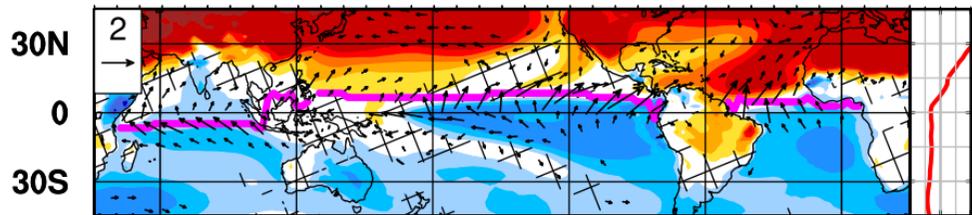


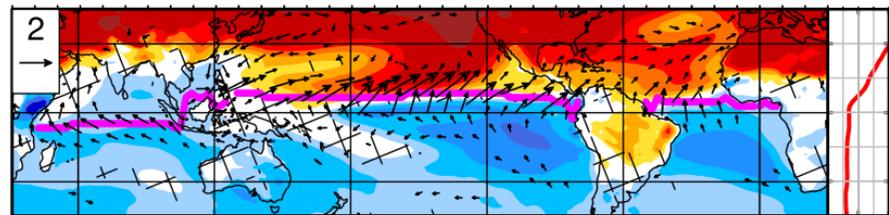
Figure 1.

a) hNA-SOM ΔT pattern

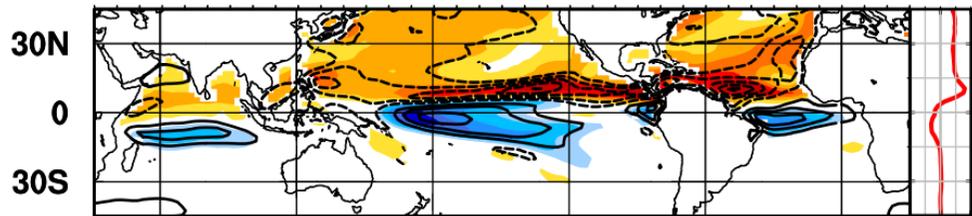
mean: 0.44 K

**b) hNP-SOM ΔT pattern**

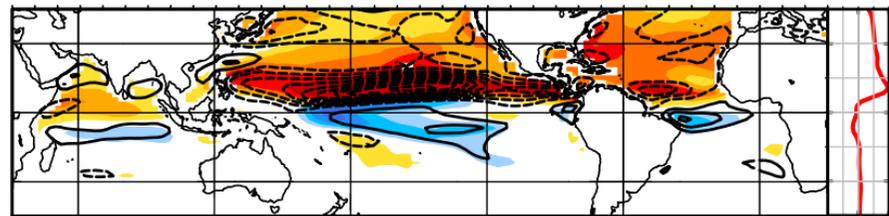
mean: 0.62 K

**c) hNA-SOM $\Delta T_{LH,a}$**

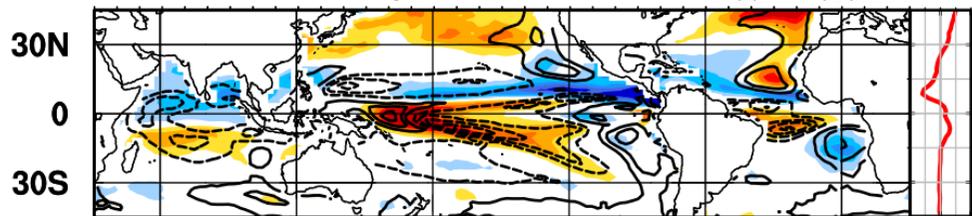
mean: 0.14 K

**d) hNP-SOM $\Delta T_{LH,a}$**

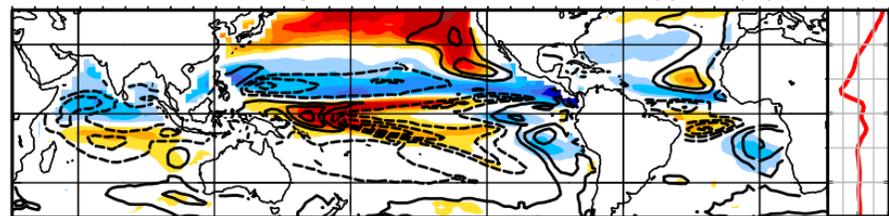
mean: 0.24 K

**e) hNA-SOM ΔT_{sw}**

mean: -0.02 K

**f) hNP-SOM ΔT_{sw}**

mean: -0.03 K



60E 120E 180 120W 60W 0

60E 120E 180 120W 60W 0

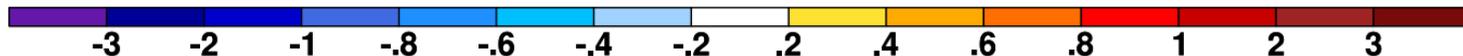


Figure 2.

