ABSTRACT: The Maritime Continent (MC), located in the heart of the Indo-Pacific warm pool, plays an important role in the global climate. However, the future MC climate is largely unknown, in particular the ENSO–rainfall teleconnection. ENSO induces a zonal dipole pattern of rainfall variability across the Indo-Pacific Ocean, that is, positive variability in the tropical Pacific and negative variability toward the MC. Here, new CMIP6 models robustly project that, for both land and sea rainfall, the negative ENSO teleconnection over the MC (drier during El Niño and wetter during La Niña) could intensify significantly under the Shared Socioeconomic Pathway 5.85 (SSP585) warming scenario. A strengthened teleconnection may cause enhanced droughts and flooding, leading to agricultural impacts and altering rainfall predictability over the region. Models also project that both the Indo-Pacific rainfall center and the zero crossing of dipole-like rainfall variability shift eastward; these adjustments are more notable during boreal summer than during winter. All these projections are robustly supported by the model agreement and scale up with the warming trend.

KEYWORDS: Maritime Continent; ENSO; Climate change

1. Introduction

The Maritime Continent (MC), uniquely positioned between the tropical Pacific (TP) and Indian Ocean (IO), is vital to the global climate (Neale and Slingo 2003; Ramage 1968). It has been challenging for climate models to satisfactorily simulate the atmospheric processes within the MC (King and Vincent 2018; Raghavan et al. 2018). Mountainous islands across the MC create many small-scale processes that could contradict the large-scale changes over the MC domain, which motivated the development of high-resolution regional models (Birch et al. 2016; Rauniyar and Walsh 2013). The MC is a region lacking coverage of in situ observations, which adds difficulties to the regional model development. In recent years, collaborative efforts have been made to better investigate the MC weather and climate, for example, the Years of the Maritime Continent project (Yoneyama and Zhang 2020).

MC is a region with potential negative impacts due to climate change (Collins et al. 2013). However, changes in the MC climate variability are not as well studied as the TP variability. The MC region is governed by the interplay of climate drivers across different time scales (Jourdain et al. 2013). In this study, we will address how ENSO influences the MC rainfall via teleconnection and focus on its changes under warming. We will investigate two aspects of future changes: magnitude change and spatial adjustment. Understanding future changes in regional rainfall variability can help MC communities building adaptation strategies and managing risks under climate change.

ENSO has worldwide impacts on precipitation variabilities (Ropelewski and Halpert 1987; Hendon 2003). Across the Indo-Pacific Ocean, El Niño induces a zonal dipole pattern of precipitation variability, that is, positive variability in the TP and “horseshoe”-shaped negative variability toward the MC (Langenbrunner and Neelin 2013). That is, the TP becomes wetter than normal, whereas the MC becomes drier. Physically, ENSO–rainfall teleconnection over the MC is part of the ENSO-induced circulation responses over the tropics (Wang et al. 2003; Lau and Nath 2003; Stuecker et al. 2015). In boreal summer when El Niño develops, a sequence of evolution begins with the eastward shifting of Walker circulation due to the anomalous warming in the eastern Pacific. The shift suppresses convection over the MC (also weakens Asian–Australian monsoon) and enhances convection in the central Pacific.

Theoretically, the suppressed convection over the MC excites an equatorial symmetric Rossby wave and subsequently forms twin low-level anticyclones (ACs) residing on each side of the equator and to the west of the heat sink (Matsuno 1966; Gill 1980). However, the two anticyclones are remarkably
asymmetrical about the equator as a result of monsoon background flows (Wang et al. 2003). One AC is located over the south Indian Ocean (SIO), which dominates during El Niño development. The other is in the western North Pacific (WNP) (Stuecker et al. 2015), which dominates during the mature and decay phases of El Niño. Wang et al. (2003) showed that the sustaining and amplification of these anticyclones depend critically on the local atmosphere-ocean interaction in the presence of favorable background monsoon flows. Suppressed convection initially establishes over the MC in JJA(0), rapidly intensifies and expands in September–November (SON)(0), and then moves northeastward to the Philippine Sea from D(0)/JF(1) to JJA(1).

Juneng and Tangang (2005) and Supari et al. (2018) showed how El Niño–induced MC rainfall variability evolves northeastward from JJA to March–May (MAM), modulated by the two AC systems over the SIO and WNP. The southern part of the MC (such as Sumatra, Java Island, and southern Borneo) is affected significantly during JJA, while the other areas are more pronounced during the DJF season (e.g., northern Borneo and the southern Philippines). Haylock and McBride (2001) reported that the Indonesian rainfall variability due to ENSO is spatially coherent during the JJA–SON, but spatially incoherent during DJF when there is no significant correlation with ENSO. Malaysia generally experiences dry anomaly conditions during El Niño. Tangang and Juneng (2004) reported that the Malaysian rainfall tends to be well correlated with ENSO during the SON and DJF, and they traced the strong relationship between ENSO and DJF Malaysia rainfall to the anomalous AC circulation over northern Borneo and the southern Philippines. CMIP3 and CMIP5 models robustly projected that over the central-eastern Pacific, the ENSO-induced rainfall variability strengthens (Bonfils et al. 2015; Perry et al. 2017; Yeh et al. 2018; Power et al. 2013; Kug et al. 2010; Chung and Power 2014; Chung et al. 2014; Chung and Power 2016, 2015). Here, we will make use of the newest CMIP6 models to investigate the changes in the ENSO teleconnection over the MC. We will discuss the relationship between changes in the AC and those in the TP. Although we focus on the MC domain, we aim to obtain a large-picture understanding of the MC-TP climate system. Beside the intensification, CMIP3 and CMIP5 models robustly project that, over the TP, ENSO-induced rainfall variability shifts eastward under warming (Yeh et al. 2018; Taschetto et al. 2020; Yan et al. 2020; Coelho and Goddard 2009; Huang and Xie 2015; Bayr et al. 2014; Kug et al. 2010; Power et al. 2013). Previous studies mainly looked at the longitude of the maximum variability over the TP (Kug et al. 2010; Power et al. 2013; Bayr et al. 2014). Here, we use CMIP6 models to investigate whether the zero crossing longitude of the zonal-dipole ENSO teleconnection will shift eastward under warming.

