZM}. CC_COR_{ZM} also yields a larger drying south of 5°S
348 and east of 180°E (~1.9 mm.day⁻¹ on average in the green box Fig. 6d), which extends farther into the western Pacific
349 than in CC_OBS_{ZM} (Fig. 7d). This represents a ~20% decrease in rainfall within the [15°S-5°S; 160°E-180°E] region
350 compared to present-day climatology. In these two experiments (Fig. 7c,d) these anomalies induce a large southward
351 shift of the climatological SPCZ position (~4°). The trade winds in the south-central part of the domain [180°E-150°W;
352 15°S-5°S] increase by 40% compared to present-day climatology in CC_COR_{ZM}, especially in the southwestern Pacific
353 where the anomaly of trade winds cross the equator to the north where maximum precipitation anomalies are found.

354 We further use simple surface moisture convergence diagnostics similar to those proposed in Widlansky et al.
355 (2013) to provide insights on the mechanisms driving the simulated rainfall response to climate change in the CMIP5
356 models and our experiments (Fig. 8). The anomalous patterns in surface moisture divergence (Figure 8a,b,c,d)
357 qualitatively match those of the precipitation changes (Fig. 7a,b,c,d) for both the CMIP5 models and our experiments,
358 indicating that this is an appropriate way to diagnose precipitation changes. Moisture divergence changes are almost
359 entirely driven by wind divergence changes in our experiments and in the CMIP5 models (Fig. 8e,f,g,h), with moisture
360 changes playing a smaller role (Fig. 8i,j,k,l). The anomalous moisture convergence in the equatorial region is also
361 largely driven by dynamical changes, with thermodynamics only damping the dynamical changes (Fig. 8e,f,g,h vs
362 i,j,k,l). Thermodynamical effects are slightly stronger in CMIP5_{ZM} (Fig. 8j) than in OBS_{ZM} (Fig. 8k), and hence damp
363 the dynamical effect more, leading to larger precipitation changes in OBS_{ZM} in the equatorial region. In the southeast
364 Pacific, the increased anomalous moisture divergence that dries the SPCZ eastern edge is also largely driven by
365 circulation changes (Fig. 8e,f,g,h).

366 Regarding the southwest Pacific and SPCZ region, the moisture divergence anomalies and related precipitation
367 changes are similar in the OBS and CMIP experiments (Fig. 7b,c and Fig. 8b,c), for which the tendency for dynamical
368 processes to dry the western portion of the SPCZ is partially compensated by thermodynamical effects (Fig.
369 8b,c,f,g,j,k). On the other hand, the COR experiment yields a stronger drying of the SPCZ western margin, that can be
370 related to stronger dynamical processes in COR (Fig. 8h) than in OBS (Fig. 8g). The wind changes responsible for these
371 moisture convergence modifications can be related to the corrected projected SST pattern in the COR experiment
372 relative to OBS (Fig. 1d,e). The SST meridional gradient between the equatorial western Pacific (160°E-180; 5°S-5°N)
373 and the southwestern Pacific (160°E-180; 5°S-15°S), and the SST zonal gradient between the southwestern Pacific
374 (150°E-180; 5°S-15°S) and the south central Pacific (180-210°E; 5°S-15°S) are stronger in CC_COR (0.33°C in
375 CC_COR vs. 0.028°C in CC_OBS and 0.58°C in CC_COR vs. 0.54°C in CC_OBS, respectively). These different SST
376 patterns induce southeasterly wind anomalies (blowing from the southeast Pacific dry zone towards the central and
377 western equatorial Pacific) that are stronger and extend further into the western Pacific in the COR experiment (Fig. 7c
378 vs 7d).

379 To summarize, 1) our model tends to enhance the precipitation response to anthropogenic forcing compared to
380 CMIP5 MMM, especially the southeastern Pacific drying, 2) the historical CMIP5 SST biases damp the intensity of the
381 equatorial rainfall response compared to OBS_{ZM}, largely through thermodynamical processes, and 3) the projected SST
382 correction yields a larger drying of the SPCZ western portion, due to southeasterlies that penetrate further west and
383 north in relation with SST gradient changes.

384

385 4.2 Interannual variability

386 A general assessment of the changes in precipitation interannual variability in the present and future
387 climate is provided in Figure 9. In PD_OBS_{ZM}, large interannual rainfall variations (4 to 8 mm.day⁻¹) are found in a
388 large region extending from 150°E to 150°W and 8°N to 20°S, with a maximum in the SPCZ core region and under the
389 ITCZ (Fig. 9a). This rainfall variability pattern is considerably affected by anthropogenic forcing (Fig. 9b,c), with a
390 near doubling of equatorial rainfall variations. In CC_COR_{ZM}, anthropogenic forcing also affects the South Pacific
391 rainfall variability, with a reduction of rainfall variability south of 5°S. This change in rainfall is likely related to drying
392 of SPCZ in its western part. Similar results are found in the CMIP5_{ZM} experiment (not shown). Finally, the OBS_{ZM} also
393 reveals an enhancement of equatorial inter-annual variability when the historical CMIP5 SST bias is also corrected. In
394 an idealized framework where the SSTs are climatological, Chung et al., (2014) found that global warming enhances
395 the nonlinear rainfall response to ENSO SST anomalies, with an increase of precipitation variability, especially along
396 the equator. Our study confirms these results in a more realistic framework.