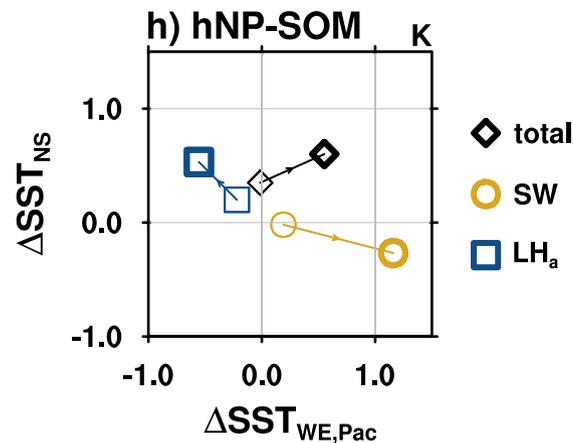
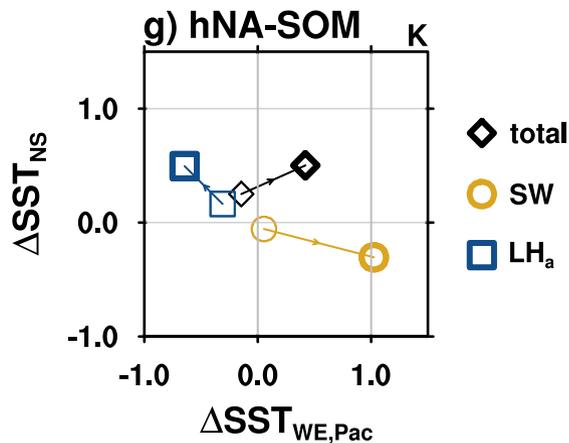
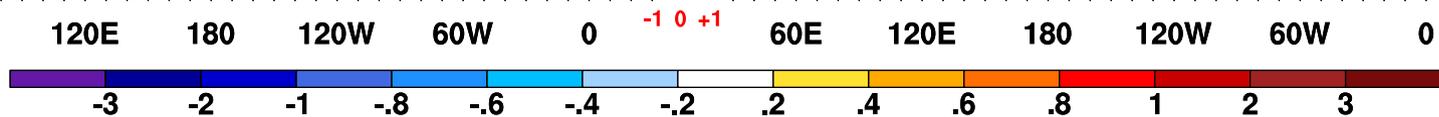
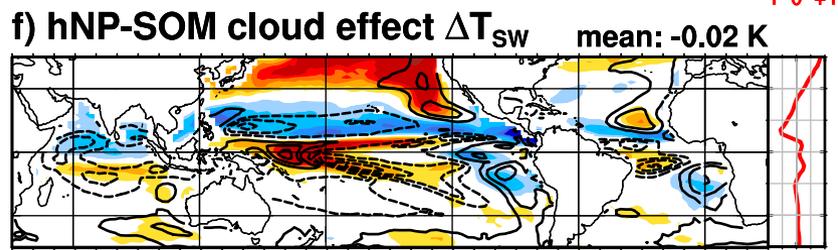
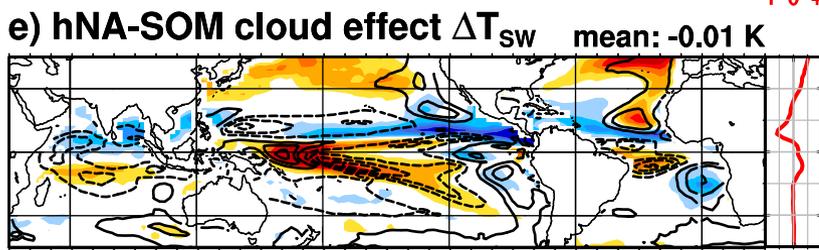
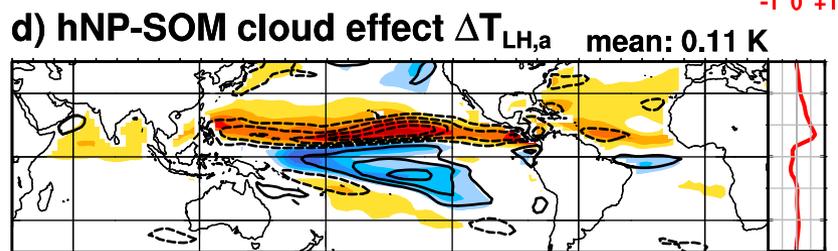
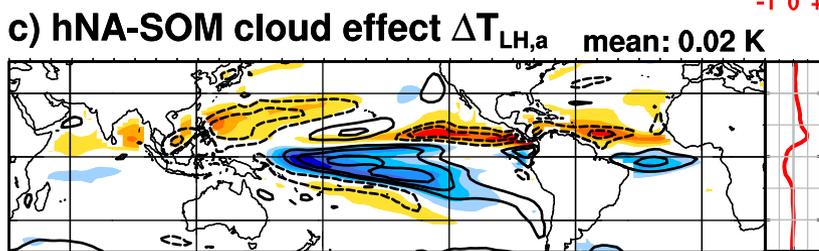
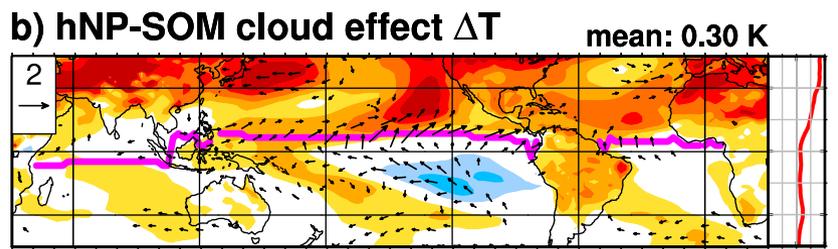
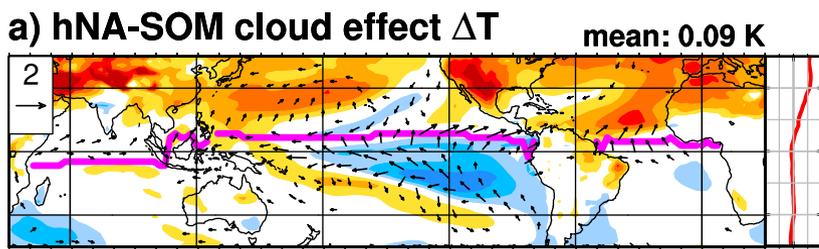
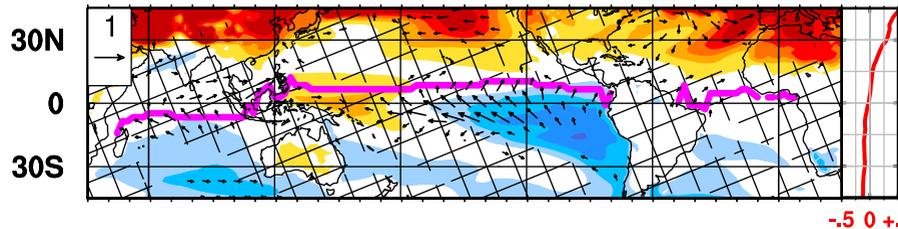
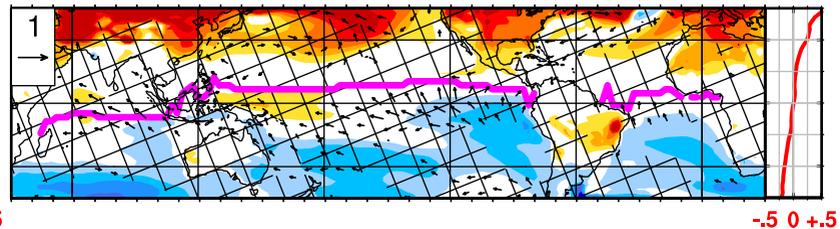


Figure 3.

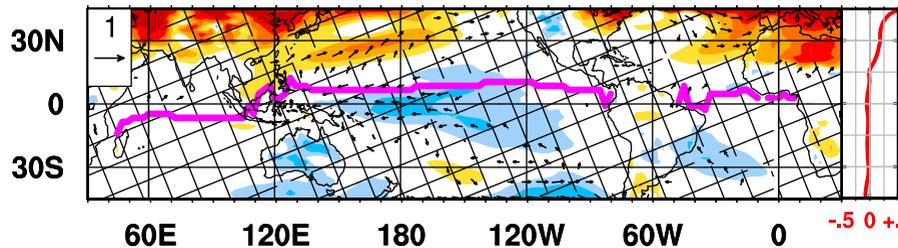
a) hNA-FOM ΔT pattern (y01-10, N=8) mean: 0.09 K