Unlike the TP, where the observed ENSO-induced precipitation variability has a large spatial coherence, precipitation variability over the MC has its spatial inhomogeneity due to a complex land–sea layout and interplay of processes (Lee and McBride 2016; Tangang et al. 2018). Previous studies noticed that Indonesian monsoon rainfall and ENSO have a low correlation (Haylock and McBride 2001; McBride et al. 2003), which was found due to the offsetting of the positive−negative correlation in east–west Indonesia. Here, we carry out a few treatments to analyze the MC. We define a central MC (MCM) domain that covers the coherent negative ENSO teleconnection. We also define an eastern MC domain (EMC) that has positive rainfall variability during ENSO. We use this EMC to highlight the eastward shift of the ENSO teleconnection. Within the MC domain, land covers a small percentage. We will investigate whether the changes in the ENSO teleconnection are applicable to both land and sea rainfall variability. Along with ENSO evolution, rainfall variability in varying subdomains of the MC may have different peak seasons and associated impacts (Juneng and Tangang 2005). Here, we will investigate whether future changes also show seasonal variations. In the observation, El Niño− and La Niña−induced teleconnections over the MC show asymmetry (Tangang et al. 2017). Here, we will investigate whether such discrepancy continues as to the future changes. To better understand the zonal shift in the teleconnection, we will also link it to the zonal adjustment in the mean rainfall.

We analyze CMIP6 models in the historical period and in the Shared Socioeconomic Pathway 5−8.5 (SSP585) warming scenario. To build confidence in the projections, we carry out model performance evaluation and diagnose individual model behaviors. We investigate model agreement on the changes. We also analyze changes over time and investigate how changes scale up with the warming trend despite natural variations. Recent studies (e.g., Hausfather et al. 2022) raised the “hot model” issues for CMIP6 models. We thus separate models into three groups (i.e., hot, median, and cold models) to provide more realistic projections matching the likely warming range by IPCC AR6.

The paper is organized as follows. In section 2, we overview data and methods. In section 3, we show the results of the observed ENSO teleconnection and its future projections by CMIP6 models. In section 4, we discuss potential impacts of changes in the ENSO teleconnection. In section 5, we discuss how our results compare with previous studies and the new understanding. Finally, we summarize and conclude in section 6.

2. Data and methods

a. Datasets

For observations, we analyze the sea surface temperature from the HadISST dataset (1870−2017) (Rayner et al. 2003) and the precipitation from the GPCP dataset (1979−2017) (Adler et al. 2003). We use the recent 35-yr historical period (1980−2014) for further analysis. To test the observation uncertainty, we also analyze the ERA5 reanalysis datasets (https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5) (1980−2014). Comparisons show that ENSO teleconnections in GPCP and ERA5 are very similar (see results section). We thus use the GPCP as the observation benchmark.

For CMIP6 model simulations (Eyring et al. 2016) (https://pcmdi.llnl.gov/CMIP6), we analyze the surface temperature (ts) and precipitation (pr) from the following experiments: historical run in the 1850−2014 period and the highest warming scenario
Table 1. Information for the 32 CMIP6 models that we analyze in this study. The first column is the model ID. The second column is the model name. The third column is the model ensemble ID that we use in this study. Note that we only choose one ensemble run from each model to create the multimodel mean. For these 32 models, we analyze variables of surface temperature (ts) and precipitation (pr) in the historical period and the SSP585 scenario. Model ECS and TCR are from Hausfather et al. (2022). We group models based on likely ECS range (2.5°–4°C).

