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Role of Tropical SST Variability on the Formation of Subtropical Dipoles

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3 **Role of tropical SST variability on the**

4 **formation of subtropical dipoles**

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21

22 **Abstract**

23 Interannual variations of Sea Surface Temperature (SST) in the midlatitudes of the
24 Southern Hemisphere play an important role in the rainfall variability over the surrounding
25 countries by modulating synoptic-scale atmospheric disturbances. These are frequently
26 associated with a northeast-southwest oriented dipole of positive and negative SST anomalies
27 in each oceanic basin, referred to as a subtropical dipole. This study investigates the role of
28 tropical SST variability on the generation of subtropical dipoles by conducting SST-nudging
29 experiments using a coupled general circulation model. In the experiments where the
30 simulated SST in each tropical basin is nudged to the climatology of the observed SST, the
31 subtropical dipoles tend to occur as frequently as the case in which the simulated SST is
32 allowed to freely interact with the atmosphere. It is found that without the tropical SST
33 variability, the zonally elongated atmospheric mode in the mid-high latitudes, called the
34 Antarctic Oscillation (AAO), becomes dominant and the stationary Rossby waves related to
35 the AAO induce the SLP anomalies in the midlatitudes, which, in turn, generate the
36 subtropical dipoles. These results suggest that the tropical SST variability may not be
37 necessary for generating the subtropical dipoles, and hence provide a useful insight into the
38 important role of the AAO in the midlatitude climate variability.

39

40 **1. Introduction**

41 More than 80 % of the surface area in the Southern Hemisphere is covered by ocean,
42 and the ocean plays a key role in the rainfall over land as moisture sources. Many countries
43 are located in the low-mid latitudes (0°-50°S), and highly dependent on rain-fed agriculture.
44 In these regions, rainfall variability is strongly linked with climate variability involving the
45 low-mid latitude oceans, and greatly influences the crop yields, infrastructure, and drinking
46 water (e.g. Jury 2002). A large number of studies on the rainfall variability discussed the
47 impacts of the tropical climate variability such as El Niño-Southern Oscillation (ENSO)
48 (Chiew et al. 1998, Grimm et al. 2000, Richard et al. 2000), but recently, much focus has been
49 placed on links with the midlatitude ocean variability (Reason 2001, Robertson et al. 2003,
50 Ummenhofer et al. 2009, Cai and Cowan 2013).

51 One of the major oceanic factors for the rainfall variability is Sea Surface Temperature
52 (SST). The SST in the midlatitudes seasonally changes due mostly to solar radiation, and
53 reaches its peak during austral summer (December-February). The SST also undergoes a
54 year-to-year variability, which becomes large during austral summer because of a shallow
55 mixed layer and high solar insolation. The interannual SST variation in the midlatitudes is
56 frequently associated with a northeast-southwest oriented dipole of SST anomalies in each
57 oceanic basin, referred to as a subtropical dipole. Behera and Yamagata (2001) first identified
58 this phenomenon in the southern Indian Ocean. Since then, several studies have shown the
59 existence of subtropical dipoles in the South Atlantic (Fauchereau et al. 2003) and South
60 Pacific (Morioka et al. 2013b).

61 The generation mechanism of the subtropical dipoles has been widely discussed in the

62 literature (Suzuki et al. 2004, Hermes and Reason 2005, Chiodi and Harrison 2007, Morioka
63 et al. 2010, 2011, 2012). At the beginning, variations in the subtropical high during late
64 austral spring generate a northeast-southwest oriented dipole of latent heat flux anomalies.
65 This induces dipole anomalies of mixed-layer thickness, and modulates warming of the mixed
66 layer by shortwave radiation. This results in the dipole SST anomalies associated with the
67 subtropical dipole during austral summer. Therefore, the variation in the subtropical high is a
68 key factor for generating the subtropical dipole.

69 However, possible sources of variability in the subtropical high are not fully understood.
70 Huang and Shukla (2006, 2008) and Morioka et al. (2013a) conducted several Coupled
71 General Circulation Model (CGCM) experiments and suggested that not only ENSO but the
72 atmospheric variability in the mid-high latitudes, called the Antarctic Oscillation (AAO;
73 Thompson and Wallace 2000), may be important for inducing the variability of the Mascarene
74 High, hence generating the Indian Ocean Subtropical Dipole (IOSD). Also, in the CGCM
75 experiments by Morioka et al. (2013a), the IOSD occurs even without the SST variability
76 outside the southern Indian Ocean, suggesting a potential role of the local air-sea feedback
77 during the development of the IOSD, which was discussed by Behera and Yamagata (2001).
78 However, the CGCM used in their studies did not include a sea-ice model and there may be
79 some uncertainties in their simulation for the mid-high latitudes climate. In fact, a large
80 number of studies suggested the importance of sea-ice variability in the high-latitude
81 atmospheric variability through changes in the surface albedo and the air temperature
82 (Simmonds and Budd 1991, Hudson and Hewitson 2001, Raphael 2003).

83 As the dominant atmospheric variability in the Southern Hemisphere, the AAO plays a

84 key role in the SST, mixed-layer, and sea-ice variability in the extratropics (Sen Gupta and
85 England 2006, Ciasto and Thompson 2008, Screen 2010). Several studies discussed that the
86 AAO is an internal atmospheric variability sustained by the positive feedback between the
87 zonal-mean flow and eddies (Lorenz and Hartmann 2001, Eichelberger and Hartmann 2007).
88 The positive (negative) AAO is driven by the eddy momentum flux convergence/divergence
89 associated with a strong (weak) anticyclonic Rossby wave breaking on the equator side of the
90 midlatitude jet (Gong et al. 2010, Wang and Magnusdottir 2011). In addition, the SST
91 anomalies associated with the AAO thermodynamically affect the atmosphere in the
92 troposphere and contribute to the persistence of the AAO, suggesting an existence of
93 ocean-to-atmosphere feedback (Sen Gupta and England 2007).

94 Another important driver for the climate variability in the Southern Hemisphere is
95 atmospheric teleconnection associated with the tropical heating anomalies (Lee et al. 2009).
96 In particular, ENSO has significant impacts on the SST anomalies in the extratropics by
97 modulating surface turbulent heat flux and Ekman heat advection (Ciasto and England 2011).
98 The atmospheric response to ENSO has an equivalent barotropic dipole structure of the
99 zonal-mean wind anomalies in the mid-high latitudes and it is also related with the AAO
100 (L'Heureux and Thompson 2006).

101 In this context, a large number of studies discussed the role of the AAO and ENSO in
102 the SST variability in the extratropical Southern Hemisphere, but little attention has been paid
103 to the SST variability associated with the subtropical dipoles as well as the impacts arising
104 from the other tropical forcing. For example, the interannual SST variability in the tropical
105 Indian Ocean shows either Indian Ocean Basin (IOB; Klein et al. 1999) or Indian Ocean

106 Dipole (IOD; Saji et al. 1999) modes. Particularly, the IOD induces the atmospheric
107 teleconnection in the Southern Hemisphere (Chan et al. 2008) and plays an important role in
108 the rainfall variability over Australia (Cai et al. 2011, Cai and Rensch 2013). The Atlantic
109 Niño is another such large interannual SST variability in the tropical Atlantic (Zebiak 1993).

110 Therefore, this study aims to investigate roles of the SST variability in each tropical
111 ocean for generation of the subtropical dipoles in the Southern Hemisphere. For this purpose,
112 we perform a series of SST-nudging experiments by using a CGCM which includes a sea-ice
113 model. The contents of this paper are as follows. In Section 2, we briefly describe CGCM
114 experiments, observational data, and methodology for the atmospheric analysis. In Section 3,
115 we compare characteristics of the subtropical dipoles among the observation and the CGCM
116 experiments. In Section 4, we perform composite analysis to discuss generation of the
117 subtropical dipole among the experiments. The discussion and conclusions are given in
118 Section 5.

119

120 **2. Model, data, and methodology**

121

122 **2-1. CGCM experiments**

123 The CGCM used in this study is SINTEX-F2 (Masson et al. 2012). The atmospheric
124 component is ECHAM5 (Roeckner et al. 2003), which has 31 vertical levels on a T106
125 Gaussian grid. The oceanic component is NEMO (Madec 2008), which includes the LIM2
126 sea ice model (Fichefet et al. 1997) and has ORCA05 horizontal resolution (0.5°) with 31
127 vertical levels. The atmospheric and oceanic fields are exchanged every 2 hours with no flux
128 correction by means of the OASIS3 coupler (Valcke et al. 2004). For the control run (CTR)

129 experiment, the SINTEX-F2 is integrated for 110 years, and monthly mean outputs from the
130 last 80 years are used for the analysis.

131 Using the SINTEX-F2, we conducted three SST-nudging experiments to examine the
132 role of tropical SST variability. Details of each experiment are given in Table 1a. In the FTP
133 experiment, the SST in the tropical Pacific (30°S-30°N) is nudged to climatology of the
134 AVHRR-only Optimum Interpolation SST version 2 data (OISST v2; Reynolds et al. 2007)
135 for the 1982-2010 period. We added a negative feedback ($-2,400 \text{ W m}^{-2} \text{ K}^{-1}$) to the surface
136 heat flux so that the temperature in the upper 50-m mixed layer is restored within 1 day.
137 Within 5 degrees near the boundaries of the SST-nudging area, a Gaussian smoothing is
138 applied. Similarly, the FTIO and FTA experiments were performed in the tropical Indian and
139 Atlantic (30°S-30°N) Oceans, respectively. All of these experiments are integrated for 50
140 years, and the monthly outputs for the last 30 years are used for the analysis by removing the
141 first 20 years for the oceanic adjustment to the atmospheric forcing.

142 For an accurate interpretation of the results from the SST-nudging experiments, it should
143 be noted that the surface heat flux correction in the SST-nudging experiments acts to inhibit
144 coupled interactions and variability in the corrected region to a large extent. In other words,
145 the differences between the CTR and each SST-nudging experiment may allow us to
146 deconvolve the impact of the SST variability in each tropical basin onto the subtropical modes
147 of the Southern Hemisphere.

148

149 **2-2. Data**

150 For comparison, we use the observed monthly mean SST data from the NOAA OISST

151 v2 (Reynolds et al. 2002). It has a horizontal resolution of $1^\circ \times 1^\circ$ and covers the 1982-2011
 152 period. Similar results are obtained for the different datasets such as the HadISST (Rayner et
 153 al. 2003) and ERSST v3 (Smith et al. 2008). We also use the Sea Level Pressure (SLP) and
 154 the geopotential height and horizontal winds at 250 hPa from the NCEP/DOE AMIP-II
 155 Reanalysis (Reanalysis-2; Kanamitsu et al. 2002). It covers the same period with a
 156 horizontal resolution of $2.5^\circ \times 2.5^\circ$. For precipitation, we use the observed data set from CPC
 157 Merged Analysis of Precipitation (CMAP; Xie and Arkin 1997). The resolution and analysis
 158 period are the same as in the NCEP Reanalysis-2. For all the above datasets, monthly
 159 anomalies were calculated after removing the linear trend and the monthly climatology.

160

161 **2-3. Wave activity flux**

162 We use a wave activity flux to examine propagation pathways of stationary Rossby
 163 waves on a zonally varying basic flow in the upper troposphere (Takaya and Nakamura 1997,
 164 2001). It provides a snapshot for three-dimensional propagation of quasi-geostrophic eddies
 165 in a phase independent manner and is parallel to their local group velocity. The formulation
 166 of the wave activity flux \vec{W}_s is given as follows:

$$167 \quad \vec{W}_s = \frac{p}{2|\vec{U}|} \begin{pmatrix} U(\psi_x'^2 - \psi' \psi'_{xx}) + V(\psi_x' \psi_y' - \psi' \psi'_{xy}) \\ U(\psi_x' \psi_y' - \psi' \psi'_{xy}) + V(\psi_y'^2 - \psi' \psi'_{yy}) \\ \frac{f_0^2}{N^2} [U(\psi_x' \psi_z' - \psi' \psi'_{xz}) + V(\psi_y' \psi_z' - \psi' \psi'_{yz})] \end{pmatrix},$$

168 where p = (pressure/1000hPa), $\vec{U} = (U, V)$ is the horizontal wind, ψ is the streamfunction,
 169 f_0 is the Coriolis parameter and N is the buoyancy frequency. Here, $()'$ indicates the
 170 deviation from the monthly climatology. The divergence (convergence) of the wave activity

171 flux corresponds to the generation (dissipation) of the Rossby waves. The wave activity flux,
172 by definition, is based on the quasi-geostrophic approximation so that it is hard to describe the
173 flux near the Equator. Since the Rossby waves are excited by the vorticity perturbation
174 associated with the thermal or orographic forcing (Hoskins and Karoly 1981), here instead,
175 we have used the divergence (convergence) of the horizontal winds in the upper troposphere
176 to describe potential sources of the Rossby waves.

177

178 **3. Impacts on characteristics of subtropical dipoles**

179 Before discussing characteristics of subtropical dipoles, it is worthwhile examining to
180 what extent the SST variances change among the CGCM experiments. Figure 1 shows the
181 standard deviation of SST anomalies during austral summer. The spatial pattern of the
182 observed SST variance is well simulated in the CTR experiment, although the simulated
183 amplitude is relatively weak. In both the observation and CTR experiment, the ENSO pattern
184 clearly appears in the tropical Pacific, and thus the SST variance is much larger than in the
185 other tropical basins. Besides in the tropical Pacific, the SST in the midlatitudes generally
186 shows a large variability compared to that in the high latitudes and this observed feature is
187 well simulated in the CTR experiment.