b) hNA-FOM ΔT pattern (y21-30, N=3) mean: 0.21 K



c) hNA-FOM-CL ΔT pattern (y01-10, N=5) mean: 0.08 K



d) cloud effect over y01-10: (a)-(c) mean: 0.01 K

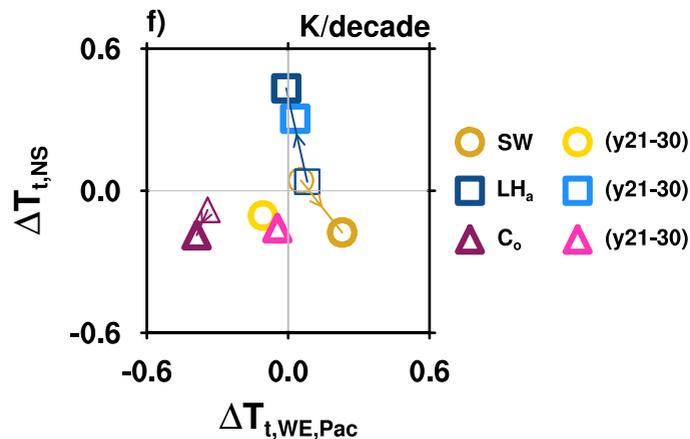
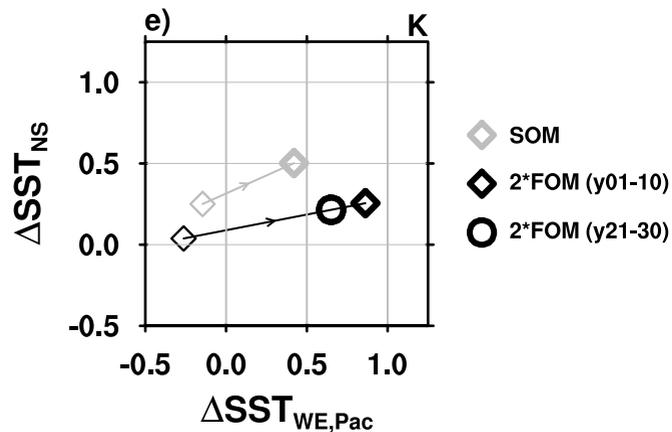
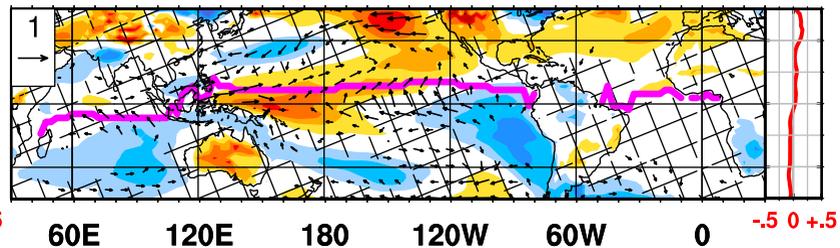
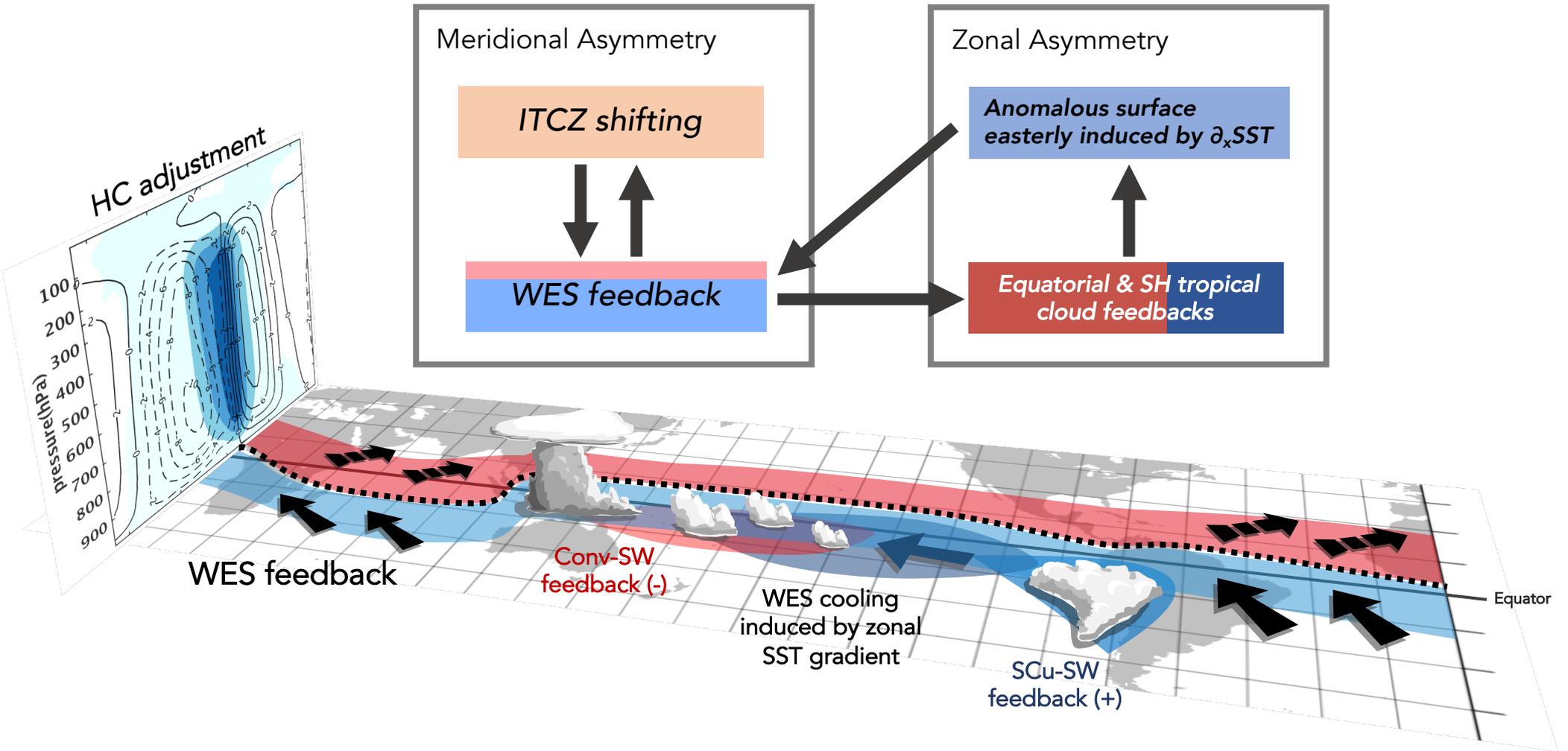


Figure 4.



Abstract

Extratropical influences on tropical sea surface temperature (SST) have implications for decadal predictability. We implement a cloud-locking technique to highlight the critical role of clouds in shaping tropical SST response to extratropical thermal forcing. With heating imposed over either extratropical Northern Atlantic or Pacific, Hadley Cells respond similarly that the trades strengthen south of the rainband. The wind-evaporation-SST (WES) feedback leads to cooling over the southern subtropics, which is enhanced in the southeastern Pacific due to the positive feedback between SST and stratiform clouds. This cooling is further extended toward the central Pacific via a WES effect associated with zonally contrasting cloud-radiative-SST feedbacks in the tropics, which is observed in both slab-ocean and dynamical-ocean experiments. We propose that the meridional and zonal SST gradients are tightly linked via WES effects and the cloud-radiative-SST feedbacks, which are largely determined by the climatological rainband position and the spatial distribution of cloud properties.