<table>
<thead>
<tr>
<th>Model ID</th>
<th>Model name</th>
<th>Ensemble ID</th>
<th>ECS</th>
<th>TCR</th>
<th>ECS screen (likely) 2.5°–4.0°C</th>
<th>TCR screen (likely) 1.4°–2.2°C</th>
<th>Group</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>ACCESS-CM2</td>
<td>r1i1p1f1</td>
<td>4.66</td>
<td>1.96</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>2</td>
<td>ACCESS-ESM1-5</td>
<td>r1i1p1f1</td>
<td>3.88</td>
<td>1.97</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>3</td>
<td>BCC-CSM2-MR</td>
<td>r1i1p1f1</td>
<td>3.02</td>
<td>1.55</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>4</td>
<td>CAMS-CSM1-0</td>
<td>r1i1p1f1</td>
<td>2.29</td>
<td>1.73</td>
<td>N</td>
<td>Y</td>
<td>Cold</td>
</tr>
<tr>
<td>5</td>
<td>CESM2</td>
<td>r1i1p1f1</td>
<td>5.15</td>
<td>2</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>6</td>
<td>CESM2-WACCAM</td>
<td>r1i1p1f1</td>
<td>4.68</td>
<td>1.91</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>7</td>
<td>CNRM-CM6-1</td>
<td>r1i1p1f2</td>
<td>4.9</td>
<td>2.22</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>8</td>
<td>CNRM-CM6-1-HR</td>
<td>r1i1p1f2</td>
<td>4.34</td>
<td>2.46</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>9</td>
<td>CNRM-ESM2-1</td>
<td>r1i1p1f2</td>
<td>4.79</td>
<td>1.83</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>10</td>
<td>CanESM5</td>
<td>r1i1p1f1</td>
<td>5.64</td>
<td>2.71</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>11</td>
<td>CanESM5-CanOE</td>
<td>r1i1p2f1</td>
<td>5.64</td>
<td>2.71</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>12</td>
<td>EC-EARTH3</td>
<td>r1i1p1f1</td>
<td>4.26</td>
<td>2.3</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>13</td>
<td>EC-EARTH3-Veg</td>
<td>r1i1p1f1</td>
<td>4.33</td>
<td>2.66</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>14</td>
<td>FGOALS-f3-L</td>
<td>r1i1p1f1</td>
<td>3</td>
<td>1.94</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>15</td>
<td>GFEDL-CM4</td>
<td>r1i1p1f1</td>
<td>3.89</td>
<td>2</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>16</td>
<td>GFEDL-ESM4</td>
<td>r1i1p1f1</td>
<td>2.65</td>
<td>1.63</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>17</td>
<td>GiSS-E2-1-G</td>
<td>r1i1p1f1</td>
<td>2.64</td>
<td>1.8</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>18</td>
<td>HadGEM3-GC31-LL</td>
<td>r1i1p1f3</td>
<td>5.55</td>
<td>2.49</td>
<td>N</td>
<td>N</td>
<td>Hot</td>
</tr>
<tr>
<td>19</td>
<td>INM-CM4-8</td>
<td>r1i1p1f1</td>
<td>1.83</td>
<td>1.3</td>
<td>N</td>
<td>N</td>
<td>Cold</td>
</tr>
<tr>
<td>20</td>
<td>INM-CM5-0</td>
<td>r1i1p1f1</td>
<td>1.92</td>
<td>1.41</td>
<td>N</td>
<td>Y</td>
<td>Cold</td>
</tr>
<tr>
<td>21</td>
<td>IPSL-CM6a-LR</td>
<td>r1i1p1f1</td>
<td>4.7</td>
<td>2.35</td>
<td>N</td>
<td>N</td>
<td>Hot</td>
</tr>
<tr>
<td>22</td>
<td>K-ACE-1-0-G</td>
<td>r1i1p1f1</td>
<td>4.75</td>
<td>2.04</td>
<td>N</td>
<td>Y</td>
<td>Hot</td>
</tr>
<tr>
<td>23</td>
<td>MCM-UA-1-0</td>
<td>r1i1p1f1</td>
<td>3.76</td>
<td>1.9</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>24</td>
<td>MiroC-ES2L</td>
<td>r1i1p1f2</td>
<td>2.66</td>
<td>1.49</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>25</td>
<td>MiroC6</td>
<td>r1i1p1f1</td>
<td>2.6</td>
<td>1.55</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>26</td>
<td>MPI-ESM1-2-HR</td>
<td>r1i1p1f1</td>
<td>2.98</td>
<td>1.64</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>27</td>
<td>MPI-ESM1-2-LR</td>
<td>r1i1p1f1</td>
<td>3.02</td>
<td>1.82</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>28</td>
<td>MRI-ESM2-0</td>
<td>r1i1p1f1</td>
<td>3.13</td>
<td>1.67</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>29</td>
<td>NESM3</td>
<td>r1i1p1f1</td>
<td>4.72</td>
<td>2.72</td>
<td>N</td>
<td>N</td>
<td>Hot</td>
</tr>
<tr>
<td>30</td>
<td>NorESM2-LM</td>
<td>r1i1p1f1</td>
<td>2.56</td>
<td>1.49</td>
<td>Y</td>
<td>Y</td>
<td>Median</td>
</tr>
<tr>
<td>31</td>
<td>NorESM2-MM</td>
<td>r1i1p1f1</td>
<td>2.49</td>
<td>1.22</td>
<td>N</td>
<td>N</td>
<td>Cold</td>
</tr>
<tr>
<td>32</td>
<td>UKESM1-0-LL</td>
<td>r1i1p1f2</td>
<td>5.36</td>
<td>2.77</td>
<td>N</td>
<td>N</td>
<td>Hot</td>
</tr>
</tbody>
</table>

(SSP585, the emission scenario considering SSP5 and radiative forcing of 8.5 W m⁻² at the end of the twenty-first century) in the 2015–99 period to assess warming-induced changes. To make a fair comparison among observations and models with varying spatial resolution, a regridding to 1.5° x 1.5° is carried out. We use 32 models that have both ts and pr for the historical period and SSP585 during the time of analysis. These 32 CMIP6 models give a reasonable sample size to make a reliable multimodel mean. We take into consideration the “hot” model issues in the CMIP6 archive (Hausfather et al. 2022), and we group models into “hot,” “median,” and “cold” models based on the equilibrium climate sensitivity (ECS) likely range (2.5°–4°C) to address the uncertainty in the future projections. Detailed model information is in Table 1.

b. Definition of subdomains

Here, we define the TP domain covering 10°S–10°N, 150°E–80°W. We define the MC covering 18°S–26°N, 80°–160°E. Considering the spatial inhomogeneity within the MC, we further define the CMC covering 10°S–10°N, 90°–150°E. It mainly covers the area with a coherent ENSO–rainfall teleconnection. We further define the EMC covering 5°S–0°, 150°–160°E. We use this small marginal EMC region to highlight the eastward shift of the ENSO teleconnection. In this study, we first calculate spatial maps of the ENSO–rainfall teleconnection. Given the overall spatial coherence within the CMC and TP, we further calculate the ENSO teleconnection using the domain-averaged rainfall.

c. Time snippets and seasonal cycles

We address the future changes based on time snippets: 1980–2014 as the base period (35 years, using the historical run), and 2065–99 as the warming period (35 years, using the SSP585 scenario). We take a 3-month running average of the monthly data to create a seasonal mean, from DJF, January–March (NDJ), to November–January (NDJ). We label the center month of the season when plotting all 12-month results, for example, “J” represents “DJF” and “D” represents...
“NDJ.” We take a 35-yr time snippet of the seasonal mean data to carry out the analysis. In each 35-yr time snippet, we carry out the preprocessing to remove the 12-month climatology and the cubic-polynomial trend to obtain the monthly anomaly of temperature and precipitation.