397 Enhanced precipitation in the equatorial region however does not appear to be robust for all ENSO
398 phases (Fig. 10abc). The large equatorial rainfall increase under climate change is indeed clear for El Niño events (Fig.
399 10ab) but is considerably weaker during La Niña events, being largely confined to the far western Pacific (Fig. 10c).
400 This is also the case in the OBS_{ZM} and CMIP5_{ZM} experiments (not shown). To further illustrate that, Figure 10d,e
401 provides scatterplots of the rainfall difference between future and present climate for each of 36 simulated austral
402 summers (black dots represent COR_{ZM} experiment, cyan dots OBS_{ZM} experiment, and brown dots CMIP5_{ZM}
403 experiment) as a function of the ENSO state in the two regions drawn on the upper panels of Figure 10: the equatorial
404 region (Fig. 10d) and the South Pacific region (Fig. 10e). The amplitude of the equatorial precipitation increase is
405 considerably larger for an El Niño (up to 2.5 mm/day) than for a La Niña (0mm/day). This illustrates a non-linearity of
406 the precipitation response to climate change during ENSO phases, as also found in the idealized framework of Chung et
407 al., (2014). The South Pacific drying in response to climate change has a weaker non-linear dependency to ENSO (-1
408 mm.day⁻¹ for La Niña vs. 2.5 mm.day⁻¹ for El Niño) than the equatorial region.

409 The changes in rainfall interannual variability near the equator are mostly driven by changes in the SPCZ
410 response to ENSO (Fig. 11). Under present-day conditions, the zonally-oriented SPCZ during extreme El Niño events
411 remains largely located in the Southern Hemisphere (red thick line; Fig. 11a), with the largest northward excursion of
412 12° occurring in PD_OBS_{ZM} during the 1997-1998 event. In CC_COR_{ZM}, the SPCZ shifts right onto the equator
413 between 165°E and 140°W (red dashed line; Fig. 11a), and between 180°E and 140°W in CC_OBS_{ZM} (not shown),
414 resulting in a northward excursion exceeding 15° during extreme El Niño events (1983, 1992, 1998, 2016) (red dashed
415 lines, Fig. 11a). During moderate El Niño events, the SPCZ in both CC_COR_{ZM} (Fig. 11b) and CC_OBS_{ZM} (not shown)
416 experiences a shift of ~3° near the dateline but a far weaker southward excursion in its eastern portion. During la Niña
417 events, the SPCZ position slightly moves southward by 2° in both present and future climates (Fig. 11c).

418 The SST cold bias in CMIP5 models yields weaker ENSO-related SPCZ present-day movements in
419 PD_CMIP5_{ZM} (black dashed line versus black thick line; Fig. 11 d,e,f) than in PD_OBS_{ZM} (black dashed line versus
420 black thick line; Fig. 11 a,b,c). Under climate change, this SPCZ variability increases in CC_CMIP5_{ZM} (STD LatE
421 PD_CMIP5_{ZM} = 3.04°; STD LatE CC_CMIP5_{ZM} = 4.96°), yielding larger northward SPCZ shifts during the future
422 conditions than in the current day, for both strong and moderate El Niños. There are thus larger northward

423 displacements of the SPCZ during El Niños in the future than in the present day, especially in our bias-corrected
424 experiments. But what about the frequency of those zonal events? Cai et al. (2012) demonstrated a near-doubling of
425 SPCZ zonal events frequency in CMIP models under climate change, and linked this increase to a decreased SST
426 meridional gradient south of the equator. To assess the robustness of this result in our experiments, we computed
427 Empirical Orthogonal Functions (EOFs) of precipitation from our regional simulations and identified zonal SPCZ
428 events from the value of the first two principal components as in Cai et al. (2012) (Figure 12). There are four SPCZ
429 zonal events (red circle, Fig. 12 a,b) in both of our present-day simulations, corresponding to 1982/1983, 1991/1992,
430 1997/1998 and 2015/2016 El Nino years. Contrary to what was found by Cai et al. (2012), we find the same number of
431 SPCZ zonal events in the CC_COR, CC_OBS and CC_CMIP5 climate change experiments, i.e. the doubling of zonal
432 SPCZ events under future climate conditions is not robust to a change in model physics (CC_CMIP5), and does not
433 occur either when additionally correcting the historical CMIP5 SST bias or the ΔSST_{GW} bias.

434 To conclude, global warming \square increases the interannual rainfall variability near the equator and decreases it
435 under the SPCZ, especially when the SST projections are corrected. The increased equatorial rainfall variability is
436 mainly related to the non-linear precipitation response to ENSO. Finally, the doubling of SPCZ zonal events discussed
437 by Cai et al. (2012) is not reproduced in our simulations.

438

439 5. Sensitivity tests

440

441 **5.1 Sensitivity of the response to the model physics configuration**

442 It is crucial to assess the robustness of our findings with regards to the regional model physics, since it has a
443 strong impact on how the SPCZ responds to climate change (e.g. Evans et al. 2016). To that end, we ran experiments
444 with a completely different set of physical parameterizations that also yields a realistic representation of the South
445 Pacific region climate (BMJ configuration; Jourdain et al., 2011; Jullien et al., 2014).