Plain Language Summary

Tropical sea surface temperature could induce changes in global climate patterns, and could be impacted by climate perturbation at mid-to-high latitudes on decadal timescales. Understanding how the tropics and extratropics interact could help predict the decadal evolution of the climate. We examine such interaction with a global climate model. Common tropical climate response patterns to the heating imposed in either extratropical Northern Atlantic or Pacific are found, determined by the following mechanisms. First, the wind speed of the trades responds in opposite manners with respect to the mean-state tropical rainband, where stronger trades and surface evaporation is present over the equator and its south that cools the sea surface. Second, this equatorial cooling is stronger over the eastern Pacific, which is associated with the mean-state properties of how clouds interact with changing surface temperature, that convective clouds over the western and central Pacific tend to damp the change, while stratiform clouds over the eastern Pacific act to amplify the change. These mechanisms, which are largely established by the mean-state rainband position and the spatial distribution of cloud properties, act to link the zonal and meridional structure of the sea surface temperature over the tropical Pacific.

1 Motivation

Tropical Pacific sea surface temperature (SST) has far-reaching climate impacts with its direct influences on atmospheric convection and thus planetary-scale stationary Rossby waves. For example, the La Niña-like cooling trend during 1998-2012 is suggested to affect the global distribution of rainfall and the occurrences of tropical storms (Delworth et al., 2015; Zhao et al., 2018). Through its critical modulation of the vertical structures of tropical atmospheric temperature and thus the efficiency of radiative climate feedbacks, the temporal evolution of tropical Pacific SST pattern is reported to control the climate sensitivity in observational records and modeling simulations (e.g. Andrews et al., 2015; Ceppi & Gregory, 2017; Dong et al., 2019), often referred to as “the pattern effect” (Stevens et al., 2016).

With the increasing demands for decadal climate predictability upon anthropogenic influences, driving mechanisms behind tropical Pacific SST trends under warming scenarios have been proposed, but have yet been comprehensively understood (Collins et al., 2018; Xie, 2020). The role of ocean dynamics in modulating the SST pattern on diverse timescales is well recognized via a coupling between the oceanic temperature structures and oceanic currents driven by atmospheric wind stress (Clement et al., 1996; Heede et al., 2020). In this study, the air-sea heat flux is of interest since recent literature has shown their major roles in affecting the SST under anthropogenic climate change. For

instance, the spatial structure of evaporation, which is largely determined by surface winds, shapes the anomalous SST pattern with increased CO₂ (L. Wang & Huang, 2016; Xie et al., 2010). Cloud radiative feedbacks, which are established by cloud properties associated with atmospheric circulations and the SST pattern itself, could also impose radiative flux anomalies and SST tendencies over the tropical Pacific (DiNezio et al., 2009).

Recently, the non-local extratropical effects in driving tropical Pacific SST trends have received increasing attention. Couplings between atmospheric adjustments, surface fluxes, and oceanic tunnels could translate anomalies in the extratropics into tropical Pacific SST changes (Amaya et al., 2019; Luo et al., 2017; Shin et al., 2021; Hwang et al., 2021). Localized energy perturbations are commonly present in the extratropics, such as those due to anthropogenic aerosol emissions and polar sea ice losses. Localized forcing in the extratropics have been found to be homogenized zonally before propagating through the atmosphere into the tropics by subtropical eddies, making tropical response insensitive to the exact locations of the forcing (Kang et al., 2014, 2018; L'Hévéder et al., 2015). In particular, Kang et al. (2018) proposed that in a slab-ocean climate model, a common La Niña-like response pattern appears along with the northward displacement of the ITCZ in response to Northern Hemispheric differential heating at different locations in the extratropics.

This study investigates the formation mechanisms behind such La Niña-like tropical SST pattern response to localized extratropical thermal forcing from Kang et al. (2018). Later in this paper, we show that the formation of such pattern is established by the climatological rainband position and the climatological spatial pattern of cloud radiative feedbacks (Section 3). This is supported by cloud-locking experiments and is relevant in both models with a slab ocean and a fully coupled dynamical ocean (Section 4). We suggest that these climatological controls on tropical Pacific SST response to extratropical forcing could be applied to other forcing scenarios and offer potential source of predictive skills of the SST on decadal timescales (Section 5).

2 Methodology

2.1 Experimental Design

Sets of a control and two perturbed experiments are performed with Community Atmosphere Model 5.0 (CAM5; Neale et al., 2010) with Community Earth System Model 1.2 (Hurrell et al., 2013). Anomalous surface flux convergence is imposed into the equilibrated preindustrial simulations (CTL) between 45°N and 65°N zonally uniformly in either the North Atlantic (hNA) or North Pacific (hNP), each with a total amount of 0.41 petawatts. The climate responses to heating are defined as the time-averaged fields from either heating deducted by those from CTL. See Table S1 and Text S1 for a full list of experiments and a detailed explanation of the experimental design.

To investigate the SST formation mechanisms driven solely by surface fluxes, a slab-ocean model (SOM) with a constant mixed-layer depth of 50 meters is coupled to CAM5. Thirty years of data are analyzed after a spinning up of thirty-five years. To evaluate whether the proposed feedbacks shown in the SOM experiments remain important when oceanic processes are involved, experiments using a full-depth dynamical ocean model (FOM) are performed. The averaged hNA response of eight ensembles over years 1-10 and three ensembles over years 21-30 are analyzed. Fewer ensembles of the hNP experiments in FOM are also conducted but are only shown in Figure S1 for brevity, while their results are consistent with the main conclusions.

The other two sets of experiments are conducted to assess the direct and indirect impact of cloud radiative effects. A cloud-locking (CL) technique is used (Ceppi & Hartmann, 2016; Y.-J. Chen et al., 2021), in which hourly cloud optical properties from CTL are prescribed in CAM5. The SOM cloud-locking simulations are performed with prein-

115 dustrial condition (CTL) and with either of the heating. The control run is spun up for
 116 ten years while the perturbed runs are spun up for twenty years, and the following twenty
 117 years are analyzed in all runs. The FOM cloud-locking simulations are performed with
 118 only the North Atlantic heating and with five ensembles running out to ten years. We
 119 define the *cloud effect* by subtracting the responses in experiments with locked clouds
 120 from those with interactive clouds.