To analyze changes across time, we merge the historical and SSP585 runs to create a long record (1850–2099, 250 years). We add the 2025–59 period (35 years, using the SSP585 scenario) as the near-future time snippet to show how changes scale with warming. We use a 35-yr running window with a 1-yr shift to create 216 time snippets. For each time snippet, we conduct the same analysis, including the de-climatolgy and cubic-polynomial detrending.

Note that there are several ways to do de-climatology and detrending. Bonfils et al. (2015) used global-averaged temperature to remove the trend. Power and Delage (2018) used spectral filtering to remove the long-term trend. Here, we use cubic polynomial to remove the trend. These preprocessing methods lead to similar conclusions. We also test detrending for the merged 1850–2099 period in the preprocessing instead of doing it for each 35 years, and we find both ways of detrending give almost the same results and our conclusions still hold.

d. ENSO states and composites

We calculate the Niño-3.4 index (5°S–5°N, 170°–120°W averaged) to provide an estimate of the warming signal across the TP, which is highly correlated with the global mean surface temperature [correlation coefficient (corrcoeff) > 0.95, with p value < 1 × 10^{-4}]. Here, we use one standard deviation of the Niño-3.4 index to indicate the ENSO amplitude manifested in ts (std ts_N34). ENSO normally peaks during the winter (DJF), so we divide the ENSO states using the threshold (Thr) of 0.5 times one standard deviation of the Niño-3.4 ts anomaly during the DJF season: El Niño year: ts anomaly > Thr; La Niña year: ts anomaly < −Thr; and a neutral year: −Thr ≤ ts anomaly ≤ Thr. We conduct a composite analysis based on these ENSO states.

We take composites to analyze El Niño- and La Niña–induced rainfall variability. For variables like the domain-averaged rainfall in the CMC, we compute the composites from April–May–June (AMJ) of the ENSO year (year 0) to the next MAM (year +1) to show the temporal evolution covering the whole ENSO life cycle. For variables like the equatorial rainfall (5°S–5°N averaged), we conduct the composite covering the JJA season of the ENSO year 0 and DJF season of the ENSO year 0/1. Here, we used the climatology as the benchmark to compute the composites; Bonfils et al. (2015), Power and Delage (2018) used the neutral state as the benchmark, and their comparison showed that both benchmarks give consistent results.

e. ENSO–rainfall correlation and covariance

Previous studies on ENSO teleconnection use either correlation or covariance metrics (Bonfils et al. 2015; Perry et al. 2017; Wang et al. 2020). Here, we first investigate the teleconnection using a correlation method (Feng and Hu 2004; Haszpra et al. 2020; Wang et al. 2020; Cai et al. 2009). We calculate the in-phase (lag = 0) Pearson correlation coefficient using the pr anomaly (domain averaged or grid cell based) and the Niño-3.4 ts anomaly. The in-phase correlation is different from the lagged correlation (between Niño-3.4 ts anomaly at DJF peak season and rainfall at different seasons) (Wang et al. 2020). Both approaches give consistent conclusions. We also investigate the teleconnection using a covariance method (Bonfils et al. 2015; Perry et al. 2017). Here, we calculate the in-phase (lag = 0) covariance of the pr anomaly with a normalized Niño-3.4 ts anomaly by dividing its standard deviation. We then analyze the ENSO-induced rainfall variability (mm day^{-1}) in the historical and future warming scenarios.

f. Spatial changes in the dipole-like ENSO teleconnection and the equatorial rainfall

We calculate the in-phase (lag = 0) covariance of the equatorial Indo-Pacific pr anomaly with the normalized Niño-3.4 ts anomaly. Across the longitude, it shows a dipole structure (negative rainfall variability in the MC and positive rainfall variability in the TP). Based on this structure, we define the zero crossing longitude. We use this metric to measure the eastward shift of the ENSO teleconnection. Note that the model’s spatial resolution is 1.5° in longitude. We thus consider the shift larger than 3° as a notable change. We also analyze the equatorial (5°S–5°N averaged) Indo-Pacific rainfall. We carry out a three-point running smoothing for the zonal profile. To highlight an eastward shift in the mean rainfall centered around the Maritime Continent, we define the longitude of the rainfall center as averaged between 90°E and 180° around the warm pool.

g. Measures of CMIP6 model behaviors

We compute multimodel means to assess the average model behavior. We also check the model agreement on the projection. We consider the agreement on the sign of change by 3/4 of the models (24 out of 32 in this study) to be a reasonable criterion for robustness. To illustrate the model diversity as to the future changes, we show each model’s value for the historical and SSP585 period in a rank based on that model’s value for the historical period. When plotting, we mark the value of the observation and multimodel mean values for both periods. We count the number of models that agree on the sign of change and mark them with distinct labels (filled circle). We use this type of model rank diagram to visualize models with larger or smaller biases compared with the observation.

Individual models could project diverse future changes for varying quantities. For two related quantities, we show scatterplots to characterize the correlated model behaviors. For example, models projecting a larger rainfall variability in CMC are also models projecting a larger rainfall variability in the TP. In such scatterplots, individual models are labeled with model IDs. The corrcoeff between two quantities is shown with the p value indicating the robustness.