446 The climatological rainfall pattern (blue circles on Fig. 4) and those induced by ENSO variations (not shown)
447 are also reasonably well simulated by that model configuration. PD_OBS_{BMJ} yields a realistic rainfall pattern (Fig. 4d)
448 and latitude of the SPCZ eastern portion (Fig. 4b). However, it displays a 3° southward bias of the SPCZ western
449 portion (Fig. 4c), resulting in a 25% underestimation of the SPCZ slope (Fig. 4a). In addition, this simulation exhibits a
450 considerably reduced dry equatorial bias relative to PD_OBS_{ZM} (not shown), but a southward bias of the SPCZ (Fig.
451 4b,c). Despite the differences in their present-day SPCZ climatology, the ZM and BMJ configurations yield a similar
452 response to climate change, with a large rainfall increase in the equatorial Pacific in both COR and OBS (Fig 13a,b vs
453 Fig 7c,d). The COR experiment also produces a large drying of the southwest and south-central Pacific (Fig. 13a vs Fig.
454 7d). The main difference is a larger ITCZ drying in the BMJ configuration (~20% in BMJ vs ~10% in ZM), and overall
455 more intense future changes. These anomalous rainfall patterns result in a southward shift (~3°) of the climatological
456 SPCZ position from present day to future (black thick line vs black dashed line; Fig 13a,b) that is similar to that in the
457 ZM configuration. In addition, as in ZM, the variability of the SPCZ eastern portion and its northward shift also
458 increase in the future for our BMJ experiments (not shown; STD LatE PD_OBS_{BMJ} = 5.3°; STD LatE CC_OBS_{BMJ} = 8°;
459 and STD LatE CC_COR_{BMJ} = 7.8° to be compared to : STD LatE PD_OBS_{ZM} = 4.24°; STD LatE CC_OBS_{ZM} = 6.21°;
460 and STD LatE CC_COR_{ZM} = 6.18°). Finally, as in BMJ, correcting the CMIP5 mean state cold bias yields a larger

461 precipitation increase in the tropical region and correcting the projected SST change yields a larger drying of the SPCZ
462 western portion (Fig. 13). Finally, as in ZM, the number of SPCZ zonal events does not increase under climate change
463 in any of our BMJ simulations.

464 Overall, the main conclusions derived from the ZM configuration remain valid under a very different set of
465 model parameterisations.

466

467 **5.2 Comparison with the CMIP models response**

468 A summary of the mechanisms responsible for the response of precipitation to climate change in the
469 equatorial and South Pacific regions in CMIP5 models and our regional experiments is finally provided in Figure 14,
470 which shows scatterplots of the dynamic and thermodynamic components of the moisture transport vs. the total
471 transport, in key regions.

472 In the equatorial region, dynamical changes (i.e. increased wind convergence) are entirely responsible for
473 the projected increased rainfall in CMIP5 models (black and gray dots in Fig. 14a,b). The same processes control the
474 south central Pacific drying (Fig. 14c,d). In the southwest Pacific (Fig. 14e,f), there tends to be a cancellation between
475 dynamical processes that tend to induce a drying through increased wind divergence and thermodynamical processes
476 that increase moisture convergence and rainfall in the CMIP5 models. Depending on the respective contributions from
477 these two processes, CMIP5 models either display increased precipitation in the southwest Pacific, resulting in a large
478 spread across models and a nearly-zero ensemble mean projected rainfall change.

479 The CMIP_{ZM} and CMIP_{BMJ} experiments lie at the upper end of the CMIP5 models distribution in both the
480 western equatorial Pacific (blue symbols in Fig. 14a,b) and the south central Pacific (blue symbols in Fig. 14c,d), the
481 dynamical component largely explaining the total changes. Correcting historical CMIP5 SST and Δ SST_{GW} biases
482 generally lead to an increase of the dynamical response and hence of the rainfall anomaly, this increase being
483 particularly prominent in the western equatorial Pacific for BMJ. In southwestern Pacific, the drying by dynamical
484 processes is about $\sim 20\%$ larger in the experiment using Δ SST_{COR_{GW}} (Fig. 14e). On the other hand, the
485 thermodynamical effects are very similar in all our experiments (Fig. 14f), and have similar amplitudes to those in
486 CMIP5 models. This yields a larger southwest Pacific drying in the experiments using Δ SST_{COR_{GW}}. In other words,
487 the cold tongue bias induces a bias in projected SST that yield an underestimated change in circulation and associated
488 moisture divergence, and thus an underestimated southwest Pacific drying.

489

490 **6. Summary and discussion**

491

492 **6.1 Summary**

493 This study investigates the influence of anthropogenic forcing on the SPCZ mean state and interannual
494 variability. A present-day simulation (PD_OBS) with a regional atmospheric model captures the main observed SPCZ
495 characteristics, with a similar degree of realism to that in the NCEP2 reanalysis. This includes the SPCZ mean state but
496 also its northward/southward displacements associated with ENSO, including the SPCZ zonal events during extreme El
497 Niño events. This accurate present-day simulation of the SPCZ justifies the use of our regional model configuration to

498 assess the SPCZ response to climate change. In contrast, a twin simulation with CMIP5 climatological SST (and the
499 trademark cold tongue bias in CMIP models) yields to a similar SPCZ bias as in the CMIP5 models, with a too zonal
500 SPCZ.

501 We further performed three simulations to estimate the influence of climate change: CC_CMIP5, CC_OBS,
502 and CC_COR. In CC_CMIP5 and CC_OBS, the boundary conditions from the PD_CMIP5 and PD_OBS experiments
503 are corrected using projected changes from CMIP5. Comparing those two experiments allows for assessing the effect of
504 the SST bias in CMIP5 models. The CC_COR simulation is similar to CC_OBS, but the projected SST change from the
505 CMIP5 ensemble mean $\Delta\text{SST}_{\text{GW}}$ is additionally corrected using an “emergent constraint” method, based on Li et al.
506 (2016). Comparing this last experiment to CC_OBS allows estimating the effect of the $\Delta\text{SST}_{\text{GW}}$ bias on the SPCZ
507 changes.