188 Interestingly, in the absence of tropical SST variability (FTP, FTIO, and FTA
189 experiments), the SST variances in the mid-latitudes do not largely change compared to those
190 in the CTR experiment. This suggests that the amplitude of the SST variability in the
191 midlatitudes may not be strongly influenced by the tropical SST variability. Also, in the
192 absence of the SST variability in the tropical Indian and Atlantic Oceans (FTIO and FTA

193 experiments), the SST variability in the tropical Pacific remarkably increases, suggesting that
194 the SST variability in the tropical Indian and Atlantic may act to suppress the ENSO signal in
195 the CTR experiment. This change in the tropical Pacific is consistent with some CGCM
196 studies (Dommenget et al. 2006, Santoso et al. 2012) but still in contradiction with other
197 studies (Yu et al. 2002, Wu and Kirtman 2004), which need further investigation in a future
198 study. Although the weak SST variability in the tropical Pacific of the CTR experiment, the
199 SST variability in the tropical Pacific for the FTIO and FTA experiments might be closer to
200 the observed SST variability.

201 To detect the SST variability associated with the subtropical dipoles, like previous studies
202 (Huang and Shukla 2006, Kataoka et al. 2012, Morioka et al. 2012), we performed Empirical
203 Orthogonal Function (EOF) analysis to all months of the area-weighted SST anomalies in
204 each subtropical-to-midlatitude oceanic basin. Here, we used the domain of 40°E-110°E and
205 15°S-50°S for the IOSD, 60°W-20°E and 15°S-50°S for the South Atlantic subtropical dipole
206 (SASD), and 150°E-70°W and 25°S-50°S for the South Pacific subtropical dipole (SPSD),
207 respectively.

208 The spatial patterns of EOF modes for each subtropical dipole are shown in Fig. 2, and
209 their corresponding EOF modes and explained variances are given in Tables 1b and c. All of
210 these EOF modes sufficiently separate from the neighboring EOF modes (North et al. 1982),
211 except for the IOSD in the FTP experiment where the first (20.8%) and second EOF (19.1%)
212 are close to each other. Also, it should be noted that the SST-nudging area in the CGCM
213 experiments extends to 30°S, thus overlaps the northeastern pole of the subtropical dipoles, so
214 the IOSD is not defined in the FTIO experiment, in a similar manner for the SASD and SPSD

215 in the FTA and FTP experiments.

216 In the observation, the IOSD is detected as the second EOF mode, whereas in the CTR
217 experiment, it is captured by the first EOF mode (Table 1b). The observed
218 northeast-southwest dipole of SST anomalies is simulated in the higher latitude, and this may
219 lead to a moderately high pattern correlation at 0.44. It is found that the first EOF mode in
220 the observation shows a zonal dipole of the SST anomalies and is linked with the IOB. This
221 IOB-related mode is also detected as the second EOF mode in the CTR experiment (figure not
222 shown). Because of the difference in the EOF analyses, the explained variance of the IOSD
223 in the CTR experiment is 24.5 %, larger than that in the observation, 14.2 % (Table 1c). In
224 the FTP and FTA experiments, the IOSD is identified as the first EOF mode with the
225 explained variances similar to that in the CTR experiment. Also, the spatial patterns of the
226 EOF modes for the IOSD resemble that for the CTR experiment (Fig. 2a). This suggests that
227 the IOSD may exist independent of the SST variability in the tropical Pacific or Atlantic
228 Oceans, which will be further discussed in Section 4.

229 In contrast with the IOSD, the SASD in the observation is detected as the first EOF mode
230 (Table 1b). This mode is also captured in the CTR experiment with a high pattern correlation
231 at 0.75, although the spatial pattern of the EOF mode shows a northeast-southwest oriented
232 dipole, while in the observation, it is associated with a north-south oriented dipole (Fig. 2b).
233 Also, the SASD is identified as the first EOF mode in the FTP and FTIO experiments, and the
234 explained variances are not much different among the experiments (Table 1c). However, the
235 spatial pattern of the EOF mode in the FTP experiment slightly deforms in such a way that the
236 amplitude of the southwestern (northeastern) SST anomaly pole becomes weaker (stronger)

237 than that in the CTR experiment. This implies that although the SASD can develop without
238 the SST variability in the tropical Pacific or Indian Oceans, the SST variability in the tropical
239 Pacific may have more influences on the SASD.

240 Similarly as in the case of the IOSD, the SPSD in the observation is captured by the
241 second EOF mode, which is successfully simulated in the CTR experiment (Fig. 2c, Table 1b).
242 The pattern correlation is high at 0.76. The SPSD is also captured as the third EOF mode in
243 the FTIO experiment and as the second EOF mode in the FTA experiment. It is found that
244 the first EOF modes in the observation and the CGCM experiments show the SST variability
245 related to ENSO, whereas in the FTIO experiment, the SST variability in the second EOF
246 mode is associated with the El Niño Modoki (Ashok et al. 2007). The spatial patterns and
247 explained variances of the SPSD are almost similar among these experiments (Fig. 2c and
248 Table 1c). This suggests that the SST variability in the tropical Indian or Atlantic Oceans
249 may not directly influence the occurrence of the SPSD.

250 The robustness of the subtropical dipoles among the experiments is further examined
251 with regard to the occurrence frequency. Due to the lack of the reliable observations before
252 1980s in the Southern Hemisphere, in particular, the South Pacific (see Fig. 1 in Morioka et al.
253 2013b), we defined subtropical dipole events as years when the principal component of the
254 EOF modes during austral summer exceeds 0.8 standard deviation. The list of event years
255 defined in the observation is given in Table 2, and the occurrence frequency of the total
256 positive and negative events are summarized in Table 3. It should be noted that to compare
257 the occurrence frequency in the CTR experiment with those in the sensitivity experiments, the
258 standard deviation is estimated for occurrence frequencies of the 50 different sets of 30-yr

259 outputs in the CTR experiment. The observed occurrence frequency of the IOSD is 48.3 %,
260 whereas in the CTR experiment, the IOSD occurs at 50.6 % close to the observation.
261 Interestingly, the occurrence frequencies in the FTP and FTA experiments do not largely
262 differ from that in the CTR experiment, suggesting that the interannual SST variability
263 associated with the IOSD may not be influenced by the tropical SST variability. Similarly,
264 the occurrence frequencies of the SASD and SPSD in the sensitivity experiments do not
265 largely change compared to that in the CTR experiment, although the FTP experiment in the
266 SASD shows a slight increase in the occurrence frequency.

267 Another important characteristic of the subtropical dipoles is a strong seasonal
268 dependence in such a way that the SST anomalies associated with the subtropical dipoles
269 develop during austral summer. Here, we calculated a standard deviation of principal
270 components of the EOF modes for each month (Fig. 3). Note that the standard deviation for
271 the 50 different sets of 30-yr outputs in the CTR experiment is also estimated. The IOSD in
272 the observation develops from late austral spring and reaches its peak during austral summer
273 (Fig. 3a). Although there are slight differences in the peak month of the IOSD, all CGCM
274 experiments reasonably simulate the strong seasonality in austral summer with comparable
275 amplitudes. Also, both the SASD and the SPSD clearly show a strong seasonality in austral
276 summer (Figs. 3b and c), but for the SASD, there exists a large difference in the peak month
277 and amplitude among the observation and the CGCM experiments. It is found that the
278 difference in the SST climatology between the observation and the model simulation is much
279 larger in the South Atlantic than in the other oceanic basins (figure not shown). This may
280 have a link with the large disparity in its standard deviation (Fig. 3b). These results suggest

281 that the seasonal phase-locking nature of the subtropical dipoles may not be largely affected
282 by the SST variability in the other tropical oceans.

283

284 **4. Potential sources for the generation of subtropical dipoles**

285 The above intriguing results about key characteristics of the subtropical dipoles motivate
286 us to further examine the generation of subtropical dipoles with a focus on the variations in
287 the subtropical highs. We conduct a composite analysis and discuss only the case for the
288 positive events, because the physical mechanism and its impact on the rainfall in surrounding
289 countries for the negative events is found to be almost similar but opposite to that for the
290 positive events.

291

292 **4-1. Indian Ocean subtropical dipole (IOSD)**

293 Figure 4a shows composite SST anomalies during austral summer of the positive IOSD.
294 Here we used the same criteria for definition of the positive IOSD events as discussed in
295 Section 3, and composited 8, 20, 8, and 6 positive events for the observation, and the CTR,
296 FTP, and FTA experiments, respectively. The SST anomalies in the observation and the CTR
297 experiment show a northeast-southwest oriented dipole in the southern Indian Ocean. In the
298 tropical Pacific, the simulated SST anomalies are associated with a strong La Niña, although
299 they are not significant in the observation. The link with La Niña in the CTR experiment is
300 also seen in the FTA experiment. However, in the FTP experiment, the dipole SST anomalies
301 associated with the IOSD exist without the SST variability in the tropical Pacific. Moreover,
302 the dipole SST anomalies in the FTP experiment resemble those in the CTR experiment,

303 indicating that ENSO may not be necessarily responsible for the occurrence of the IOSD as in
304 the observation, despite of the overemphasized ENSO in the CTR experiment.

305 Since the dipole SST anomalies during the IOSD are closely linked with the SLP
306 anomalies (Behera and Yamagata 2001), we examine composite SLP anomalies in Fig. 4b.
307 Note that the SLP anomalies are calculated for Nov. (0) - Jan. (1) because of the delayed
308 oceanic response to the atmospheric forcing (Fauchereau et al. 2003). In the reanalysis data,
309 the Mascarene High anomalously strengthens, whereas in the CTR experiment, the
310 strengthening of the Mascarene High occurs in the southern part of its mean state ($\sim 30^\circ\text{S}$)
311 during austral summer of the analysis period, indicating anomalous southward shift of the
312 Mascarene High. In the CTR experiment, the positive SLP anomalies are zonally elongated
313 in the mid-latitudes and are associated with the negative SLP anomalies over Antarctica,
314 suggesting a possible link with the positive AAO. The negative SST anomaly in the tropical
315 Indian Ocean may act to weaken the Hadley Cell and cause the negative SLP anomalies in the
316 subtropical region, but this cannot explain the positive SLP anomalies in the southern Indian
317 Ocean. Although not visible in the reanalysis data, the potential link with the AAO in the
318 CTR experiment is also found in the FTP and FTA experiments, but the zonal elongation of
319 the positive and negative SLP anomalies is much clearer in the FTP experiment.

320 In order to determine whether the stronger relationship with ENSO and the AAO in the
321 CTR experiment is due to model bias in simulating the variability inside or outside the
322 southern Indian Ocean, the model results have been further projected onto the observed EOF
323 pattern in Fig. 2a, and the resulting time series have been used as an alternative index for the
324 IOSD to compute composites in the model experiments. As expected, the derived SST

325 anomalies in the CTR experiment compare favorably to the observed ones in the southern
326 Indian Ocean (figure not shown). However, the main differences in the SST and SLP
327 anomalies between the observation and the CTR experiment still remain, especially for the
328 stronger relationship with ENSO and the AAO. This clearly suggests that origins of the
329 biases are probably linked to the overemphasized ENSO and the AAO, not to the fact that the
330 simulated EOF pattern for the IOSD in Fig. 2a is different from the observed one.

331 To further assess the link with the AAO in the model experiments, we defined the AAO
332 by using the geopotential height in the upper troposphere and calculated the correlation
333 coefficient between the IOSD and the AAO. The first EOF mode of the geopotential height
334 anomalies at 250 hPa in the reanalysis data and the CGCM experiments are associated with
335 zonal circular positive anomalies in the midlatitudes and the negative anomalies over
336 Antarctica, representing the AAO (Fig. 5). The variance explained by the first EOF mode in
337 the CTR experiment (14.0 %) is very close to that in the reanalysis data (15.0 %). The
338 correlation coefficient between the time series of the principal component of the
339 corresponding EOF mode for the AAO during Nov.-Jan. and that for the IOSD during
340 Dec.-Feb. is found to be significantly large at 0.45 in the CTR experiment, although not
341 significant in the reanalysis data (Table 4). As expected, both the FTP and FTA experiments
342 show a significantly high correlation with the AAO, and in particular, the FTP experiment
343 shows higher correlation (0.72) than that in the CTR experiment. The comparison of the CTR
344 and FTP experiments implies that, without the SST variability in the tropical Pacific, the
345 AAO may play more important role in inducing the IOSD (Hermes and Reason 2005,
346 Morioka et al. 2013a).

347 To investigate a physical link with the AAO in the model experiments, we calculated
348 composite anomalies of geopotential height at 250 hPa and wave activity flux during
349 Nov.(0)-Jan.(1) (Fig. 6a). Positive geopotential height anomalies in the southern Indian
350 Ocean are located above the positive SLP anomalies in the reanalysis data and the CTR
351 experiment, representing an equivalent barotropic structure in the troposphere. In the CTR
352 experiment, the positive anomalies are zonally elongated in the midlatitudes and associated
353 with a propagation of stationary Rossby waves from the negative anomalies over Antarctica,
354 suggesting the physical connection with the AAO. A similar feature is also found in the FTP
355 and FTA experiments.