3. Results

a. Observed ENSO–rainfall teleconnection across the Maritime Continent

When El Niño takes place, the TP is often wetter than normal and the MC is drier than normal (Fig. 1a). Within the
MC, CMC mainly covers the area with a coherent ENSO–rainfall teleconnection (Fig. 1b). Across 12 calendar months, the Niño-3.4 ts variability is positively correlated with the TP rainfall [Fig. 1e; corr (ts_N34, pr_TP) = 0.88 for JJA, 0.95 for DJF] while negatively correlated with the CMC rainfall [Fig. 1c; corr (ts_N34, pr_CMC) = −0.77 for JJA, −0.83 for DJF]. Here, EMC is a marginal area between the MC and TP, and EMC has a positive ENSO–rainfall teleconnection [Fig. 1d; corr (ts_N34, pr_EMCC) = 0.55 for JJA, 0.31 for DJF], unlike the CMC with a negative teleconnection. We later use EMC to highlight the eastward shift of the ENSO teleconnection under warming.

The CMC domain has 15% land coverage, compared with the oceanic domains of the TP and EMC. We apply land–sea masks to separate the land and sea rainfall over the CMC region. We find that summer (JJA) rainfall teleconnection across the CMC is strong over both land and sea [corr (ts_N34, pr_CMC_land) = −0.78 and corr (ts_N34, pr_CMC_sea) = −0.72] (Fig. 1c). We later show that ENSO–rainfall teleconnection over CMC enhances under warming for all three cases, that is, all-surface, land-only, and sea-only precipitation. In this study, we will mainly show results for the all-surface rainfall.

ENSO-induced rainfall variability over the TP is spatially coherent and stable across seasons (Fig. 2). However, rainfall teleconnection across the MC varies in space and across seasons. The summer rainfall variability is near the southern and central MC, while the winter variability is northeastward compared with JJA, which agrees with the seasonal circulation migration seen in Juneng and Tangang (2005) and Supari et al. (2018).

b. ENSO–rainfall teleconnection simulated by CMIP6 models

We analyze the tropical rainfall in the historical period (1980–2014) using 32 available CMIP6 models. CMIP6 models overall reproduce the dipole-shaped ENSO teleconnections across the Indo-Pacific Ocean for all four seasons (Fig. 3). Across individual CMIP6 models, most models give realistic ENSO teleconnection patterns (Fig. A1). Only a few outlier models produce very weak rainfall variability (19: INM-CM4-8, 20: INM-CM5-0) or have a misplaced variability pattern (23: MCM-UA-1-0).

As to the teleconnection intensity (Fig. 4, Table 2), models match the observed TP rainfall covariance during summer but slightly underestimate that for winter. For the CMC, models overestimate the ENSO-induced negative rainfall variability across seasons. Over the EMC, models overestimate the positive variability. To check the observation, we show that the ENSO teleconnection in the ERA5 reanalysis is very similar to the observation (GPCP). We thus show the GPCP results as the observation benchmark.
We then investigate the model diversity on the teleconnection across 32 models. With the physical bond of the Walker circulation movement, ENSO-induced rainfall variability over the CMC is closely linked to that over the TP. Here, we show that models with larger positive ENSO-induced rainfall variability in the TP are also models with larger negative variabilities in the CMC (Figs. 5a, b). This relation holds for all four seasons, although it develops from summer and becomes stronger in the winter peak season (for the historical period, corrcoeff_DJF $= -0.69$ with $p$ value $< 10^{-5}$). We also show that ENSO-induced rainfall variability across individual models is correlated with the ENSO amplitude (Figs. 5c, d) (for the historical period, corrcoeff_DJF $= -0.69$ with $p$ value $< 10^{-5}$). That is, models with a weak and unrealistic ENSO teleconnection are those models with weak SST variability. It indicates that the model performance on ENSO–rainfall teleconnection is tightly linked to the model performance on ENSO itself (Cai et al. 2009).

c. Intensification of the ENSO–rainfall teleconnection under warming

To address the future changes in the ENSO teleconnection, we analyze the tropical rainfall near the end of the twenty-first century (2065–99) under a high-emission scenario (SSP585). Under warming, the ENSO teleconnection shows a zonal dipole structure similar to the historical period: the TP is overall wetter than normal during El Niño and drier during La Niña; CMC is overall drier than normal during El Niño and wetter during La Niña (Fig. 6). Under warming, CMIP6 models robustly project that ENSO-induced positive rainfall variability over the TP will increase (Fig. 7a) [from 0.58 to 0.80 mm day$^{-1}$ (38% increase)]
for JJA with 28/32 models agreed; from 0.89 to 1.19 mm day$^{-1}$ (34% increase) for DJF with 28/32 models agreed.

We also show that changes in ENSO teleconnection over the TP scale with changes in ENSO amplitude (Fig. 7b). Models projecting a larger increase in ENSO amplitude also project a greater increase in the rainfall variability (corrcoeff = 0.56 with $p$ value $< 10^{-4}$ for JJA). It indicates that the ENSO physics still bonds the ENSO ts variability and the ENSO–rainfall teleconnection under a changing climate. We also note that future change of ENSO amplitude tends to be larger during the summer–fall season than during the boreal winter (e.g., std ts_N34 from 0.84° to 0.98°C with 24/32 models agreed for JJA, from 1.13° to 1.19°C with only 16/32 models agreed for DJF). In section 5c, we will have more discussion as to ENSO teleconnection changes compared with ENSO changes across CMIP6 models.

As to changes over the CMC domain, CMIP6 models robustly project that ENSO-induced negative rainfall variability will enhance, on average, from −0.26 to −0.59 mm day$^{-1}$ (127% enhancement) for JJA with 29/32 models agreed (Fig. 8a), and from −0.46 to −0.62 mm day$^{-1}$ (35% enhancement) for DJF with 27/32 models agreed. Note that one outlier model, MCM-UA-1-0 (model 23), projects weakened summer (JJA) precipitation variability in both the TP and CMC (Fig. 8b); this model was shown to have unrealistic ENSO responses in the tropics and was thus not reliable (Fig. A1).