508 To assess the sensitivity to model physics, we compared our CC_CMIP5_{ZM} simulation to the CMIP5 multi-
509 model mean. The SPCZ response to climate change is slightly more intense in CC_CMIP5_{ZM} than in CMIP5 MMM,
510 with a slightly stronger increase of precipitation in the equatorial region ($\sim 2.2\text{mm}\cdot\text{day}^{-1}$ in CMIP5 MMM and
511 $2.5\text{mm}\cdot\text{day}^{-1}$ CMIP5_{ZM} experiment), and a stronger drying in the southeastern Pacific ($-1\text{mm}\cdot\text{day}^{-1}$ compared to -
512 $0.4\text{mm}\cdot\text{day}^{-1}$), leading to a shrinkage of the SPCZ. Our model setup hence reproduces qualitatively the rainfall changes
513 in CMIP5 MMM, but magnifies the amplitude of the precipitation response to anthropogenic forcing.

514 To evaluate the impact of historical CMIP5 SST biases, we compared the CC_OBS_{ZM} simulation to
515 CC_CMIP5_{ZM}. Correcting the CMIP5 historical SST bias leads to a much stronger equatorial Pacific rainfall increase in
516 response to climate change ($\sim 3.5\text{mm}\cdot\text{day}^{-1}$ vs. $2.5\text{mm}\cdot\text{day}^{-1}$ in the CMIP5_{ZM} experiment), mainly due to a weaker
517 negative feedback of the thermodynamic component of moisture transport in CC_OBS_{ZM}. The drying in the southwest
518 Pacific is hardly influenced by the CMIP5 SST bias ($0.9\text{mm}\cdot\text{day}^{-1}$ rainfall decrease in CC_OBS_{ZM} vs. $1\text{mm}\cdot\text{day}^{-1}$ in
519 CMIP5_{ZM}). These changes are associated with a $\sim 4^\circ$ southward shift of the mean SPCZ location west of 170°W .

520 Finally, an original aspect of our study is our evaluation of the impact of plausible biases in the projected SST
521 $\Delta\text{SST}_{\text{GW}}$, by comparing CC_COR_{ZM} with CC_OBS_{ZM}. In CC_COR_{ZM} the equatorial rainfall increase is weaker than in
522 CC_OBS ($3\text{mm}\cdot\text{day}^{-1}$ vs. $3.5\text{mm}\cdot\text{day}^{-1}$), while the southeastern Pacific becomes much drier ($-1.9\text{mm}\cdot\text{day}^{-1}$ vs. -
523 $0.9\text{mm}\cdot\text{day}^{-1}$), and the drying extends further west to 150°E . Correcting the projected SSTs changes the SST gradients
524 in the southwestern Pacific, leading to southeasterlies that penetrate further west and north. This induces more moisture
525 divergence through dynamical processes, and hence a larger drying.

526 All the climate change simulations display enhanced rainfall variability in the equatorial region, but this
527 change is not the same across ENSO phases. The precipitation increase is strongest during extreme El Niño years,
528 illustrating the non-linearity of the precipitation response to climate change during ENSO phases, as also found in the
529 idealized framework of Chung et al., (2014). Finally, our analyses show that the number of SPCZ zonal events under
530 climate change conditions remains similar to the present-day conditions in all our simulations, in contrast with the result
531 of Cai et al. (2012) in CMIP3-5 models, questioning the robustness of this result.

532 Last but not least, additional regional simulations using a completely different set of physical
533 parameterizations give qualitatively similar patterns for the future SPCZ mean change and how these are influenced by
534 CMIP biases, underlining the robustness of our results.

535

536 **6.2 Discussion**

537 Several studies have already investigated the impact of climate change onto the mean climate within the
538 SPCZ region, either by directly comparing global climate models (Brown et al. 2012, 2013) or by comparing
539 atmospheric experiments forced with and without climate change boundary conditions anomalies (Widlansky et al. 2013;
540 Brown et al. 2015; Evans et al. 2016). A direct analysis of CMIP5 models outputs reveals that the climate change
541 warming pattern results in a robust precipitation increase in the equatorial Pacific (e.g. Watanabe et al. 2014; Li et al.,
542 2016), a weak drying of the southeast SPCZ margin in the CMIP5 models but no consistent shift in climatological
543 SPCZ position (Brown et al. 2012, 2013). Despite stronger rainfall change in our sets of experiments than in CMIP5
544 MMM, we find that the increased equatorial rainfall and dryer eastern SPCZ are robust features. On the other hand,
545 CMIP5 MMM, our CC_OBS_{ZM} and CC_CMIP5_{ZM} simulations show an insignificant drying in the western SPCZ part,
546 while, by contrast, our CC_COR_{ZM} simulation yields a strong significant drying (-20%) in this region.

547 Our results in the OBS experiment are in qualitative agreement with the atmospheric experiments of
548 Widlansky et al. (2013). Indeed their climate-change bias-corrected simulations pointed toward a significant drying of
549 the SPCZ and wetter equatorial Pacific conditions during austral summer. In contrast, results from the present-day SST
550 bias-corrected experiments from Brown et al. (2015) still indicate a future wettening of the southwest Pacific. We
551 however suggest our results to be more reliable than Brown et al. (2015), due to a more realistic experimental strategy.
552 Indeed, Brown et al. (2015) used an idealized framework by forcing their atmospheric model with annually averaged
553 SST, which may have impaired the realism of their projected rainfall response. As acknowledged by these authors, their
554 objective was indeed not to provide a reliable projection of future rainfall but rather to demonstrate the sensitivity of the
555 rainfall response to the underlying SST patterns. Evans et al. (2016) also used a present-day SST bias-corrected
556 strategy, with several regional climate models forced by the same CMIP global warming anomalous pattern. While they
557 showed that their projected rainfall response was not very sensitive to the choice of the CMIP model used to force their
558 regional model boundaries, their projections in the southwest Pacific region were more sensitive to the
559 parameterizations used in their regional model. Despite this sensitivity, they indicated that the majority (7 out of 9) of
560 their configurations projected a significant drying of the southwest Pacific, in line with Widlansky et al. (2013) and the
561 present study.