356 To reveal potential sources of the stationary Rossby waves, we analyzed composite
357 anomalies of the velocity potential and divergent winds at 250 hPa and the rainfall during
358 Nov.(0)-Jan.(1) in Figs. 6b, and 7. Both the reanalysis data and the CTR experiment show an
359 anomalous wind divergence west of the Antarctic Peninsula, but there is a large disparity in
360 the western tropical Pacific. In the CTR experiment, anomalous divergence occurs in
361 association with La Niña when the seasonal rainfall enhances in the western tropical Pacific
362 (Fig. 7). The associated diabatic heating acts to generate the Rossby waves propagating along
363 the great circle, generating the Pacific-South American (PSA; Mo and Paegle 2001) like
364 teleconnection pattern in the South Pacific. The anomalous rainfall increase is also found
365 over adjacent countries in the southern Indian Ocean, in particular, southern Africa in the
366 CTR experiment, although not significant in the observation. This may be due to the
367 anomalous moisture supply from the southern Indian Ocean linked with the westward
368 extension of the anomalous Mascarene High (Fig. 4b). On the other hand, the rainfall

369 increase in Australia may be more related to La Niña in the tropical Pacific.

370 A similar feature of the wind divergence anomalies can be seen in the FTA experiment
371 associated with the increased rainfall in the western tropical Pacific (Fig. 7). However, in the
372 FTP experiment, the spatial pattern of the wind divergence anomalies resembles that in the
373 reanalysis data (Fig. 6b), except that the anomalous divergence occurs in the tropical Indian
374 Ocean in association with the enhanced rainfall in the region (Fig. 7). A comparison of the
375 reanalysis and the FTP experiment suggests that without the SST variability in the tropical
376 Pacific and associated interferences, the anomalous wind divergences in the upper
377 troposphere of mid-high latitudes may excite the stationary Rossby waves and induce the
378 AAO-related positive SLP anomalies in the southern Indian Ocean.

379

380 **4-2. South Atlantic subtropical dipole (SASD)**

381 The physical mechanism for the SASD is found to be almost similar to that for the IOSD.
382 Similarly as in the IOSD, we conducted the composite analysis and used 6, 21, 9, and 7
383 positive events for the observation, and the CTR, FTP, and FTIO experiments, respectively.
384 Composite SST anomalies in the observation and the CTR experiment show a
385 northeast-southwest oriented dipole in the South Atlantic (Fig. 8a). These are associated with
386 the negative SST anomalies in the central tropical Pacific related to La Niña. On the other
387 hand, in the FTP experiment, the dipole SST anomaly pattern slightly deforms in such a way
388 that the northeastern pole of the negative SST anomaly extends to the southwest of South
389 Africa. This is expected from the spatial pattern of the corresponding EOF mode in Fig. 2b.
390 Also, in the FTP and FTIO experiments, the SST anomalies in the tropical Pacific show no

391 significant signal, implying that ENSO may not be a necessary condition for generating the
392 SASD.

393 Similarly as in the IOSD, St. Helena High in the reanalysis data and the CTR experiment
394 anomalously strengthens in its southern part of the mean state ($\sim 30^{\circ}\text{S}$) during austral summer
395 of the analysis period, indicating anomalous southward shift of St. Helena High (Fig. 8b).
396 The positive SLP anomalies do not appear to be directly linked with the SST variability in the
397 tropical Atlantic, because the negative SST anomalies in the region tend to weaken the
398 Hadley Cell and cause the negative SLP anomalies in the subtropical region. Rather, they are
399 associated with the negative (positive) SLP anomalies in the western (eastern) tropical Pacific
400 related to La Niña. Also, over Antarctica, the negative SLP anomalies dominate, indicating a
401 possible link with the positive AAO. On the other hand, the link with the positive AAO is not
402 visible in the FTP and FTIO experiments. In fact, the correlation coefficients between the
403 AAO and the SASD in the reanalysis data and the CTR experiment are significantly high, but
404 those in the FTP and FTIO experiment are not significant (Table 4). This implies that in both
405 FTP and FTIO experiments, there may exist some internal atmospheric processes responsible
406 for generating the positive SLP anomalies in the South Atlantic.

407 To reveal the physical mechanism on the positive SLP anomalies during the positive
408 SASD, composite anomalies of the geopotential height and wave activity flux at 250 hPa are
409 shown in Fig. 9a. In the reanalysis data, the positive geopotential height anomalies in the
410 South Atlantic are associated with a clear propagation of stationary Rossby waves from the
411 positive anomalies in the South Pacific through the negative anomalies west of the Antarctic
412 Peninsula. A similar propagation of the Rossby waves is found in the CTR experiment.

413 However, in the FTP experiment, the positive anomalies in the South Atlantic are associated
414 with a propagation of the Rossby waves from the positive anomalies west of Antarctic
415 Peninsula through the negative anomalies over the southern South America. Also, in the
416 FTIO experiment, the positive anomalies in the South Atlantic are associated with the
417 negative anomalies west of Antarctic Peninsula. These suggest that the atmospheric
418 variability west of Antarctic Peninsula may have influence on the positive anomalies in the
419 South Atlantic through the Rossby wave propagation, although no significant link with the
420 AAO is found in the FTP and FTIO experiments.

421 For potential sources of the stationary Rossby waves, composite anomalies of the
422 velocity potential and divergent winds at 250 hPa in the reanalysis data and the CTR
423 experiment show anomalous wind divergence near the Maritime Continent and New Zealand
424 as well as the Antarctic Peninsula (Fig. 9b). In particular, the anomalous wind divergences
425 over the Maritime Continent are related to the enhanced rainfall associated with La Niña (Fig.
426 10). The significant rainfall increase is also found over southeast Brazil both in the
427 observation and the CTR experiment, which may be partly related to the strengthening of the
428 St. Helena High supplying more moisture from the South Atlantic (Fig. 8b). Although the
429 spatial patterns of the wind divergence anomalies in the FTP and FTIO experiments resemble
430 that in the CTR experiment, there is a clear difference in the wind divergence anomalies in the
431 midlatitudes. Therefore, in the absence of the SST variability in the tropical Pacific or Indian
432 Oceans, the anomalous wind divergences in the midlatitudes may act to generate the
433 stationary Rossby waves, hence the positive SLP anomalies in the South Atlantic.

434 Finally, a strong coherence between the SASD and the IOSD is found in the analyzed

435 results. This intriguing aspect was discussed by Hermes and Reason (2005), suggesting that
436 they are highly correlated under the large-scale atmospheric forcing related to the AAO. In
437 the reanalysis data and the CTR experiment (Figs. 8b and 9a), the stationary Rossby waves
438 clearly propagate from the positive SLP anomalies in the South Atlantic through the negative
439 SLP anomalies over Antarctica to the positive SLP anomalies in the southern Indian Ocean.
440 This suggests that the atmospheric Rossby waves play an important role in physically
441 connecting the SLP anomalies, hence the subtropical dipoles in the two oceans. However,
442 some exceptions are found when some of the IOSDs occur independent of the SASDs and
443 some of the positive SASDs occur in association with the negative IOSDs and vice versa. In
444 these cases, the atmospheric signal related to the AAO is absent and less prominent than the
445 above coherent cases (Hermes and Reason 2005). This may be related to the non-linear
446 behavior of the mid-high latitude atmosphere.

447

448 **4-3. South Pacific subtropical dipole (SPSD)**

449 The physical mechanism on the generation of the SPSPD is found not to largely differ
450 from the IOSD and the SASD. Similarly, we performed composite analysis to 7, 18, 9, and 7
451 positive events for the observation, and the CTR, FTIO, and FTA experiments, respectively.
452 The positive SPSPDs in the observation and the CTR experiment show slightly east-west
453 oriented dipole of positive and negative SST anomalies in the central South Pacific (Fig. 11a).
454 These are associated with the negative SST anomalies southeast of Australia and the positive
455 SST anomalies west of South America. Interestingly, both the observation and the CTR
456 experiment show an association with La Niña in the tropical Pacific. The similar dipole SST

457 anomaly is found in the FTIO and FTA experiments, but in the FTA experiment, the SST
458 anomaly in the tropical Pacific is associated with El Niño in the central tropical Pacific. This
459 implies that a phase of ENSO may have nothing to do with the occurrence of the SPSD.

460 Composite SLP anomalies in the reanalysis data and the CTR experiment (Fig. 11b) show
461 a wave train of positive and negative SLP anomalies in the central South Pacific, as discussed
462 in Morioka et al. (2013b). The wave train of the SLP anomalies does not seem to be directly
463 linked with the negative SST anomaly in the tropical Pacific, because it acts to cause
464 weakening of the Hadley Cell and hence the negative SLP anomalies in the subtropical region.
465 Rather, associated with the La Niña condition, both the reanalysis and the CTR experiment
466 show positive (negative) SLP anomalies in the eastern (western) tropical Pacific. On the
467 other hand, in the FTIO and FTA experiments, the wave train of the SLP anomalies is
468 associated with the negative SLP anomalies in the midlatitudes and the positive SLP
469 anomalies over Antarctica, indicating a possible link with the negative AAO. Although the
470 correlation coefficients between the SPSD and the AAO are not significant in the reanalysis
471 data and the CTR experiment, they are significantly negative in the FTIO and FTA
472 experiments.

473 In the upper troposphere, both the reanalysis data and the CTR experiment clearly show a
474 wave train of positive and negative geopotential height anomalies in the South Pacific (Fig.
475 12a). They are associated with a propagation of stationary Rossby waves from negative
476 anomalies southeast of Australia, which are partly related to the wave propagation from the
477 negative anomalies in the southern Indian Ocean through the positive anomalies over
478 Antarctica. A similar wave propagation is found in the FTA and FTIO experiments, but the

479 Rossby waves propagate between the negative anomalies in the midlatitudes and the positive
480 anomalies over Antarctica, indicating a strong connection with the negative AAO.

481 One of the potential sources of the stationary Rossby waves in the reanalysis data and the
482 CTR experiment may be the anomalous wind divergence in the western tropical Pacific (Fig.
483 12b). This wind divergence anomaly, although not significant, is related to the enhanced
484 rainfall over the Maritime Continent associated with La Niña (Fig. 13). Another possible
485 source may be the anomalous wind divergence in the southern Indian Ocean for the reanalysis
486 data, as discussed by Morioka et al. (2013b). On the other hand, in the FTIO experiment,
487 anomalous wind divergences significantly occur southwest of Australia as well as in the
488 tropical Indian Ocean, which are related to the anomalous rainfall increase in the southern
489 tropical Indian Ocean (Fig. 13). Since the SST variability is suppressed in the tropical Indian
490 Ocean, the enhanced rainfall may be due to an internal atmospheric variability in the tropical
491 basin. Also, in the FTA experiment, the wind divergence anomaly is found in the central
492 tropical Pacific related to El Niño and also south of South Africa (Fig. 12b). These results
493 suggest that even without the SST variability in the tropical Indian or Atlantic Oceans,
494 anomalous wind divergences in the midlatitudes may act to induce the stationary Rossby
495 waves, hence generate the wave train of the positive and negative SLP anomalies in the South
496 Pacific.

497

498 **5. Summary and discussion**

499 The role of tropical SST variability in the generation of subtropical dipoles is
500 investigated by conducting a series of CGCM experiments in the tropical oceans and the

501 results for each subtropical dipole are summarized in the schematic diagram (Fig. 14). The
502 CGCM reasonably reproduces the northeast-southwest oriented dipole of positive and
503 negative SST anomalies associated with the subtropical dipole in each oceanic basin, although
504 there exist some discrepancies for the IOSD due to the overemphasized ENSO and the AAO
505 in the model. In the experiments where the SST variability in each tropical basin is
506 suppressed by nudging the simulated SST to the observed climatology, the subtropical dipoles
507 in each oceanic basin are found to develop in association with the variations in the subtropical
508 highs. Interestingly, the occurrence frequencies of subtropical dipoles as well as their strong
509 seasonality in austral summer do not largely change in most cases of the CGCM experiments.
510 Composite analysis reveals that in the absence of the tropical SST variability, the IOSD and
511 the SPSD are associated with the AAO, whereas the SASD is associated with a wave train of
512 SLP anomalies in the midlatitudes. It is found that the Rossby waves generated in the
513 midlatitudes play an important role in inducing the SLP anomalies, hence generating the
514 subtropical dipoles. The results for the IOSD are consistent with the previous CGCM study
515 by Morioka et al. (2013a), although the link with the AAO is not significant in the composite
516 fields of reanalysis data (Table 4) and our model has a tendency to simulate the IOSD in
517 association with the AAO.

518 The most intriguing result in this study is that, when the tropical SST variability is
519 suppressed, the AAO plays an important role in the atmospheric variability of the midlatitudes,
520 and is able to induce the SST variability associated with the subtropical dipoles. The
521 extratropical SST response to the AAO and/or ENSO has been widely discussed in the
522 previous literature (Sen Gupta and England 2006, Ciasto and Thompson 2008, Ciasto and

523 England 2011), but few studies paid attention to the SST variability associated with
524 subtropical dipoles and examined the relative contribution from the AAO and the tropical
525 SST variability to the generation of subtropical dipoles. Our CGCM results suggest that the
526 tropical SST variability acts to suppress or counteract the AAO activity inherently existing in
527 mid-high latitudes. This tendency is clearly seen in the standard deviation of the geopotential
528 height anomalies at 250 hPa during Nov.-Jan. (Fig. 15a). Both the reanalysis data and the
529 CTR experiment show large interannual variations in the mid-high latitudes, which are
530 closely linked with the AAO. Interestingly, the standard deviation over the Antarctica in the
531 FTP and FTA experiments becomes larger than that in the CTR experiment. However, in the
532 FTIO experiment, the amplitude of the standard deviation does not largely change compared
533 to the CTR experiment.