We also show that models projecting a stronger negative enhancement of CMC rainfall variability are models with a more positive enhancement of TP rainfall variability (corrcoeff = −0.55 with $p$ value $= 10^{-3}$ for JJA) (Fig. 8b). It indicates that the whole dipole-like precipitation variability across the Indo-Pacific Ocean becomes stronger under warming, across
four seasons (Fig. 9). Also note that changes over the CMC are larger during the spring to autumn than during the winter (Fig. 9b).

d. Intensification of both land and sea rainfall teleconnection over the CMC

Over the CMC, we further investigate the land–sea differences as to the future changes. In the observation, the ENSO-induced variability in the summer (JJA) sea rainfall \[\text{cov} \,(\text{ts\_N34, pr\_CMB\_sea}) = -0.45 \text{ mm day}^{-1}\] is similar to the variability in all-surface rainfall \[\text{cov} \,(\text{ts\_N34, pr\_CMB\_all}) = -0.51 \text{ mm day}^{-1}\]. Land rainfall variability is larger \[\text{cov} \,(\text{ts\_N34, pr\_CMB\_land}) = -0.84 \text{ mm day}^{-1}\] (Fig. 10). CMIP6-modeled land and sea rainfall variabilities in the historical period are underestimated compared with the observation. Under warming, models project that ENSO–rainfall teleconnection enhances for both land and sea rainfall, and changes are more notable in summer than in winter (for land: from \(-0.49\) to \(-0.86 \text{ mm day}^{-1}\) at JJA, from \(-0.08\) to \(-0.09 \text{ mm day}^{-1}\) at DJF; for sea, from \(-0.21\) to \(-0.54 \text{ mm day}^{-1}\) at JJA, from \(-0.53\) to \(-0.72 \text{ mm day}^{-1}\) at DJF) (Fig. 10). Given the consistency between the land and sea rainfall variability, we will mainly show results of the all-surface precipitation.

e. Intensification of rainfall teleconnections for both El Niño and La Niña

To address the discrepancy between El Niño– and La Niña–induced ENSO teleconnection, we next analyze rainfall variability composites across the life cycle of El Niño and La Niña years (Fig. 11, Table 3). ENSO SST variability normally peaks in boreal winter. Over the TP, El Niño–induced positive rainfall variability develops through the boreal spring–summer to peak at winter \[\text{pr\_EN\_obs = 0.60 \text{ mm day}^{-1} \text{ for JJA, 1.38 mm day}^{-1} \text{ for DJF}}\] (Fig. 11a). It is stronger than the La Niña–induced negative rainfall variability \[\text{pr\_LN\_obs = -0.38 \text{ mm day}^{-1} \text{ for JJA (63% of pr\_EN\_obs), -1.06 mm day}^{-1}}\] for DJF (77% of pr\_EN\_obs). Over the CMC, El Niño–induced negative rainfall variability develops through the boreal spring–summer to peak at fall \[\text{pr\_EN\_obs = -0.70 \text{ mm day}^{-1} \text{ for JJA, -1.41 mm day}^{-1} \text{ for October–December (OND)}}\] (Fig. 11b). La Niña–induced positive rainfall variability is weaker than El Niño–induced variability \[\text{pr\_LN\_obs = 0.49 \text{ mm day}^{-1} \text{ for JJA (70% of pr\_EN\_obs), 0.94 mm day}^{-1}}\] for OND (67% of pr\_EN\_obs). ENSO-induced rainfall variability in the CMC tends to peak earlier during fall than that in the TP due to the interplay with the seasonal migration of the monsoon system (Chakraborty 2018; Feng and Hu 2004). Along with ENSO evolution, rainfall variability in varying subdomains of the MC have different peak seasons and associated impacts (Juneng and Tangang 2005).

Under warming, the JJA rainfall variabilities over the TP enhance (from \(0.48\) to \(0.58 \text{ mm day}^{-1}\) for El Niño year, and from \(-0.38\) to \(-0.50 \text{ mm day}^{-1}\) for La Niña year, Fig. 11a). DJF peak season rainfall variability increases from 1.11 to 1.47 \text{ mm day}^{-1} in El Niño year 0/1, and from \(-0.95\) to \(-1.29 \text{ mm day}^{-1}\) in La Niña year 0/1. Future changes show a small discrepancy between El Niño–induced (0.36 \text{ mm day}^{-1}) at DJF peak season) and La Niña–induced (0.34 \text{ mm day}^{-1}) at

Table 2. ENSO–rainfall teleconnection in observation, reanalysis, and models. The first column is the ENSO teleconnection measure. The second column is the observation (ts from HadISST, pr from GPCP). The third column is the reanalysis (ERA5). The fourth column is the CMIP6 historical model for the historical period. The fifth column is the 32-CMIP6-model mean for the 2065–99 period in the SSP585 scenario. Units are mm day\(^{-1}\).