562 In our COR experiment, we have tested the sensitivity of the rainfall response to the characteristics of
563 the projected SST pattern. Brown et al. (2015) suggested that the amplitude of the rainfall response could be highly
564 sensitive to the pattern of SST warming. Huang and Ying (2015) and Li et al. (2016) confirmed that the well-known
565 cold tongue bias in most CMIP models could lead to an underestimation in the east-minus-west gradient of the
566 projected sea surface warming along the equator. In contrast, Widlansky et al. (2013) argued that the SST warming
567 pattern was independent of the climatological SST bias by comparing the SST warming pattern from a climate model
568 run with and without flux adjustment strategy. In line with the previous statistical studies (Huang and Ying, 2015; Li et
569 al. 2016), we find that biases in the pattern of CMIP5 sea surface warming considerably influence the future changes of
570 regional precipitations. Future precipitation changes in the western tropical Pacific are indeed driven by two main
571 processes, which respective contributions are strongly influenced by the underlying background SSTs: a “wet gets
572 wetter” thermodynamic response and a “warmest gets wetter” dynamical response due to the overall tropical circulation
573 slowdown (Widlansky et al. 2013).

574 Despite the absence of consensus on how SST variability associated with ENSO may change in the
575 future (e.g. Cai et al., 2015), CMIP3-5 models project a future increase in the frequency of zonal SPCZ events, which

576 has been attributed to mean changes in meridional SST gradient between the equator and the South Pacific (Cai et al.
577 2012). In line with recent studies using similar experimental strategy (Chung and Power 2016; Evans et al. 2016), our
578 results suggest that the frequency of zonal SPCZ events may not increase in the future, in apparent contradiction with
579 Cai et al. (2012) results. Several reasons could explain this apparent discrepancy. First, while CMIP5 mean present-day
580 biases have been corrected in all the later studies, the common CMIP model biases in Cai et al. (2012), including the
581 cold tongue bias, may largely responsible for the unrealistic increase of the zonal SPCZ events frequency in CMIP
582 simulations. Second, the present and future SST variability is kept constant in our simulations and those of Chung and
583 Power (2016): it may well be that, in contradiction to what has been initially proposed by Cai et al. (2012), the mean
584 SST changes may not be the dominant mechanism explaining the zonal SPCZ events frequency increase in their study.
585 Rather, consistent changes in SST interannual variability in CMIP models may be partly responsible for this increase.
586 Other hypotheses explaining this discrepancy include as well the diversity of the $\Delta\text{SST}_{\text{GW}}$ patterns across the CMIP5
587 models, not taken into account in our multi-model mean of $\Delta\text{SST}_{\text{GW}}$ CMIP5 pattern or the lack of atmosphere–ocean
588 feedbacks in these forced atmospheric simulations compared to CMIP models.

589 In any case, this study highlights the importance of a realistic simulation of the tropical Pacific mean state for
590 improving regional climate projections. Tropical Pacific SST biases, including the prominent cold tongue bias,
591 influence the projected rainfall response directly but also indirectly by altering the projected SST changes. While a
592 long-term objective is clearly to reduce long-standing robust SST biases in the tropical Pacific in CMIP models, this
593 study also advocates for a shorter-term strategy, by using standalone atmospheric experiments where present-day and
594 future SST projection are bias-corrected. This study also highlights the need to perform this kind of strategy with
595 different models: while the bias correction approach used in this bias-corrected forced atmospheric study is likely to
596 improve the reliability of future projection, atmosphere-only present-day simulations still exhibit significant SPCZ
597 biases (Lintner et al. 2016) that may still impair the reliability of the projected changes, as illustrated by the significant
598 quantitative differences in the projected rainfall changes in our two model configurations (Fig. 7d vs. Fig. 13a). It is also
599 possible that the reanalysis products used to build the boundary conditions influence the SPCZ response to
600 anthropogenic forcing, since the present characteristics of the SPCZ are more or less captured according to the
601 reanalysis used (eg, Niznik et al. Lintner 2013). Therefore a sensitivity experiment to this parameter will be tested.
602 Finally, the sensitivity of the projected changes to the statistical correction method used to correct the projected SST
603 changes should also be tested: results with the simple correction strategy proposed by Li et al. (2016) should be
604 compared with more complex strategies, such as the one proposed by Huang and Ying (2015) which extracts the
605 leading bias modes of historical SST CMIP5 simulations and then projecting the tropical Pacific SST changes onto
606 these historical bias modes.

607

References :

Alexander, L.V., Zhang, X., Peterson, T.C., Caesar, J., Gleason, B., Klein Tank, A.M.G., Haylock, M., Collins, D.,
Trewin, B., Rahimzadeh, F., et al. (2006). Global observed changes in daily climate extremes of temperature and
precipitation. *J. Geophys. Res.* *111*.

Bellenger, H., Guilyardi, E., Leloup, J., Lengaigne, M., and Vialard, J. (2014). ENSO representation in climate models:
from CMIP3 to CMIP5. *Clim. Dyn.* *42*, 1999–2018.