534 Since the atmospheric variability in the upper troposphere of the mid-high latitudes is
535 also influenced by the atmospheric teleconnection through the rainfall variability induced by
536 the SST variability in the tropical oceans (Cai et al. 2011, Ding et al. 2012), we calculated the
537 standard deviation of the rainfall anomalies in the same season (Fig. 15b). Both the
538 observation and CTR experiment show large rainfall variability in the western tropical Pacific
539 and to a lesser extent in the tropical Indian Ocean. The rainfall variability in these regions is
540 suppressed in the FTP and FTA experiments, but does not largely change in the FTIO
541 experiment. This implies that the atmospheric teleconnection from the western tropical
542 Pacific and the tropical Indian Oceans becomes weaker in the FTP and FTA experiments, and
543 the AAO becomes able to play a major role in the SLP variations of the mid-high latitudes.
544 The difference in the rainfall variability in the experiments may be due to the basin

545 interactions through the change in the walker circulation related to that in the SST variability
546 (Saravanan and Chang 2000), but this is beyond the scope of this study.

547 Finally, this study provides much implication for both simulation and prediction of
548 subtropical dipoles. To realistically simulate and accurately predict the subtropical dipoles, it
549 is necessary to represent the AAO signal in the model at least in one season prior to the
550 occurrence of the subtropical dipoles. A previous study by Yuan et al. (2013) suggests that it
551 is difficult to predict the subtropical dipoles that are not associated with ENSO. The AAO,
552 itself, is the internal atmospheric variability sustained by the eddy-mean interaction and exists
553 even without the SST variability (Yamazaki and Shinya 1999, Limpasuvan and Hartmann
554 1999). However, some studies claim an important role of ocean-to-atmosphere feedback in
555 the AAO (Sen Gupta and England 2007), which is worthwhile to examine in the further
556 CGCM study. In addition, there are external factors such as ozone influencing the AAO
557 activity (Sexton 2001), which are not taken into account in the present study. For the realistic
558 simulation of the AAO and hence the subtropical dipoles, it is necessary to reproduce at least
559 both the oceanic and atmospheric mean states in the mid-high latitudes such as the SST,
560 sea-ice, and westerly jet. In this regard, many efforts are required for realistic simulation and
561 accurate prediction of the AAO to represent the climate variability in the midlatitudes of the
562 Southern Hemisphere.

563

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750 **Tables**

751 **Table 1:** (a) Details of CGCM experiments performed in this study. (b) EOF modes of SST
 752 anomalies in each subtropical-to-midlatitude oceanic basin to define the IOSD, SASD, and
 753 SPSD for the observation (OISST) and CGCM (CTR, FTP, FTIO, and FTA) experiments. (c)
 754 Explained variances (in %) for the corresponding EOF mode in (b).

(a) CGCM experiments

	SST nudging area	Analysis period
CTR	None	80 yr
FTP	Tropical Pacific (30°S-30°N)	30 yr
FTIO	Tropical Indian (30°S-30°N)	30 yr
FTA	Tropical Atlantic (30°S-30°N)	30 yr

(b) EOF mode

	OISST	CTR	FTP	FTIO	FTA
IOSD	2	1	1		1
SASD	1	1	1	1	
SPSD	2	2		3	2

(c) Explained variance %

	OISST	CTR	FTP	FTIO	FTA
IOSD	14.2	24.5	20.8		25.8
SASD	20.6	19.0	23.1	21.5	
SPSD	13.2	11.8		11.1	12.3

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757 **Table 2:** Observed event years for the IOSD, SASD, and SPSD. Pos (Neg) corresponds to
 758 positive (negative) events.

Event years	
	IOSD
Pos	1986, 1987, 1992, 1998, 2003, 2005, 2006, 2010
Neg	1982, 1989, 1991, 1995, 1999, 2002
	SASD
Pos	1989, 1996, 1999, 2001, 2008, 2010
Neg	1982, 1987, 1997, 2000, 2002, 2009
	SPSD
Pos	1983, 1984, 1987, 2001, 2003, 2005, 2008
Neg	1986, 1990, 1994, 1997, 2009, 2010

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762 **Table 3:** Occurrence frequency (in %) of the IOSD, SASD, and SPSD for the observation
 763 (OISST) and CGCM (CTR, FTP, FTIO, and FTA) experiments. Both the positive and
 764 negative events are counted together. The values in the parenthesis of the CTR experiment
 765 indicate one standard deviation calculated from 50 different sets of 30-yr outputs.

	Occurrence frequency					%
	OISST	CTR	FTP	FTIO	FTA	
IOSD	48.3	50.6 (± 5.8)	48.3			51.7
SASD	41.4	50.6 (± 6.2)	58.6	48.3		
SPSD	44.8	49.4 (± 10.3)		58.6		55.2

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767

768 **Table 4:** Correlation coefficients among the IOSD, SASD, and SPSD indices during
769 Dec.-Feb. and the AAO index during Nov.-Jan. for the NCEP Reanalysis-2 (NCEP2) data and
770 CGCM (CTR, FTP, FTIO, and FTA) experiments. The asterisk indicates a correlation
771 coefficient exceeding 90 % (95 %) confidence level by the two-tailed t-test for the NCEP2
772 and FTP, FTIO, and FTA experiments (CTR experiment).

Correlation with AAO

	NCEP2	CTR	FTP	FTIO	FTA
IOSD	0.16	0.45*	0.72*		0.54*
SASD	0.67*	0.39*	-0.28	-0.17	
SPSD	0.22	-0.22		-0.42*	-0.60*

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776 **Figure captions**

777 **Figure 1:** Standard deviation (in °C) of SST anomalies during austral summer for (a) the
778 observation (OISST) and (b) the CGCM CTR experiment. (c-e) As in (b), but for the FTP,
779 FTIO, and FTA experiments, respectively.

780 **Figure 2:** (a) Spatial patterns (in °C) of the EOF modes for the IOSD defined in Table 1b.
781 (b-c) As in (a), but for the SASD and SPSD, respectively. Positive values are shaded.

782 **Figure 3:** (a) Monthly standard deviation of the principal component of the corresponding
783 EOF modes for the IOSD defined in Table 1b. The red shade in the CTR experiment indicates
784 one standard deviation calculated from 50 different sets of 30-yr outputs. (b-c) As in (a), but
785 for the SASD and SPSD, respectively.

786 **Figure 4:** (a) Composite anomalies of SST (in °C) during austral summer of the positive
787 IOSD for the observation (OISST) and CGCM (CTR, FTP, and FTA) experiments. (b) As in
788 (a), but for the SLP (in hPa) during Nov.(0)-Jan.(1). For comparison, NCEP Reanalysis-2
789 (NCEP2) are shown. Anomalies exceeding the 90% (95%) confidence level by two-tailed
790 t-test are shown for the observation, FTP, and FTA experiments (CTR experiment).

791 **Figure 5:** Spatial patterns of the first EOF mode of the geopotential height anomalies at 250
792 hPa (contour interval is 10 m). The NCEP Reanalysis-2 (NCEP2), the CGCM (CTR, FTP,
793 FTIO, and FTA) experiments are shown, respectively. The positive values are shaded.
794 Values on the top of the panel indicate the explained variance.

795 **Figure 6:** (a) As in Fig. 4b, but for the geopotential height at 250 hPa (color, in m) and wave
796 activity flux (arrow, in $\text{m}^2 \text{s}^{-2}$). (b) As in (a), but for the velocity potential at 250 hPa (color, in
797 $10^6 \text{m}^2 \text{s}^{-1}$) and divergent wind (arrow, in m s^{-1}).

798 **Figure 7:** As in Fig. 4b, but for the rainfall (color, in mm month^{-1}).

799 **Figure 8:** As in Fig. 4, but for the positive SASD. The FTP and FTIO experiments are shown.

800 **Figure 9:** As in Fig. 6, but for the positive SASD. The FTP and FTIO experiments are shown.

801 **Figure 10:** As in Fig. 7, but for the positive SASD. The FTP and FTIO experiments are
802 shown.

803 **Figure 11:** As in Fig. 4, but for the positive SPSD. The FTIO and FTA experiments are
804 shown.

805 **Figure 12:** As in Fig. 6, but for the positive SPSD. The FTIO and FTA experiments are
806 shown.

807 **Figure 13:** As in Fig. 7, but for the positive SPSD. The FTIO and FTA experiments are
808 shown.

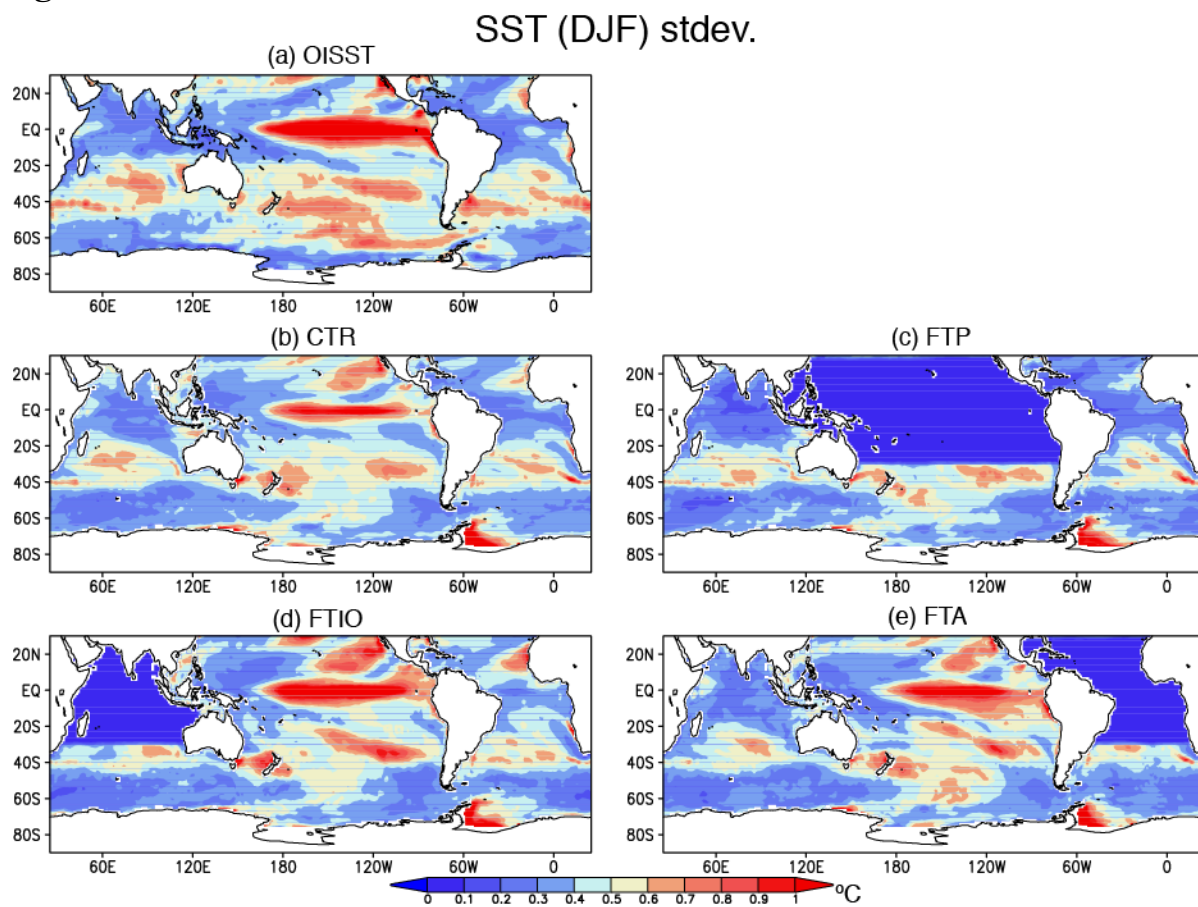
809 **Figure 14:** Schematic diagram illustrating the generation mechanism of the IOSD, SASD,
810 and SPSD derived from the reanalysis data and the CTR experiment. The warm (cold) color
811 indicates the positive (negative) SST anomalies and the black shade indicates the anomalous
812 wind divergence in the upper troposphere as possible sources of the Rossby waves. The
813 dashed line shows the propagation pathway of the anomalous high (H) and low (L) pressure
814 in the upper troposphere.

815 **Figure 15:** As in Fig. 1, but for (a) geopotential height at 250 hPa (in m) during Nov.-Jan.. (b)
816 As in (a), but for the precipitation (in mm month^{-1}).