121 2.2 Energetic decomposition of SST response

122 An energy budget analysis of oceanic mixed layer is used to attribute SST response
 123 to surface energy fluxes (Xie et al., 2010; L. Zhang & Li, 2014). Briefly, the time ten-
 124 dency of the mixed-layer temperature (T) can be written as:

$$125 T_t = \frac{1}{\rho c_p H} (Q_{SW} + Q_{LW} + Q_{LH} + Q_{SH} + Q_{C_o}) \quad (1)$$

126 where the subscript t denotes a time derivative, ρ the density of sea water, c_p the spe-
 127 cific heat capacity of sea water at constant pressure, H the mixed-layer depth, and Q
 128 the inward energy flux. Fluxes include shortwave (SW) and longwave (LW) radiative fluxes,
 129 latent heat flux (LH), sensible heat flux (SH), and column-integrated heat convergence
 130 by oceanic heat transport (C_o). T_t contributed by each energy flux term can be writ-
 131 ten as:

$$132 T_{t,flux} = \frac{1}{\rho c_p H} Q_{flux}, \quad flux = \{SW, LW, LH, SH, C_o\} \quad (2)$$

133 In FOM experiments, $T_{t,flux}$ are calculated to show the contribution of fluxes to the trend
 134 of T with an averaged H used. In SOM experiments, Q_{C_o} is unchanged by design and
 135 the system reaches equilibrium thus $T_t = 0$. The temperature difference between the
 136 control state and the forced states in SOM runs can then be written as:

$$137 \Delta T_{flux} = \frac{1}{\overline{dQ/dT}} \Delta Q_{flux}, \quad flux = \{SW, LW_{dn}, LH_a, SH\} \quad (3)$$

138 where Δ denotes the differences between the forced and the control states, and $\overline{dQ/dT}$
 139 is the linear dependence of total surface energy flux to T evaluated at the arithmetic-
 140 mean states between the forced and the control states, which consists of a blackbody long-
 141 wave radiative term and a latent heat term associated with its bulk formula. After re-
 142 moving the linear dependent terms to T , downward longwave radiative flux (LW_{dn}) and
 143 non-Newtonian latent heat flux that depends solely on near-surface atmospheric condi-
 144 tion (LH_a) appear that replace LW and LH, respectively. Finally, ΔT_{flux} are calculated
 145 to show the contribution of the change in fluxes to the change in T . See Text S2 for a
 146 detailed derivation and expression of each term.

147 3 The climatological controls of the response patterns

148 We first present the responses in SOM experiments. Regardless of heating being
 149 imposed in either North Atlantic or North Pacific, steady-state surface temperature re-
 150 sponse patterns are highly similar, with a spatial correlation of 0.86 within 30°S and 30°N
 151 (comparing Figure 1a with 1b). Both the interhemispheric and zonal SST gradients strengthen
 152 along the equatorial Pacific. Responding to the imposed heating in the Northern Hemi-
 153 sphere, an anomalous cross-equatorial overturning circulation develops to transport en-
 154 ergy southward through its upper branch and to shift moisture and the ITCZ northward
 155 through its lower branch (shown by the low-level wind response in Figures 1a-b). The
 156 robust similarities of the tropical responses are consistent with Kang et al. (2018) who
 157 proposed that extratropical forcing is being homogenized before affecting the tropics.

158 First, we discuss the formation mechanism behind the anomalous meridional SST
 159 structure, which is largely established by the climatological position of the tropical rain-
 160 band (pink lines in Figures 1a-b). The climatological rainband is located at the transi-
 161 tion region between the northeast and the southeast trades. With the rainband shifts

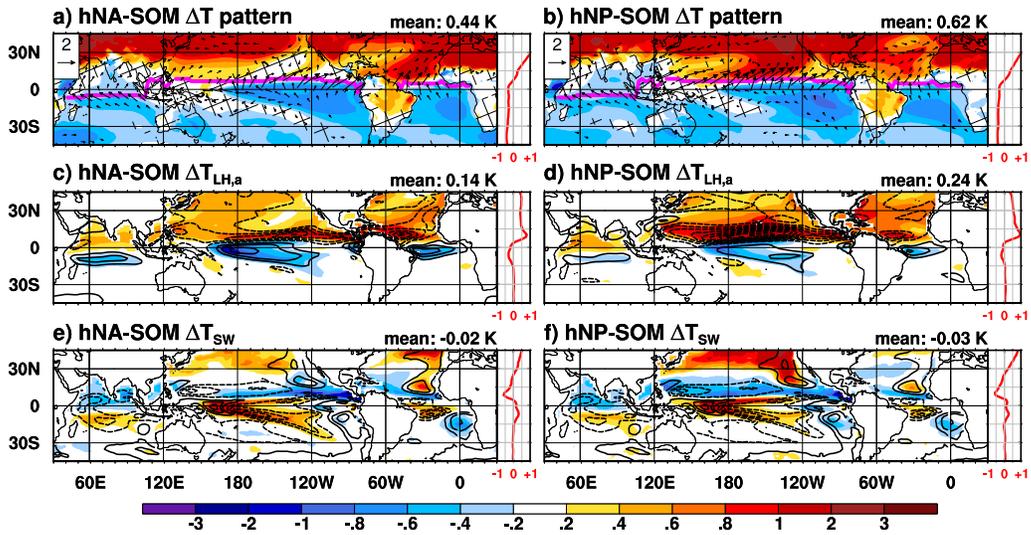


Figure 1. (a-b) responses of surface temperature pattern (shading; K) and 993-hPa winds (arrows), with climatological ITCZ defined at where the annual rainfall on marine region has meridional maxima (pink line), and the hatched regions show where the shading patterns are statistically insignificant at 5% level using two-tailed Student *t*-tests. In (c-d), shading shows $\Delta T_{LH,a}$ (units: K) and contours show the response of 10-meter wind speeds (spacing: 0.2 m s^{-1}). In (e-f), shading shows ΔT_{SW} (units: K) and contours show the surface SW-SST feedbacks estimated as the slope from the linear regression of surface shortwave radiative flux on SST using CTL-SOM (spacing: $5 \text{ W m}^{-2} \text{ K}^{-1}$). The patterns of responses are calculated by removing their means over 30°S - 30°N . Mean values of the shadings over 30°S - 30°N are labeled, and zonal mean values of the shadings are attached on the right of each panel. For contours, dashed indicates negative, solid indicates positive, and zero lines are omitted.

162 toward the north in response to the Northern Hemispheric heating, anomalous low-level
 163 southerlies dominate over the tropics, which strengthen the southeasterlies on the south
 164 of the rainband and result in increased evaporation (negative $\Delta T_{LH,a}$) and cooler SST,
 165 while the opposite is shown in the north. This dipole pattern of $\Delta T_{LH,a}$ dominates the
 166 response of the anomalous meridional SST structure, while it is partly damped by the
 167 shortwave radiative effect (ΔT_{SW} ; Figures 1e-f) associated with the shift of the rainband.
 168 As the rainband shifts toward the warmer Northern Hemisphere, the response of cloud
 169 cover and shortwave reflectivity are anti-symmetric to the climatological rainband, lead-
 170 ing to cooling on the north and heating on the south.

171 The role of climatological rainband in determining the center of the meridional dipole
 172 SST response is consistent with previous literature (e.g. Chiang & Bitz, 2005), known
 173 as the ITCZ blocking effect, stating that anomalous warming is not able to penetrate
 174 the climatological rainband and could even induce an anomalous cooling south of the rain-
 175 band. Kang et al. (2020) further demonstrate that the climatological ITCZ location acts
 176 as a barrier to the propagation of the surface temperature anomalies in response to ex-
 177 tratropical forcing using a set of aqua-planet experiments. This effect produced by the
 178 climatological rainband is also shown in natural variabilities, as H. Zhang et al. (2014)
 179 suggest that the anomalous SST and surface winds associated with the South Pacific Merid-
 180 ional Mode no longer reach the equatorial region under a climatological state with the
 181 rainband located south of the equator.