Bracegirdle, T.J., and Stephenson, D.B. (2013). On the Robustness of Emergent Constraints Used in Multimodel
Climate Change Projections of Arctic Warming. *J. Clim.* *26*, 669–678.

- Bretherton, C.S., and Park, S. (2009). A New Moist Turbulence Parameterization in the Community Atmosphere Model. *J. Clim.* *22*, 3422–3448.
- Brown, J.N., Matear, R.J., Brown, J.R., and Katzfey, J. (2015). Precipitation projections in the tropical Pacific are sensitive to different types of SST bias adjustment: precipitation and SST bias adjustment. *Geophys. Res. Lett.* *42*, 10,856–10,866.
- Brown, J.R., Power, S.B., Delage, F.P., Colman, R.A., Moise, A.F., and Murphy, B.F. (2011). Evaluation of the South Pacific Convergence Zone in IPCC AR4 Climate Model Simulations of the Twentieth Century. *J. Clim.* *24*, 1565–1582.
- Brown, J.R., Moise, A.F., and Delage, F.P. (2012). Changes in the South Pacific Convergence Zone in IPCC AR4 future climate projections. *Clim. Dyn.* *39*, 1–19.
- Brown, J.R., Moise, A.F., and Colman, R.A. (2013). The South Pacific Convergence Zone in CMIP5 simulations of historical and future climate. *Clim. Dyn.* *41*, 2179–2197.
- Cai, W., Lengaigne, M., Borlace, S., Collins, M., Cowan, T., McPhaden, M.J., Timmermann, A., Power, S., Brown, J., Menkes, C., et al. (2012). More extreme swings of the South Pacific convergence zone due to greenhouse warming. *Nature* *488*, 365–369.
- Cai, W., Borlace, S., Lengaigne, M., van Rensch, P., Collins, M., Vecchi, G., Timmermann, A., Santoso, A., McPhaden, M.J., Wu, L., et al. (2014). Increasing frequency of extreme El Niño events due to greenhouse warming. *Nat. Clim. Change* *4*, 111–116.
- Cai, W., Santoso, A., Wang, G., Yeh, S.-W., An, S.-I., Cobb, K.M., Collins, M., Guilyardi, E., Jin, F.-F., Kug, J.-S., et al. (2015). ENSO and greenhouse warming. *Nat. Clim. Change* *5*, 849–859.
- Chen, F., and Dudhia, J. (2001). Coupling an Advanced Land Surface–Hydrology Model with the Penn State–NCAR MM5 Modeling System. Part I: Model Implementation and Sensitivity. *Mon. Weather Rev.* *129*, 569–585.
- Chung, C.T.Y., and Power, S.B. (2016). Modelled impact of global warming on ENSO-driven precipitation changes in the tropical Pacific. *Clim. Dyn.* *47*, 1303–1323.
- Chung, C.T.Y., Power, S.B., Arblaster, J.M., Rashid, H.A., and Roff, G.L. (2014). Nonlinear precipitation response to El Niño and global warming in the Indo-Pacific. *Clim. Dyn.* *42*, 1837–1856.
- Collins, M., An, S.-I., Cai, W., Ganachaud, A., Guilyardi, E., Jin, F.-F., Jochum, M., Lengaigne, M., Power, S., Timmermann, A., et al. (2010). The impact of global warming on the tropical Pacific Ocean and El Niño. *Nat. Geosci.* *3*, 391–397.
- Collins, W.D., Rasch, P.J., Boville, B.A., Hack, J.J., McCaa, J.R., Williamson, D.L., Kiehl, J.T., Briegleb, B., Bitz, C., Lin, S.-J., et al. (2004). Description of the NCAR Community Atmosphere Model (CAM 3.0). 226.
- Deser, C., Phillips, A.S., and Alexander, M.A. (2010). Twentieth century tropical sea surface temperature trends revisited: twentieth century tropical SST trends. *Geophys. Res. Lett.* *37*, n/a-n/a.
- Evans, J.P., Bormann, K., Katzfey, J., Dean, S., and Arritt, R. (2016). Regional climate model projections of the South Pacific Convergence Zone. *Clim. Dyn.* *47*, 817–829.
- Folland, C.K. (2002). Relative influences of the Interdecadal Pacific Oscillation and ENSO on the South Pacific Convergence Zone. *Geophys. Res. Lett.* *29*.
- Gouriou, Y., and Delcroix, T. (2002). Seasonal and ENSO variations of sea surface salinity and temperature in the South Pacific Convergence Zone during 1976–2000: sea surface salinity and temperature in the SPCZ. *J. Geophys. Res. Oceans* *107*, SRF 12-1–SRF 12-14.
- Grose, M.R., Brown, J.N., Narsey, S., Brown, J.R., Murphy, B.F., Langlais, C., Gupta, A.S., Moise, A.F., and Irving, D.B. (2014). Assessment of the CMIP5 global climate model simulations of the western tropical Pacific climate system