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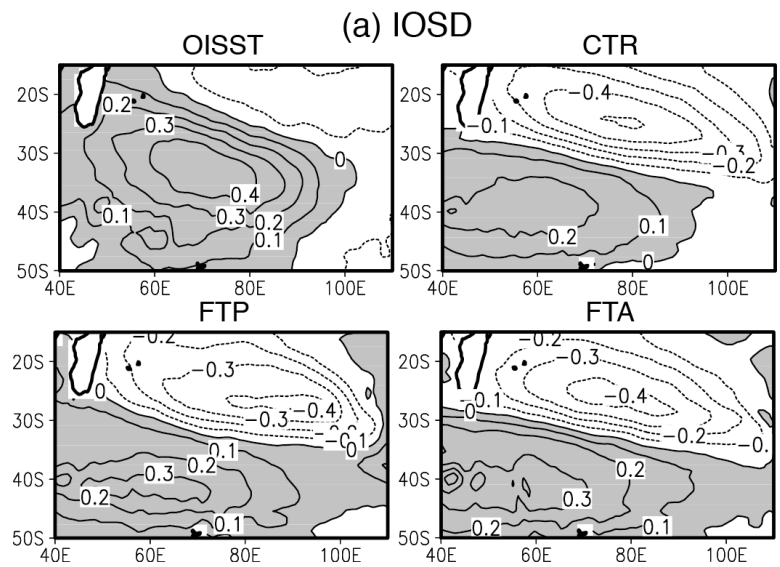
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819 **Figures**

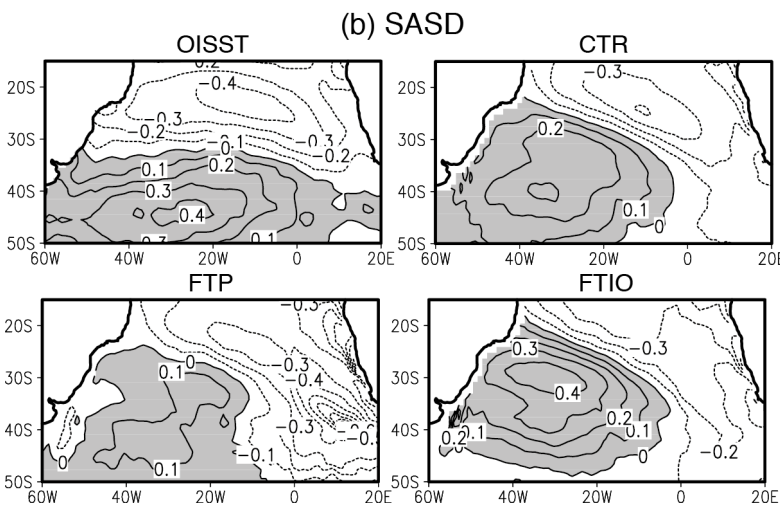


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821 **Figure 1:** Standard deviation (in °C) of SST anomalies during austral summer for (a) the
822 observation (OISST) and (b) the CGCM CTR experiment. (c-e) As in (b), but for the FTP,
823 FTIO, and FTA experiments, respectively.
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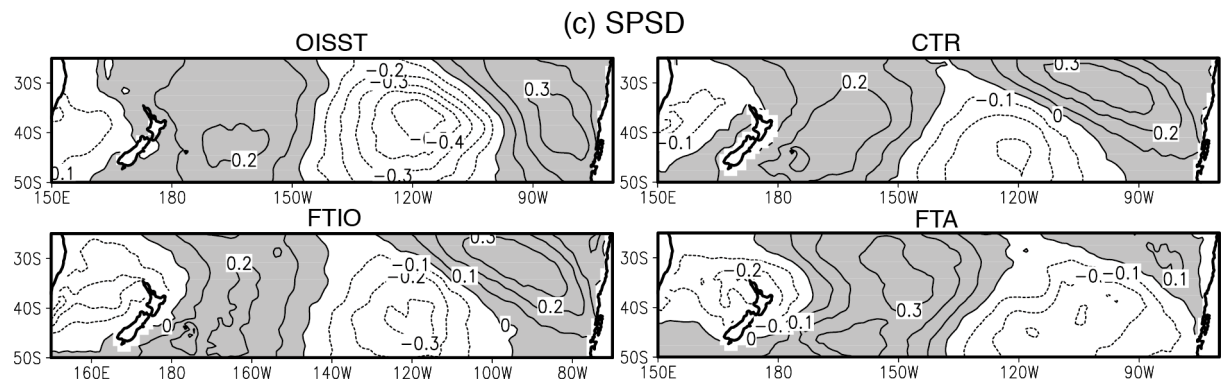
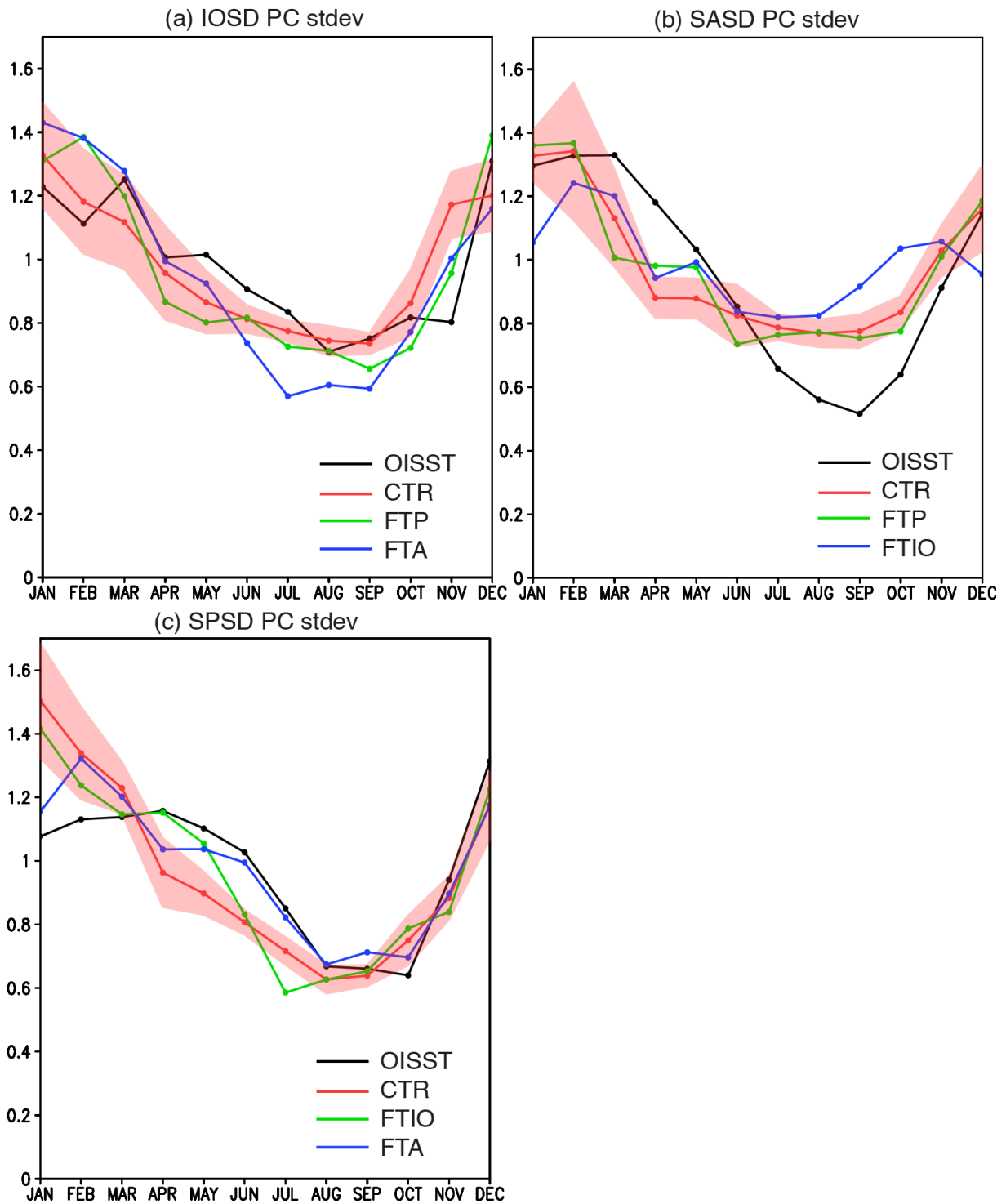


Figure 2: (a) Spatial patterns (in $^{\circ}\text{C}$) of the EOF modes for the IO SD defined in Table 1b.

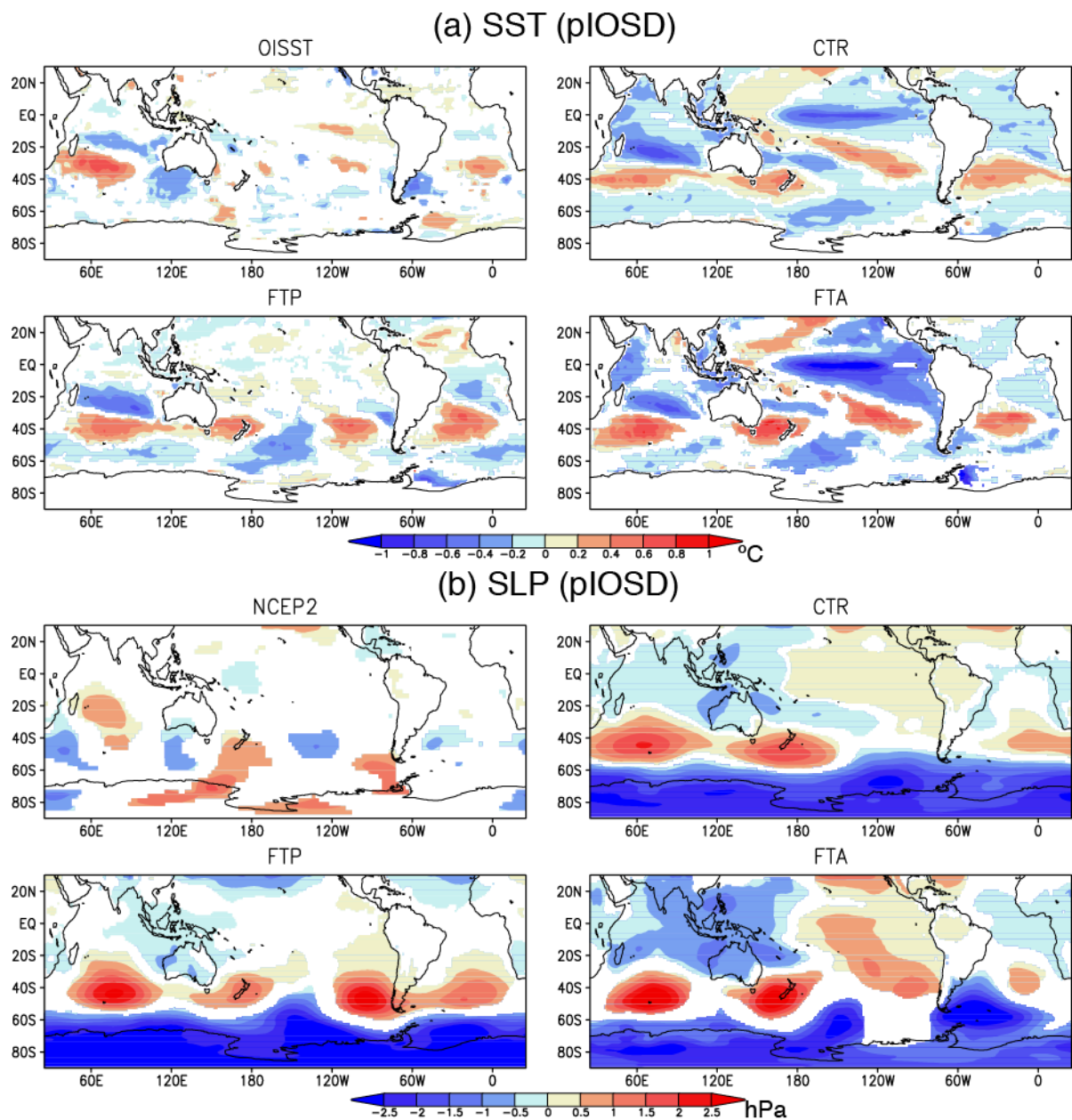
(b-c) As in (a), but for the SASD and SPSD, respectively. Positive values are shaded.



831

832 **Figure 3:** (a) Monthly standard deviation of the principal component of the corresponding
 833 EOF modes for the IOSED defined in Table 1b. The red shade in the CTR experiment indicates
 834 one standard deviation calculated from 50 different sets of 30-yr outputs. (b-c) As in (a), but
 835 for the SASD and SPSD, respectively.