182 Next, we turn our attention to the mechanism behind the enhanced zonal SST gra-
 183 dient over the equatorial Pacific, which is established by the spatial structure of clima-
 184 tological cloud regimes. The response of cloud cover under different environments to the
 185 latent heat cooling across the equatorial region shapes the zonal structure of the SST
 186 response. Shown as the contours in Figures 1e-f, shortwave-SST feedback over the trop-
 187 ics is positive over the east of the basins where stratiform clouds prevail and is negative
 188 along the convective ITCZ and the South Pacific Convergence Zone. With increased low-
 189 level atmospheric stability followed by the decrease in SST, marine stratiform clouds in-
 190 crease over the eastern equatorial Pacific (Hanson, 1991; Klein & Hartmann, 1993). As-
 191 sociated with the anomalous cross-equatorial Hadley Cell, the strengthened subtropical
 192 high in South Pacific may also produce anomalous cold advection and further increase
 193 the marine stratiform clouds (Wei et al., 2018). In contrast, the decreased SST and strength-
 194 ened subsidence both inhibit deep convections and anvil clouds over the western-central
 195 tropical Pacific, allowing more solar insolation to heat up the surface (Lau et al., 1997;
 196 Ramanathan & Collins, 1991; Meehl & Washington, 1996). Distinct cloud radiative-SST
 197 feedback in the eastern and western equatorial Pacific is recently highlighted in Park et
 198 al. (2022). These zonally contrasting cloud-radiative feedbacks appear to be the most
 199 dominant term that contributes positively to the enhanced zonal SST gradient over the
 200 equatorial Pacific (comparing ΔT_{SW} in Figures 1e-f with Figures 1a-b). The contribu-
 201 tions from longwave and sensible heat fluxes are at less importance in shaping the over-
 202 all SST response pattern (not shown), whereas $\Delta T_{LH,a}$ serves as the main damping term
 203 of the zonal SST gradient response.

204 4 The coupling between clouds and circulation

205 In Section 3, we suggest the mechanisms underlying the similar pattern of SST re-
 206 sponse over the tropical Pacific, considering their dependence on climatology. Namely,
 207 the northward displacement of the tropical rainband and the zonal contrast in the cli-
 208 matological cloud regimes together shape the tropical SST pattern response to the ex-
 209 tratropical thermal forcing. In this section, we use a cloud-locking technique (see Sec-
 210 tion 2) to verify the role of clouds in shaping the structure of the SST response.

211 The enhanced zonal SST gradient over the equatorial Pacific is mostly contributed
 212 by the *cloud effect* (as defined in Section 2), which is obtained by comparing simulations

213 with interactive and locked cloud radiative properties. As shown in the scatter plots in
 214 Figures 2g-h, the changes in the equatorial Pacific zonal SST gradient are close to zero
 215 or even negative in the cloud-locking simulations (thin black diamond), whereas they are
 216 largely positive in the cloud-interactive simulations (bold black diamond). The essen-
 217 tial role of clouds in shaping the equatorial Pacific zonal SST gradient is consistent with
 218 the surface energy budget analysis discussed in Section 3. ΔT_{SW} is the only term that
 219 contributes to the enhanced zonal SST gradient over the equatorial Pacific (yellow cir-
 220 cles), supporting the hypothesis that zonally contrasting cloud feedbacks enhance the
 221 SST gradient.

222 In addition to direct radiative effects, the cloud effect also manifests the non-radiative
 223 effects originated from the interaction between clouds and circulations. For the triangle-
 224 shaped surface cooling that maximizes in the southeastern Pacific and extends to the cen-
 225 tral equatorial Pacific, it is produced by two parts of cloud effects together. The south-
 226 eastern Pacific cooling is caused by the decrease in shortwave surface radiative flux as-
 227 sociated with the increase in stratiform clouds (Figures 2e-f), as mentioned earlier in the
 228 discussion of the zonal SST gradient; on the other hand, the central equatorial Pacific
 229 cooling could be attributed to the increase in evaporation due to cloud effects (Figures
 230 2c-d). We interpret the increased evaporation due to the change in clouds as part of the
 231 *indirect effects by clouds*. The increase in the zonal SST gradient associated with cloud
 232 and shortwave radiative flux changes over the tropical Pacific (Figures 2e-f) drives anoma-
 233 lous surface easterlies (arrows in Figures 2a-b) (Lindzen & Nigam, 1987) and leads to
 234 further increase in evaporation over the central equatorial Pacific (Figures 2c-d). Namely,
 235 the change in cloud cover ultimately drives the latent heat cooling over the central equa-
 236 torial Pacific. This increased evaporation as part of the cloud effect explains a large por-
 237 tion of the anomalous surface latent heat flux in the cloud-interactive simulations (com-
 238 paring Figures 2c-d with Figures 1c-d), and contributes to more than half of the enhance-
 239 ment of the interhemispheric SST gradient over the tropics (ΔSST_{NS} ; Figures 2g-h). In
 240 summary, direct cloud radiative effects contribute to the zonal structure of the anoma-
 241 lous SST, while indirect (i.e., non-radiative) cloud effects promote the meridional struc-
 242 ture of the anomalous SST.

243 One may suspect that the slab-ocean setting exaggerates the cloud effects due to
 244 lack of vertical mixing and circulation feedbacks in the ocean. It is verified in Figure 3,
 245 which shows the SST response to the North Atlantic heating with a fully interactive ocean
 246 model (FOM), that the importance of clouds can still be seen within the first few decades.
 247 The North Pacific heating experiments lead to similar conclusions (Figure S1). Despite
 248 weaker magnitudes, the SST pattern response in the FOM experiments is qualitatively
 249 similar to those in the slab-ocean experiments (compare Figure 3a with Figure 1a; also
 250 see Figure 3e). The triangle-shaped cooling over the southeastern and equatorial-central
 251 Pacific is statistically significant in the FOM experiments and is persistent till year 30
 252 (Figures 3a-b). This cooling can be attributed to the cloud effect (Figure 3d) and is ab-
 253 sent in the cloud-locking simulations over the first decade (Figure 3c). Shown by Fig-
 254 ure 3f, shortwave cloud radiative effect dominates the zonal gradient of equatorial Pa-
 255 cific SST tendencies over the first decade, while the cloud indirect effect of latent heat
 256 flux contributes primarily to the meridional gradient of tropical SST tendencies over both
 257 the first and the third decade. The oceanic heat convergent term always damps the gra-
 258 dients of the SST tendencies (purple triangles in Figure 3f). These results support the
 259 existence of our proposed mechanisms in a model setting with a fully interactive ocean,
 260 with a stronger confidence over the first decade after the forcing is imposed.