- and comparison to CMIP3: assesment of CMIP5 climate models for the western tropical pacific. *Int. J. Climatol.* *34*, 3382–3399.
- Held, I.M., and Soden, B.J. (2006). Robust Responses of the Hydrological Cycle to Global Warming. *J. Clim.* *19*, 5686–5699.
- Hong, S.-Y., Dudhia, J., and Chen, S.-H. (2004). A Revised Approach to Ice Microphysical Processes for the Bulk Parameterization of Clouds and Precipitation. *Mon. Weather Rev.* *132*, 103–120.
- Huang, P., and Ying, J. (2015). A Multimodel Ensemble Pattern Regression Method to Correct the Tropical Pacific SST Change Patterns under Global Warming. *J. Clim.* *28*, 4706–4723.
- Huang, P., Xie, S.-P., Hu, K., Huang, G., and Huang, R. (2013). Patterns of the seasonal response of tropical rainfall to global warming. *Nat. Geosci.* *6*, 357–361.
- Iacono, M.J., Delamere, J.S., Mlawer, E.J., Shephard, M.W., Clough, S.A., and Collins, W.D. (2008). Radiative forcing by long-lived greenhouse gases: Calculations with the AER radiative transfer models. *J. Geophys. Res.* *113*.
- Irving, D., Perkins, S., Brown, J., Sen Gupta, A., Moise, A., Murphy, B., Muir, L., Colman, R., Power, S., Delage, F., et al. (2011). Evaluating global climate models for the Pacific island region. *Clim. Res.* *49*, 169–187.
- Janjić, Z.I. (1994). The Step-Mountain Eta Coordinate Model: Further Developments of the Convection, Viscous Sublayer, and Turbulence Closure Schemes. *Mon. Weather Rev.* *122*, 927–945.
- Jourdain, N.C., Marchesiello, P., Menkes, C.E., Lefèvre, J., Vincent, E.M., Lengaigne, M., and Chauvin, F. (2011). Mesoscale Simulation of Tropical Cyclones in the South Pacific: Climatology and Interannual Variability. *J. Clim.* *24*, 3–25.
- Juillet-Leclerc, A., Thiria, S., Naveau, P., Delcroix, T., Le Bec, N., Blamart, D., and Corrège, T. (2006). SPCZ migration and ENSO events during the 20th century as revealed by climate proxies from a Fiji coral. *Geophys. Res. Lett.* *33*.
- Jullien, S., Marchesiello, P., Menkes, C.E., Lefèvre, J., Jourdain, N.C., Samson, G., and Lengaigne, M. (2014). Ocean feedback to tropical cyclones: climatology and processes. *Clim. Dyn.* *43*, 2831–2854.
- Kanamitsu, M., Ebisuzaki, W., Woollen, J., Yang, S.-K., Hnilo, J.J., Fiorino, M., and Potter, G.L. (2002). NCEP–DOE AMIP-II Reanalysis (R-2). *Bull. Am. Meteorol. Soc.* *83*, 1631–1643.
- Kidwell, A., Lee, T., Jo, Y.-H., and Yan, X.-H. (2016). Characterization of the Variability of the South Pacific Convergence Zone Using Satellite and Reanalysis Wind Products. *J. Clim.* *29*, 1717–1732.
- Kiladis, G.N., von Storch, H., and Loon, H. (1989). Origin of the South Pacific Convergence Zone. *J. Clim.* *2*, 1185–1195.
- Kirono, D.G.C., Kent, D.M., Hennessy, K.J., and Mpelasoka, F. (2011). Characteristics of Australian droughts under enhanced greenhouse conditions: Results from 14 global climate models. *J. Arid Environ.* *75*, 566–575.
- Knutson, T.R., Sirutis, J.J., Garner, S.T., Vecchi, G.A., and Held, I.M. (2008). Simulated reduction in Atlantic hurricane frequency under twenty-first-century warming conditions. *Nat. Geosci.* *1*, 359–364.
- Lefèvre, J., Menkes, C., Bani, P., Marchesiello, P., Curci, G., Grell, G.A., and Frouin, R. (2016). Distribution of sulfur aerosol precursors in the SPCZ released by continuous volcanic degassing at Ambrym, Vanuatu. *J. Volcanol. Geotherm. Res.* *322*, 76–104.
- Li, G., and Xie, S.-P. (2014). Tropical Biases in CMIP5 Multimodel Ensemble: The Excessive Equatorial Pacific Cold Tongue and Double ITCZ Problems*. *J. Clim.* *27*, 1765–1780.