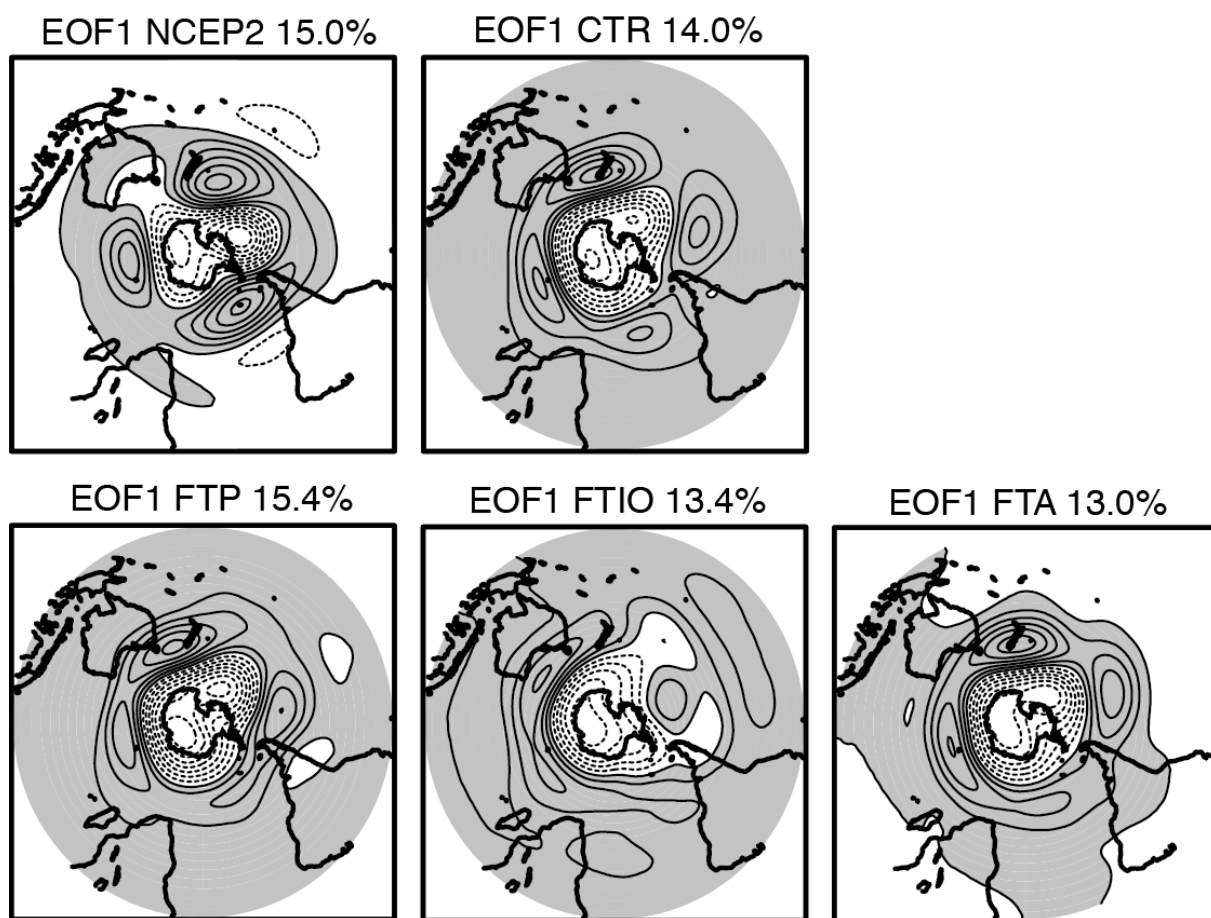
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838 **Figure 4:** (a) Composite anomalies of SST (in $^{\circ}\text{C}$) during austral summer of the positive
 839 IOSD for the observation (OISST) and CGCM (CTR, FTP, and FTA) experiments. (b) As in
 840 (a), but for the SLP (in hPa) during Nov.(0)-Jan.(1). For comparison, NCEP Reanalysis-2
 841 (NCEP2) are shown. Anomalies exceeding the 90% (95%) confidence level by two-tailed
 842 t-test are shown for the observation, FTP, and FTA experiments (CTR experiment).

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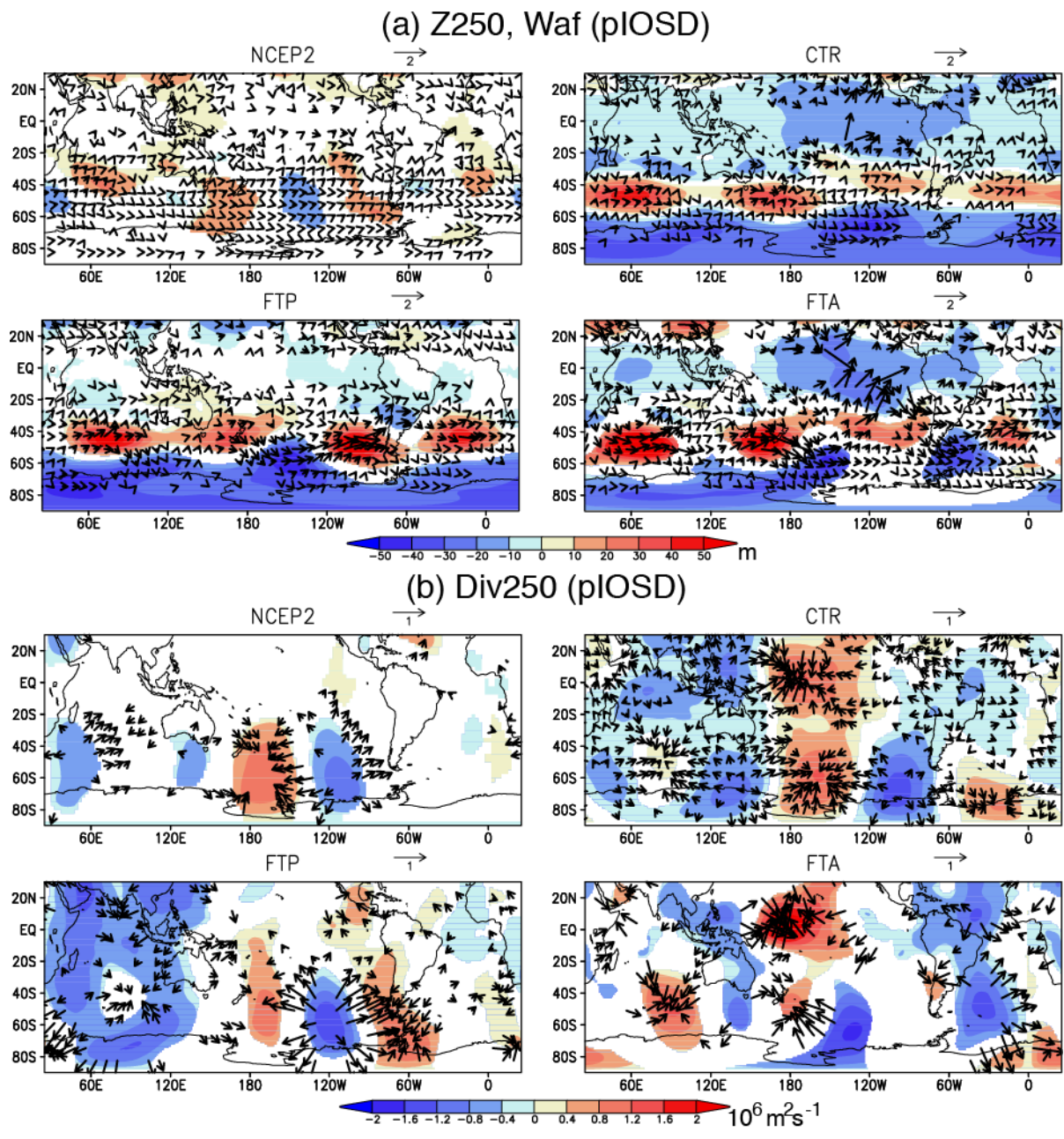


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845 **Figure 5:** Spatial patterns of the first EOF mode of the geopotential height anomalies at 250
 846 hPa (contour interval is 10 m). The NCEP Reanalysis-2 (NCEP2), the CGCM (CTR, FTP,
 847 FTIO, and FTA) experiments are shown, respectively. The positive values are shaded.
 848 Values on the top of the panel indicate the explained variance.

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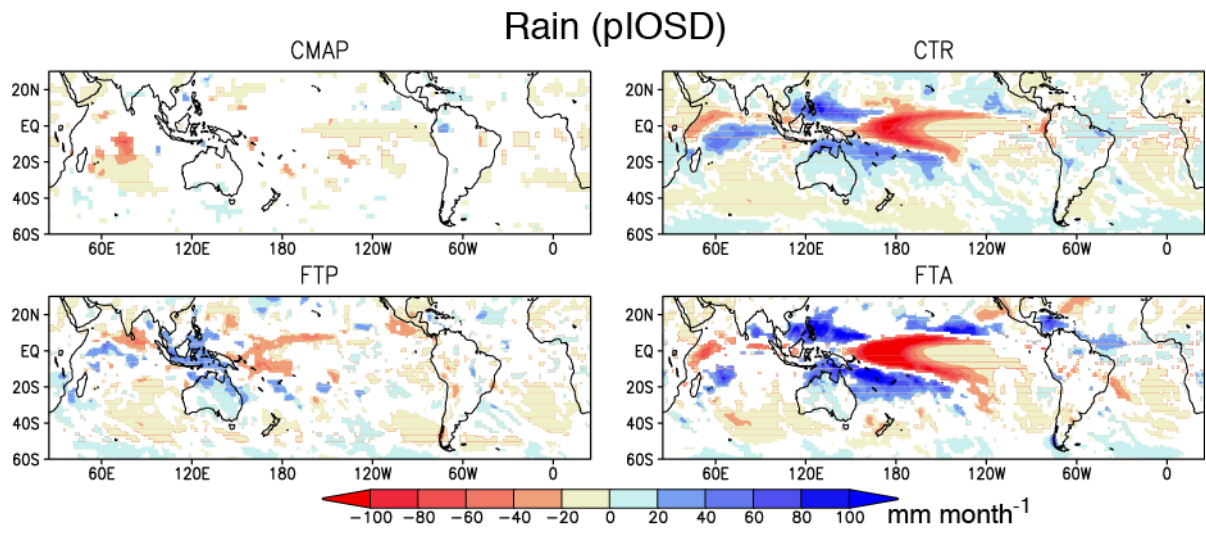


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852 **Figure 6:** (a) As in Fig. 4b, but for the geopotential height at 250 hPa (color, in m) and wave
 853 activity flux (arrow, in $\text{m}^2 \text{s}^{-2}$). (b) As in (a), but for the velocity potential at 250 hPa (color, in
 854 $10^6 \text{m}^2 \text{s}^{-1}$) and divergent wind (arrow, in m s^{-1}).

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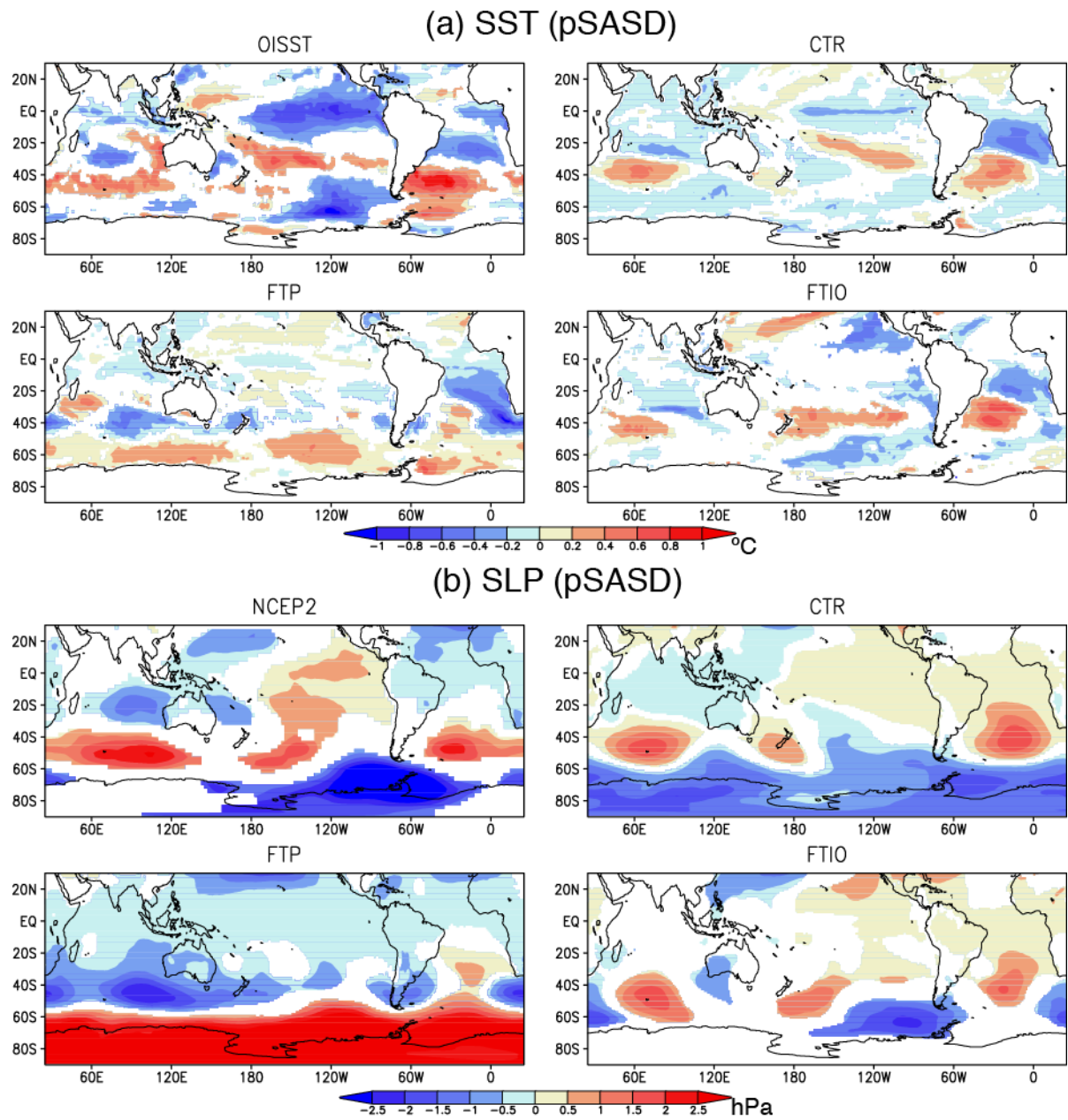


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858 **Figure 7:** As in Fig. 4b, but for the rainfall (color, in mm month⁻¹).