261 5 Summary and Discussion

262 Kang et al. (2018) and L'Hévéder et al. (2015) report that the tropical SST pat-
 263 tern is insensitive to the longitudinal location of extratropical thermal forcing in a slab-
 264 ocean setting. In this study, we investigate the formation mechanisms of such tropical

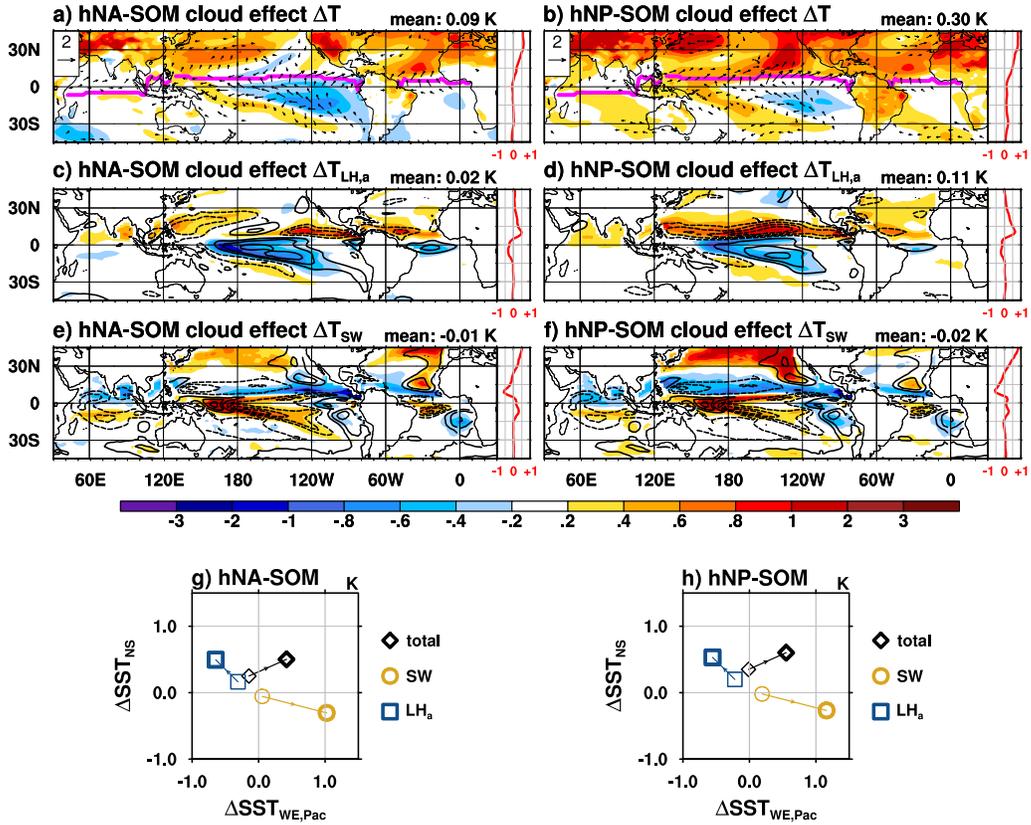


Figure 2. (a-f) as Figures 1(a-f), but of the cloud effect except for the SW-SST feedback in e-f (contours); (g-h) SST gradient metrics (see definition later in the caption) from the cloud-locking runs (thin markers) and the cloud-interactive runs (bold markers) associated with the surface fluxes labelled in the legend, where the arrows between the thin and the bold markers depict the cloud effect. The tropical meridional SST gradient (ΔSST_{NS}) is defined as the areal mean SST over 0°-20°N minus that over 20°S-0° across all longitudes; the equatorial Pacific zonal SST gradient ($\Delta SST_{WE,Pac}$) is defined as the areal mean SST over 160°E-180° minus that over 110°W-90°W within 5°S-5°N.

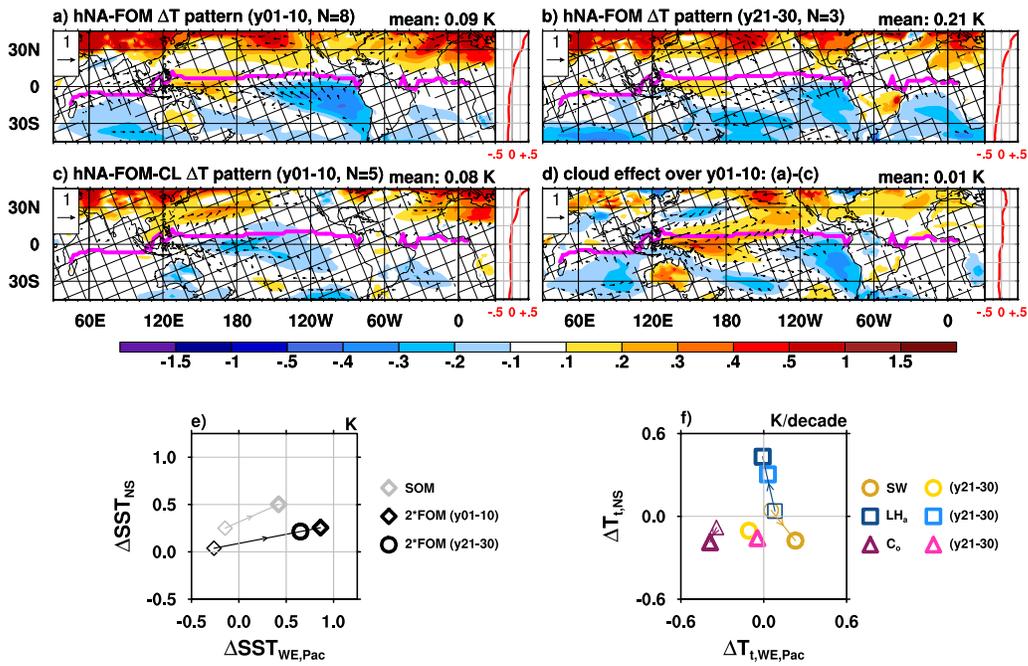


Figure 3. As Figure 1a, but of the ensemble means from the fully coupled simulations averaged over (a) years 1-10 and (b) years 21-30. (c) shows the response in the cloud-locking simulations over years 1-10 and (d) shows the corresponding cloud effect. (e) is as Figure 2e but only shows those for the overall SST gradients (fully coupled results are multiplied by two for clarity). (f) is also as Figure 2e but shows those for SST tendencies (T_t) instead of SST anomalies over years 1-10, while symbols with lighter colors represent the responses over years 21-30.