- Li, G., Du, Y., Xu, H., and Ren, B. (2015). An Intermodel Approach to Identify the Source of Excessive Equatorial Pacific Cold Tongue in CMIP5 Models and Uncertainty in Observational Datasets. *J. Clim.* *28*, 7630–7640.
- Li, G., Xie, S.-P., Du, Y., and Luo, Y. (2016). Effects of excessive equatorial cold tongue bias on the projections of tropical Pacific climate change. Part I: the warming pattern in CMIP5 multi-model ensemble. *Clim. Dyn.* *47*, 3817–3831.
- Lin, Y.-L., Farley, R.D., and Orville, H.D. (1983). Bulk Parameterization of the Snow Field in a Cloud Model. *J. Clim. Appl. Meteorol.* *22*, 1065–1092.
- Matsueda, M., and Palmer, T.N. (2011). Accuracy of climate change predictions using high resolution simulations as surrogates of truth: accuracy of climate change predictions. *Geophys. Res. Lett.* *38*, n/a-n/a.
- McGree, S., Whan, K., Jones, D., Alexander, L.V., Imielska, A., Diamond, H., Ene, E., Finaulahi, S., Inape, K., Jacklick, L., et al. (2014). An updated assessment of trends and variability in total and extreme rainfall in the western Pacific: trends and variability in extreme rainfall in the western Pacific. *Int. J. Climatol.* *34*, 2775–2791.
- Mechoso, C.R., Robertson, A.W., Barth, N., Davey, M.K., Delecluse, P., Gent, P.R., Ineson, S., Kirtman, B., Latif, M., Treut, H.L., et al. (1995). The Seasonal Cycle over the Tropical Pacific in Coupled Ocean–Atmosphere General Circulation Models. *Mon. Weather Rev.* *123*, 2825–2838.
- Meehl, G.A., Covey, C., Taylor, K.E., Delworth, T., Stouffer, R.J., Latif, M., McAvaney, B., and Mitchell, J.F.B. (2007). THE WCRP CMIP3 Multimodel Dataset: A New Era in Climate Change Research. *Bull. Am. Meteorol. Soc.* *88*, 1383–1394.
- Meng, Q., Latif, M., Park, W., Keenlyside, N.S., Semenov, V.A., and Martin, T. (2012). Twentieth century Walker Circulation change: data analysis and model experiments. *Clim. Dyn.* *38*, 1757–1773.
- Murphy, B.F., Ye, H., and Delage, F. (2015). Impacts of variations in the strength and structure of El Niño events on Pacific rainfall in CMIP5 models. *Clim. Dyn.* *44*, 3171–3186.
- Nguyen, K.C., Katzfey, J.J., and McGregor, J.L. (2012). Global 60 km simulations with CCAM: evaluation over the tropics. *Clim. Dyn.* *39*, 637–654.
- Niznik, M.J., Lintner, B.R., Matthews, A.J., and Widlansky, M.J. (2015). The Role of Tropical–Extratropical Interaction and Synoptic Variability in Maintaining the South Pacific Convergence Zone in CMIP5 Models. *J. Clim.* *28*, 3353–3374.
- Noh, Y., Cheon, W.G., Hong, S.Y., and Raasch, S. (2003). Improvement of the K-profile model for the planetary boundary layer based on large eddy simulation data. *Bound.-Layer Meteorol.* *107*, 401–427.
- Parvathi, V., Suresh, I., Lengaigne, M., Izumo, T., and Vialard, J. (2017). Robust Projected Weakening of Winter Monsoon Winds Over the Arabian Sea Under Climate Change. *Geophys. Res. Lett.* *44*, 9833–9843.
- Perkins, S.E. (2011). Biases and Model Agreement in Projections of Climate Extremes over the Tropical Pacific. *Earth Interact.* *15*, 1–36.
- Power, S., Delage, F., Chung, C., Kociuba, G., and Keay, K. (2013). Robust twenty-first-century projections of El Niño and related precipitation variability. *Nature* *502*, 541–545.
- Skamarock, W.C., and Klemp, J.B. (2008). A time-split nonhydrostatic atmospheric model for weather research and forecasting applications. *J. Comput. Phys.* *227*, 3465–3485.
- Solomon, A., and Newman, M. (2012). Reconciling disparate twentieth-century Indo-Pacific ocean temperature trends in the instrumental record. *Nat. Clim. Change* *2*, 691–699.
- Taylor, K.E., Stouffer, R.J., and Meehl, G.A. (2012). An Overview of CMIP5 and the Experiment Design. *Bull. Am. Meteorol. Soc.* *93*, 485–498.

Tokinaga, H., Xie, S.-P., Deser, C., Kosaka, Y., and Okumura, Y.M. (2012). Slowdown of the Walker circulation driven by tropical Indo-Pacific warming. *Nature* 491, 439–443.

Trenberth, K.E. (1976). Spatial and temporal variations of the Southern Oscillation. *Q. J. R. Meteorol. Soc.* 102, 639–653.

van Vuuren, D. P., and Coauthors, 2011: The representative concentration pathways: An overview. *Climatic Change*, 109, 5–31, <https://doi.org/10.1007/s10584-011-0148-z>.

Vecchi, G.A., Clement, A., and Soden, B.J. (2008). Examining the tropical Pacific's response to global warming. *Eos Trans. Am. Geophys. Union* 89, 81–83.

Vincent, D.G. (1994). The South Pacific Convergence Zone (SPCZ): A Review. *Mon. Weather Rev.* 122, 1949–1970.

Vincent, E.M., Lengaigne, M., Menkes, C.E., Jourdain, N.C., Marchesiello, P., and Madec, G. (2011). Interannual variability of the South Pacific Convergence Zone and implications for tropical cyclone genesis. *Clim. Dyn.* 36, 1881–1896.

Walsh, K. (2015). Fine resolution simulations of the effect of climate change on tropical cyclones in the South Pacific. *Clim. Dyn.* 45, 2619–2631.

Watanabe, M., Kamae, Y., and Kimoto, M. (2014). Robust increase of the equatorial Pacific rainfall and its variability in a warmed climate: increasing pacific rainfall variability. *Geophys. Res. Lett.* 41, 3227–3232.

Widlansky, M.J., Timmermann, A., Stein, K., McGregor, S., Schneider, N., England, M.H., Lengaigne, M., and Cai, W. (2013). Changes in South Pacific rainfall bands in a warming climate. *Nat. Clim. Change* 3, 417–423.

Xie, S.-P., Deser, C., Vecchi, G.A., Collins, M., Delworth, T.L., Hall, A., Hawkins, E., Johnson, N.C., Cassou, C., Giannini, A., et al. (2015). Towards predictive understanding of regional climate change. *Nat. Clim. Change* 5, 921–930.

Zhang, G.J., and McFarlane, N.A. (1995). Sensitivity of climate simulations to the parameterization of cumulus convection in the Canadian climate centre general circulation model. *Atmosphere-Ocean* 33, 407–446.

Zheng, Y., Lin, J.-L., and Shinoda, T. (2012). The equatorial Pacific cold tongue simulated by IPCC AR4 coupled GCMs: Upper ocean heat budget and feedback analysis: The pacific cold tongue bias analysis. *J. Geophys. Res. Oceans* 117, n/a-n/a.

608
609