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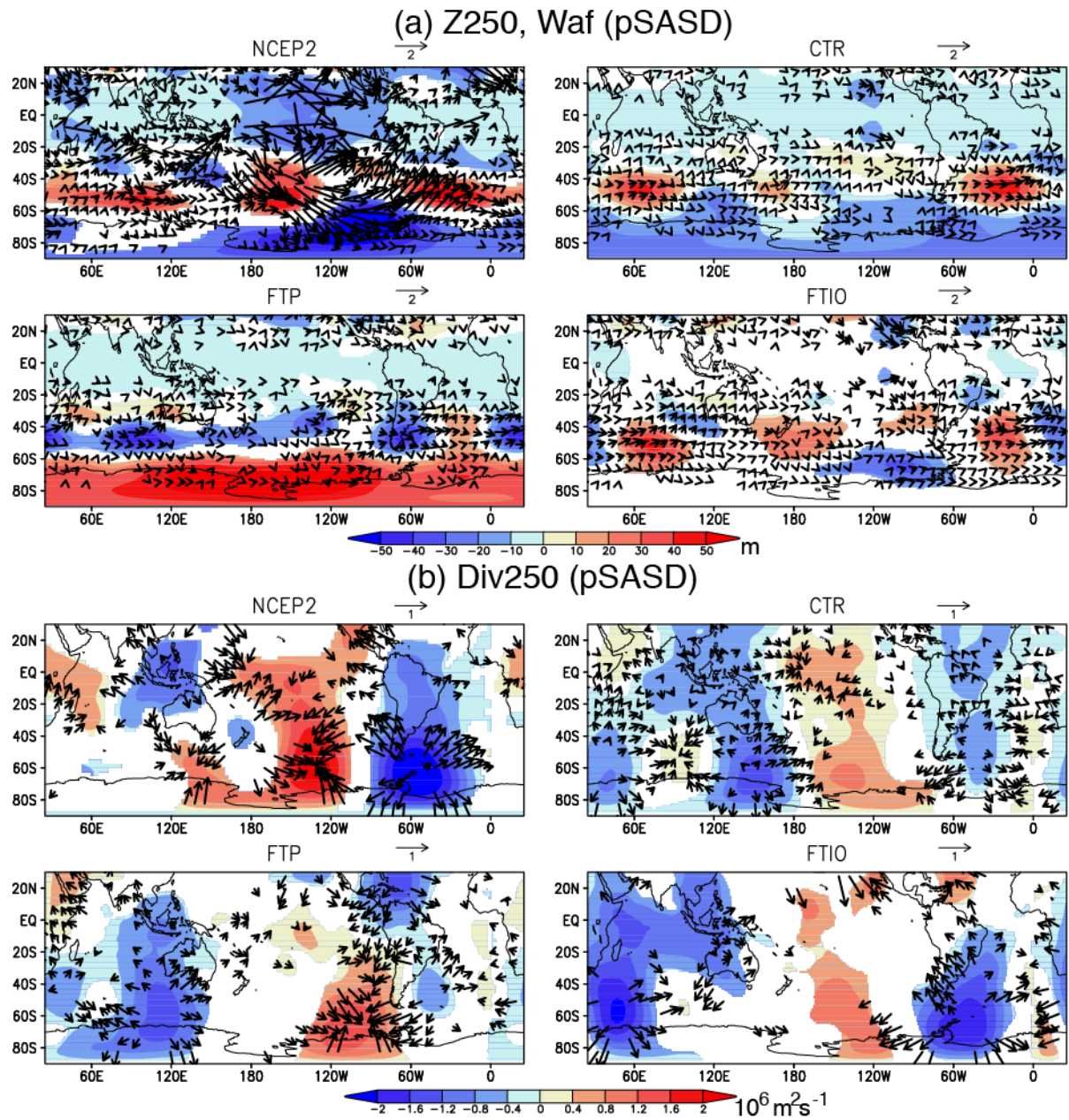
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862 **Figure 8:** As in Fig. 4, but for the positive SASD. The FTP and FTIO experiments are shown.

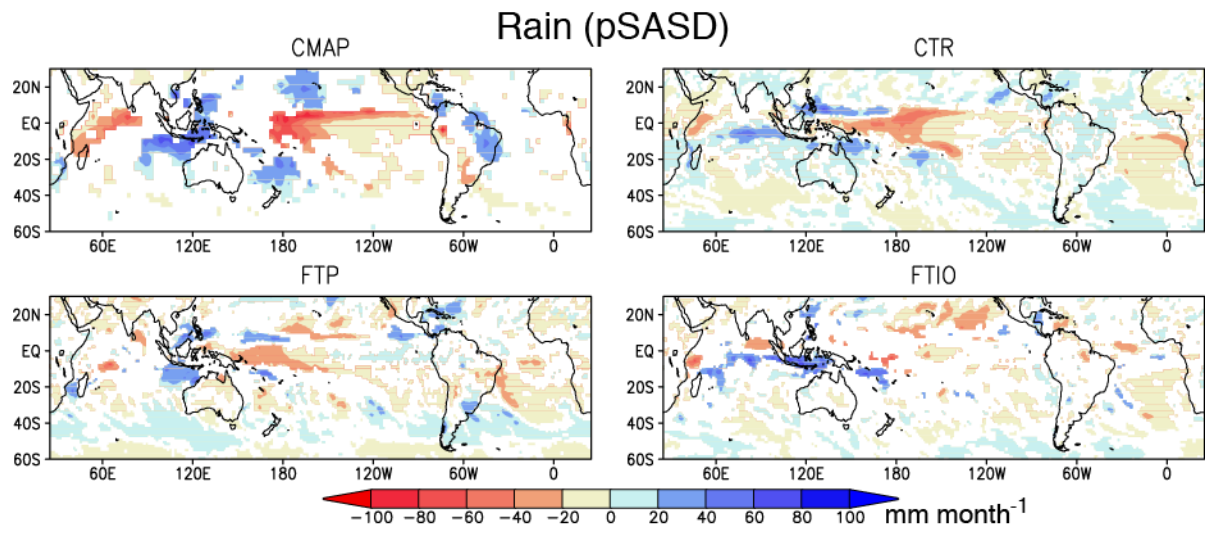
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865 **Figure 9:** As in Fig. 6, but for the positive SASD. The FTP and FTIO experiments are shown.

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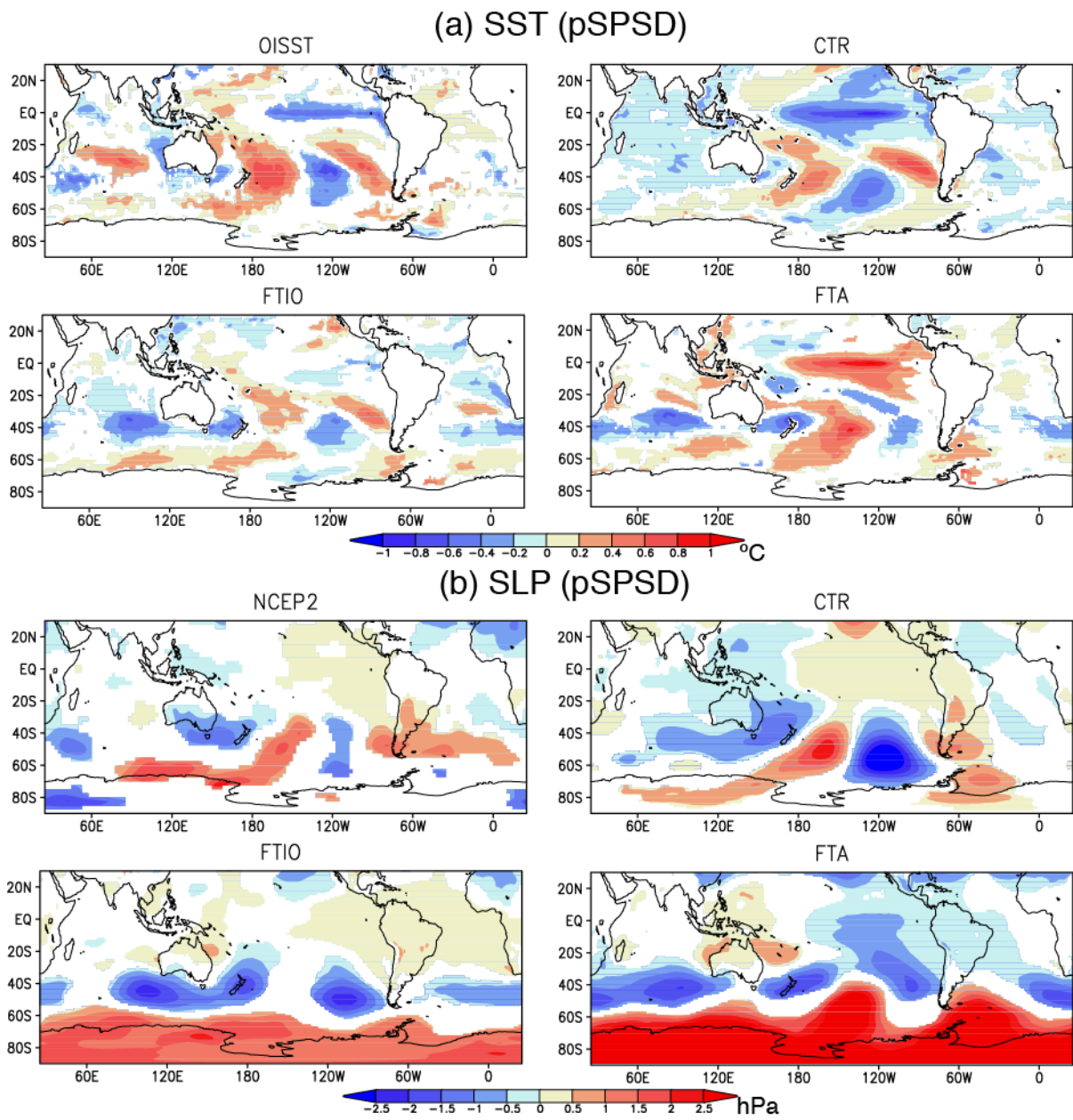
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868 **Figure 10:** As in Fig. 7, but for the positive SASD. The FTP and FTIO experiments are

869 shown.

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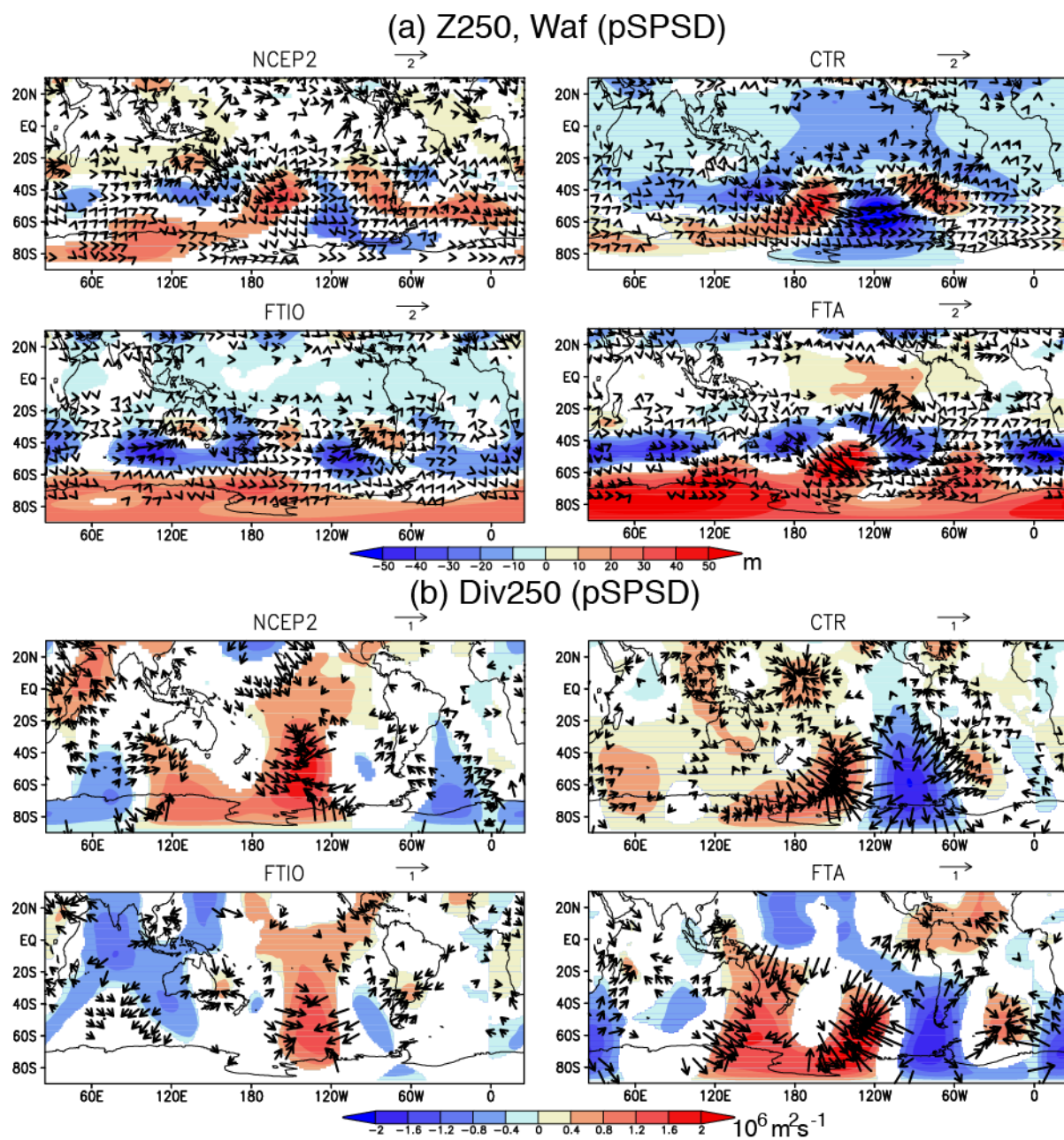
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873 **Figure 11:** As in Fig. 4, but for the positive SPSD. The FTIO and FTA experiments are
 874 shown.