<table>
<thead>
<tr>
<th>ENSO–rainfall teleconnection covariance measure</th>
<th>Observation</th>
<th>Reanalysis (ERA5)</th>
<th>CMIP6 historical</th>
<th>CMIP6 SSP585</th>
</tr>
</thead>
<tbody>
<tr>
<td>\text{cov} ,(\text{ts_N34, pr_TP}) in JJA</td>
<td>0.57</td>
<td>0.54</td>
<td>0.58</td>
<td>0.80</td>
</tr>
<tr>
<td>\text{cov} ,(\text{ts_N34, pr_TP}) in DJF</td>
<td>1.07</td>
<td>1.12</td>
<td>0.89</td>
<td>1.19</td>
</tr>
<tr>
<td>\text{cov} ,(\text{ts_N34, pr_CMB}) in JJA</td>
<td>-0.51</td>
<td>-0.64</td>
<td>-0.26</td>
<td>-0.59</td>
</tr>
<tr>
<td>\text{cov} ,(\text{ts_N34, pr_CMB}) in DJF</td>
<td>-0.81</td>
<td>-0.81</td>
<td>-0.46</td>
<td>-0.62</td>
</tr>
<tr>
<td>\text{cov} ,(\text{ts_N34, pr_EMC}) in JJA</td>
<td>0.81</td>
<td>1.29</td>
<td>1.51</td>
<td>0.32</td>
</tr>
<tr>
<td>\text{cov} ,(\text{ts_N34, pr_EMC}) in DJF</td>
<td>1.08</td>
<td>0.87</td>
<td>2.46</td>
<td>2.68</td>
</tr>
</tbody>
</table>
DJF) rainfall variabilities. Over the CMC, El Niño–induced rainfall variabilities enhance (from $0.33$ to $0.67$ mm day$^{-1}$ at JJA, and from $-0.61$ to $-0.72$ mm day$^{-1}$ at OND of ENSO year 0) (Fig. 11b). Here, summer rainfall variability tends to increase more than fall variability ($0.34$ mm day$^{-1}$ for JJA, $0.1$ mm day$^{-1}$ for OND). La Niña–induced rainfall variabilities also enhance (from $0.32$ to $0.55$ mm day$^{-1}$ at JJA, and from $0.51$ to $0.59$ mm day$^{-1}$ at OND of ENSO year 0), and these changes ($0.23$ mm day$^{-1}$ for JJA, $0.08$ mm day$^{-1}$ for OND) are slightly smaller than the changes in the El Niño–induced variability.

f. Eastward shift of the ENSO–rainfall teleconnection under warming

In addition to the enhancement, ENSO teleconnection adjusts its spatial structure under warming. The covariance between the equatorial Indo-Pacific rainfall and the anomalous Niño-3.4 index displays a longitudinal dipole structure (positive in the TP but negative in the MC) (Fig. 12, Table 4) (dipole zero crossing longitude $\text{lon}_0\text{-cross}_{\text{obs}} = 142^\circ$E for JJA, $154^\circ$E for DJF in the observation). The multimodel mean shows that the zero crossing longitude of the dipole is near the eastern margin of the MC [$\text{lon}_0\text{-cross}_{\text{model}} = 137^\circ$E for JJA (8 bias), $137^\circ$E for DJF (17 bias)]. Such model biases indicate that the modeled TP teleconnection extends too far west into the MC. See more discussion in section 5a.

Under warming, models project that the zero crossing longitude will shift eastward under warming (from $137^\circ$ to $146^\circ$E for JJA, shifting 9$^\circ$) (Fig. 12b). Most models (29/32) agree on the sign of change, except for a few models. Model CNRM-CM6-1-HR (model 8) and model GISS-E2-1-G (model 17) do not project much change. Model INM-CM4-8 (model 19)
projects a rather westward shift; this model shows one outlier that does not produce reasonable ENSO teleconnection (Fig. A2). We also note that the projected eastward shift in the zero crossing longitude of the rainfall covariance is more notable during summer than winter (Fig. 13).

Here, we highlight that the eastward shift can result in the reduction of the ENSO teleconnection over a small EMC domain, which is near the dipole zero crossing but on the TP side of positive rainfall variability. This EMC region is usually wetter than normal during El Niño, same as the TP (Fig. 1). However, CMIP6 models robustly project (27/32 models agree) the summer (JJA) positive rainfall variability will reduce [cov (ts_N34, pr_EMC) = 1.51 mm day^{-1} in the historical period, to 0.32 mm day^{-1} in the warming scenario] (Fig. 14a). Also note that the reduction is mainly in the summer compared with the winter (for DJF, 2.46-2.68 mm day^{-1}, little change with no robust agreement across models) (Fig. 14b), which matches with the eastward shift of the ENSO teleconnection.

### g. Eastward shift of rainfall teleconnections for both El Niño and La Niña

The composite analysis also agrees on the projected eastward shifts for both El Niño– and La Niña–induced rainfall variability (Fig. 15). El Niño– and La Niña–induced zonal dipole structures are very similar. As to the zero crossing longitude, El Niño’s location is slightly more east than La Niña in the observation (for JJA, 0-cross_EN = 143°E, 0-cross_LN = 140°E; for DJF, 0-cross_EN = 157°E, 0-cross_LN = 149°E) and in the models (for JJA, 0-cross_EN = 139°E, 0-cross_LN = 138°E; for DJF, 0-cross_EN = 132°E, 0-cross_LN = 130°E).

Under warming, zero crossing longitude for the summer (JJA) season of the El Niño year shifts from 139° to 150°E.
For the La Niña year, the zero crossing longitude shifts from 138° to 148°E (shifting 10°). For the winter season (DJF) of the El Niño year 0/1, zero crossing longitude shifts from 132° to 134°E (shifting 2°). For the DJF of La Niña year 0/1, the zero crossing longitude shifts from 130° to 131°E (shifting 1°). These winter variations (similar range to the model’s spatial resolution of 1.5°) are considered insignificant changes compared with the summer season (~10°).

h. ENSO–rainfall teleconnection shifts seasonally along with the mean rainfall center

We further investigate the relationship between spatial changes in the rainfall variability and the rainfall itself. Walker circulation can be represented in the equatorial zonal streamfunction and the vertical velocity at the 500-hPa level (Bayr et al. 2014). Vertical velocity at 500 hPa also generally matches the precipitation pattern (Zhang et al. 2021). Here, we use the

Unauthenticated | Downloaded 10/30/23 02:38 AM UTC
mean rainfall spatial changes as simple diagnostics to indicate
changes in the convection and the Walker circulation. We
mainly focus on the changes in the zonal structure, but we are
aware that regionally developed circulation over the tropics
could break the trans-Pacific Walker circulation into smaller
regional scales and may lead to greater increases in precipita-
tion in the western equatorial Pacific compared with the east-
er equatorial Pacific, as shown in Sohn et al. (2019).