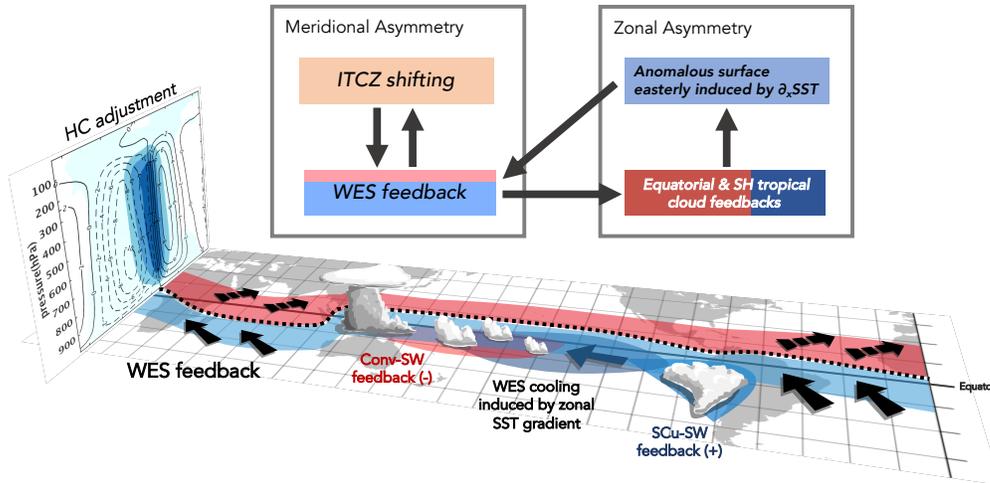


Figure 4. A schematic of the feedbacks behind the La Niña-like SST formation. Hadley Cell adjustment due to the forcing is shown on the left. On the map, the climatological ITCZ is indicated by the dashed line, with near-surface wind change (arrows) and associated SST change by the WES feedback (band-structured shading) when the ITCZ shifts north. Cloud-associated responses near the equator are also depicted with realistic style and its effects on SST response are as the shading patches. A flow chart of the discussed feedbacks is on the top.

265 SST pattern in CESM1. A schematic of our proposed mechanisms that lead to the com-
 266 mon SST pattern is displayed in Figure 4. First, responding to heating in extratropical
 267 Northern Hemisphere, an anomalous cross-equatorial Hadley Cell develops. On the south
 268 of the climatological rainband, the enhanced southeasterly trades lead to enhanced evap-
 269 orative cooling. With the cooling, zonally contrasting cloud cover responses emerge over
 270 the equatorial Pacific. As a result, zonal SST gradient and near-surface easterlies over
 271 the southern subtropical Pacific are strengthened. In other words, WES feedback and
 272 cloud-SST feedbacks together build up a positive feedback between the zonal and merid-
 273 ional structures of tropical SST changes over the tropical Pacific. This formation of the
 274 anomalous SST pattern is triggered when hemispheric differential heating is presented
 275 in the extratropics. While the energetic framework provides zonal-mean predictability
 276 under extratropical hemispheric differential heating (as summarized in Kang et al., 2020),
 277 combining our findings with the zonal homogenization of the forcing (Kang et al., 2018)
 278 yields spatial predictability of tropical responses.

279 The robust control of climatology fields to tropical SST response also implies that
 280 mean-state biases in CESM could lead to uncertainties regarding the mechanisms pro-
 281 posed in this study. However, although the mean-state biases may quantitatively influ-
 282 ence the magnitudes of SST response, the mechanisms discussed should be at work even
 283 without a perfect climatology. For example, the WES cooling on the south of the rain-
 284 band should be qualitatively similar among the GCMs with the ITCZ located on the north
 285 of the equator in most seasons. The mechanisms that link zonal to meridional SST gra-
 286 dient should be at work as long as the overall signs of the cloud-SST feedback pattern
 287 agree with observations over the tropics, which is true in most of the GCMs (X. Chen
 288 et al., 2019). As supportive evidence, Hwang et al. (2017) report a linear relationship
 289 between interhemispheric tropical SST asymmetry and zonal SST gradient over the south-
 290 ern subtropical Pacific in abrupt 4xCO₂ experiments in CMIP5 models, despite the pres-

291 ence of double-ITCZ biases and quantitatively inaccurate cloud-SST feedbacks in most
 292 of the models.

293 The importance of WES effects and cloud-SST feedbacks on extratropical-to-tropical
 294 teleconnections have been discussed extensively in the literature of seasonal footprint-
 295 ing mechanisms (Vimont et al., 2001), meridional modes (Chiang & Vimont, 2004), and
 296 central Pacific El Niño-Southern Oscillation (Yu et al., 2015), that anomalies of subtrop-
 297 ical trade winds could alter latent heat flux and thus local SST gradients, leading to the
 298 propagation of SST anomalies into the deep tropics. Our result highlights the critical
 299 role of clouds in shaping such teleconnection patterns. This message is consistent with
 300 recent modeling studies suggesting that the positive cloud-SST feedback over the east-
 301 ern basins could either amplify or drive SST variabilities on decadal timescales (Bellomo
 302 et al., 2014, 2015; Clement et al., 2015; Burgman et al., 2017; Middlemas et al., 2019).
 303 While literature of meridional modes mostly focuses on variabilities within individual
 304 oceanic basins, we demonstrate a cross-hemispheric influence of high-latitude heating
 305 on meridional mode-like response via the adjustments of Hadley Cells and the fol-
 306 lowing responses of surface fluxes and clouds.

307 The mechanisms we reported here linking meridional to zonal SST gradient largely
 308 depend on surface fluxes in the subtropics, which is different from the mechanisms re-
 309 lated to oceanic upwelling over the eastern equatorial Pacific proposed by previous lit-
 310 erature (e.g. Chiang et al., 2008; Fedorov et al., 2015). It is worth noting that our FOM
 311 simulations are only run out to thirty years after the system being perturbed, and oceanic
 312 processes on longer timescales might take place and alter the response patterns if we ex-
 313 tend the simulations (e.g. Kang et al., 2020; K. Wang et al., 2018), which is partly shown
 314 in Figure S1b. Our result at least suggests that within thirty years, when only the “fast”
 315 responses in FOM simulations are present, the mechanisms discussed are robust. The
 316 interaction between circulation, surface turbulent fluxes, and clouds should be empha-
 317 sized when predicting, say, the changes in climate patterns after a few decades given cer-
 318 tain anthropogenic climate change. The results also suggest that better representation
 319 of cloud properties over the tropics is essential for modeling the associated mechanisms
 320 and could potentially improve the predictive skill of tropical SST pattern on decadal timescales.

321 Open Research

322 CESM 1.2 can be downloaded and installed following the official documentation
 323 provided by NCAR and UCAR: <https://www.cesm.ucar.edu/models/cesm1.2/>. To
 324 reproduce the simulations, follow the details described in Text S1, including the model
 325 settings, the spatial structures of the extratropical heating, the imposed form of heat-
 326 ing, and the procedures of how to save and impose cloud optical properties. The list of
 327 simulations is provided as Table S1.

328 Acknowledgments

329 We appreciate Dr. Paulo Ceppi for kindly providing the source codes for cloud locking
 330 in CESM1. We are also grateful to Prof. Shang-Ping Xie for his thoughtful comments.
 331 Y.-T. Hwang and Y.-J. Chen are supported by the Ministry of Science and Technology
 332 of Taiwan (MOST 110-2628-M-002-002).

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