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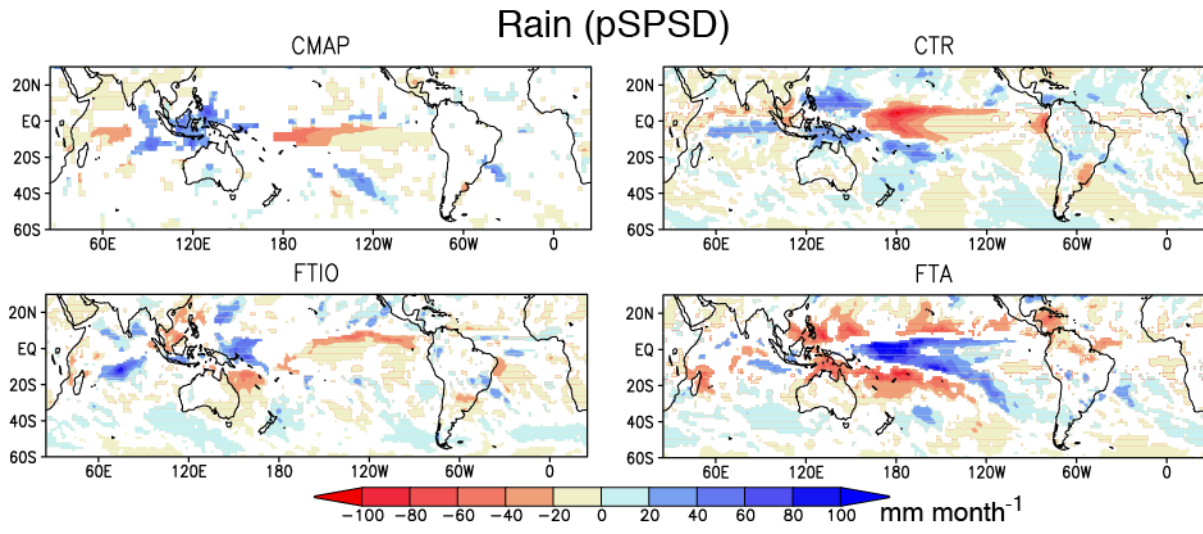


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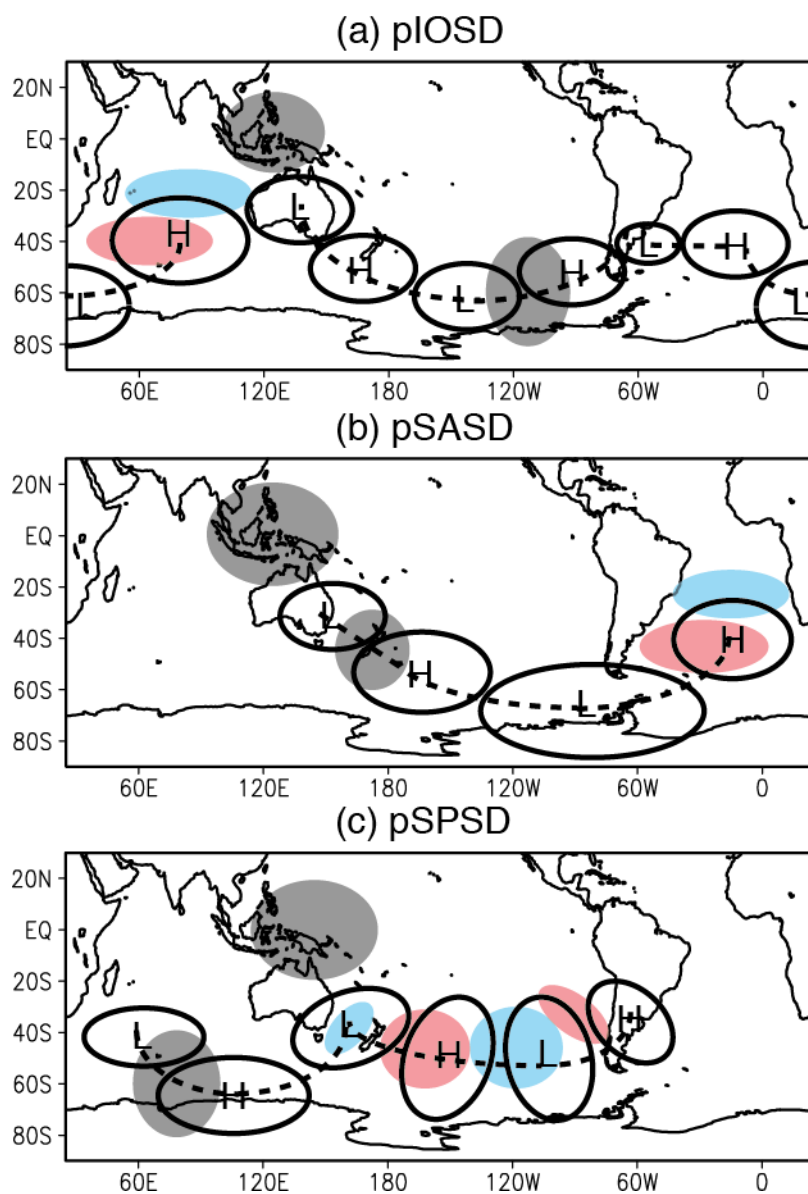
877 **Figure 12:** As in Fig. 6, but for the positive SPSSD. The FTIO and FTA experiments are
 878 shown.

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 882 **Figure 13:** As in Fig. 7, but for the positive SPSD. The FTIO and FTA experiments are
 883 shown.
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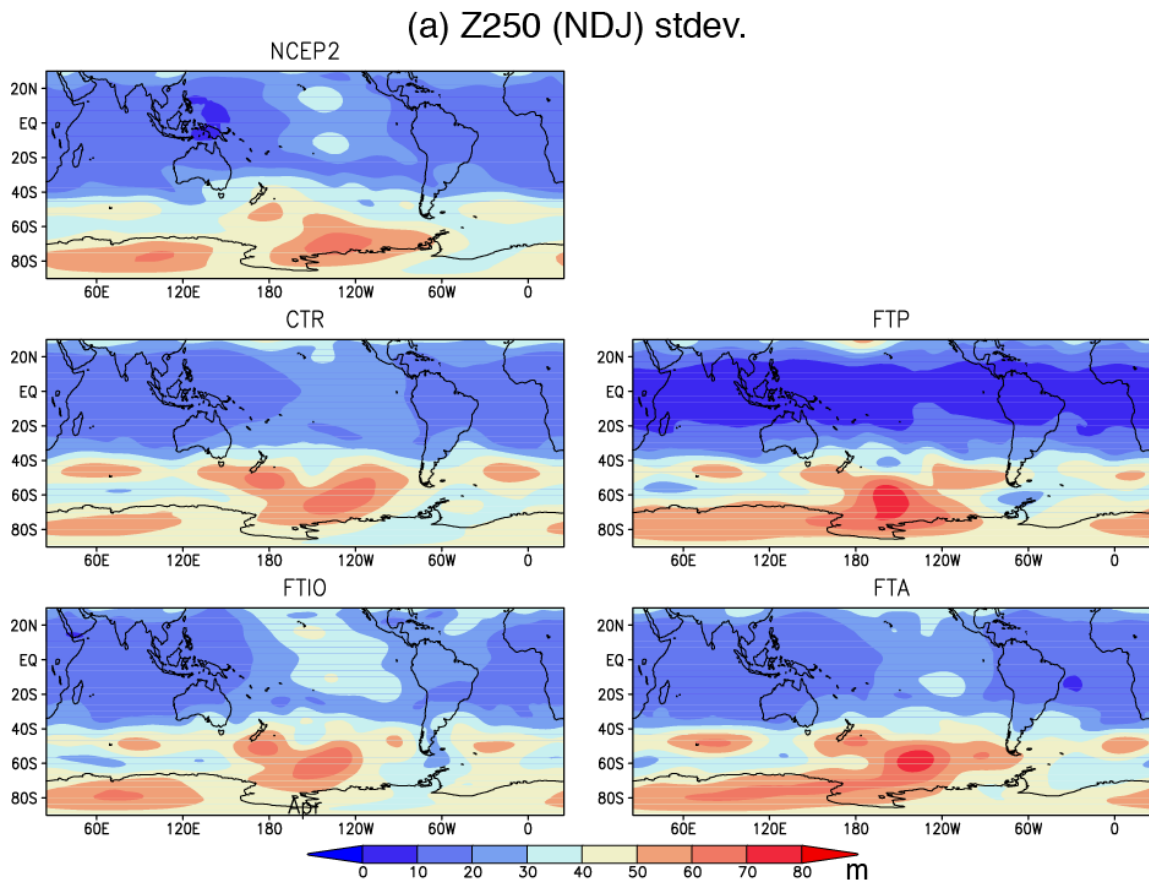
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886 **Figure 14:** Schematic diagram illustrating the generation mechanism of the IOSD, SASD,
 887 and SPSD derived from the reanalysis data and the CTR experiment. The warm (cold) color
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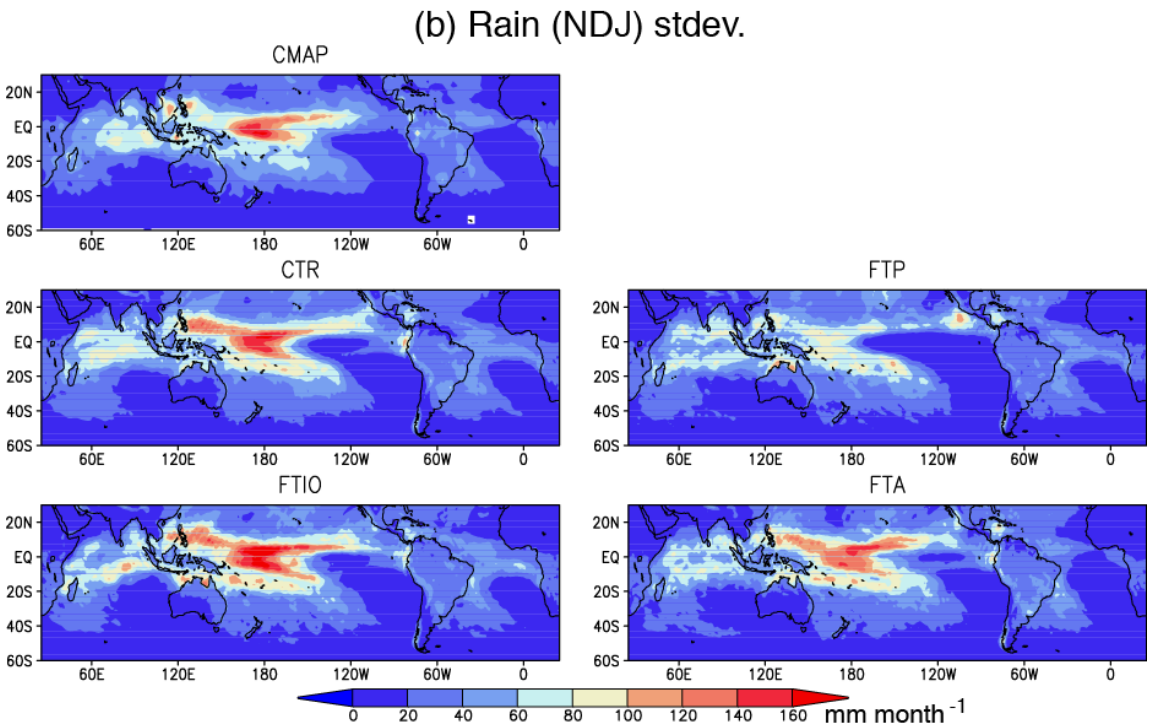
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Figure 15: As in Fig. 1, but for (a) geopotential height at 250 hPa (in m) during Nov.-Jan.. (b)

897

As in (a), but for the precipitation (in mm month⁻¹).