By defining the longitude of the Indo-Pacific precipitation
center around the warm pool, we show that CMIP6 models
robustly project that the rainfall center will move eastward
under warming (Fig. 16b) (on average, from 132° to 137°E for
JJA, 31/32 models agreed) and the shift is larger during sum-
mer than winter (from 131° to 132°E for DJF) (Fig. 16d). It
matches well with the seasonal reduction of the rainfall tele-
connection over EMC (Fig. 14b) and the shift in the ENSO
teleconnection (Fig. 13).

Physically, zonal shifts of the rainfall anomaly are naturally
linked to that of the mean rainfall climatology. This linkage is
manifested in the model diversity across historical simula-
tions. Models with the teleconnection zero crossing toward
the east are often models with the mean rainfall center toward
the east (corcoeff = 0.78 with p value = 10^{-3} for JJA)
(Fig. 17). Meanwhile, models projecting a further eastward
shift in the ENSO teleconnection are generally those projecting
a further eastward shift in the rainfall center (corcoeff = 0.41
with p value = 0.02 for JJA) (Fig. 17d).

i. Changes in the teleconnection scale up with the
warming trend

Thus far, we have analyzed the historical (1980–2014) and
end of the twenty-first century (2065–99) time snippets of
CMIP6 models. We show that under the warming climate, the
ENSO-induced dipole-like precipitation variability could en-
hance and shift eastward. Here, we create a long record
(1850–2099) by merging the CMIP6 model historical and
SSP585 runs. We show that our future change estimates based
on time snippets are robust and agree with the changes over
time (Fig. 18). Niño-3.4 mean temperature is 26.66°C during
the historical base period, 27.96°C (1.3°C warming) in the
near future (2025–59), and 29.86°C (3.2°C warming) in the far
future (2065–99). Along with increasing warming levels, the
multimodel mean changes are shown to scale up with the
warming even with natural variations.

![Fig. 9. Intensification of ENSO–rainfall teleconnection over the TP and CMC. (a) 12-month ENSO–precipitation
covariance across the TP. Observations (GPCP; black curve), the multimodel mean of 32 CMIP6 models for the his-
torical period (blue curve), and the multimodel mean for the SSP585 scenario (red curve) are shown. The shades indi-
cate the 95% model range. (b) As in (a), but for rainfall teleconnection over the CMC.](Unauthenticated | Downloaded 10/30/23 02:38 AM UTC)
Changes in the teleconnection in “hot,” “median,” and “cold” models

Recent studies have shown that CMIP6 models tend to have higher ECS, and projected warming levels could be overestimated (Hausfather et al. 2022). To address the model uncertainty and provide more realistic projections, we separate 32 CMIP6 models into three groups: 13 “median” models ($2.5^\circ \leq \text{ECS} \leq 4^\circ$, match the IPCC AR6 assessed warming), 15 “hot” models (ECS > $4^\circ$), and 4 “cold” models (ECS < $2.5^\circ$). We then show results of the three groups of models (Fig. 19, Table 5). Instead of showing absolute values, here we show changes from the 1980–2014 base period.

As to the tropical ts ($\text{ts}_\text{N34}$) increase from the base period to the 2065–99 period (Fig. 19a), all-model mean projects $3.2^\circ$, hot-model mean projects $3.83^\circ$, median-model mean projects $2.77^\circ$, and cold-model mean projects $2.27^\circ$. As to the std of ts$_\text{N34}$ increase (Fig. 19b), all-model mean projects $0.14^\circ$, hot-model projects a larger increase than the cold model. As to the increase of positive rainfall covariance over the TP (Fig. 19c), all-model mean projects $0.22$ mm day$^{-1}$. Hot models project a larger increase than the cold models.

As to the decrease of negative (i.e., enhanced) rainfall covariance over the TP (Fig. 19d), all-model mean projects $0.33$ mm day$^{-1}$. Hot models project a larger enhancement than the cold models. As to the eastward shift of zero crossing longitude (Fig. 19e), all-model mean projects $9.75^\circ$. Hot models project a larger shift than the cold models. As to the eastward shift of warm pool rainfall center (Fig. 19f), all-model mean projects $5.4^\circ$. Hot models project a larger shift than the cold models. Note that cold-model change appears to be slightly larger than the median model, but the average based on only four cold models is not robust and the change is within the noise range compared with the variability.

In summary, we show that all three groups of models agree on the sign of change for aspects we investigate; this consistency suggests that our conclusions based on all 32 models are robust and meaningful. We also show that models with higher ECS (i.e., hot models) project a faster warming and a stronger ENSO. Hot models project a stronger enhancement of ENSO teleconnection over the TP and CMC. They also project a larger eastward shift of the ENSO teleconnection zero crossing and a larger eastward shift of the warm pool rainfall center. These agree with our conclusions that changes scale with

Table 3. El Niño– and La Niña–induced rainfall variability in observation and models. The second column is the observation (ts from HadISST, pr from GPCP). The third column is the 32-CMIP6-model mean for the historical period. The fourth column is the 32-model mean for the 2065–99 period in the SSP585 scenario. Units are mm day$^{-1}$.

<table>
<thead>
<tr>
<th>ENSO-induced rainfall variability</th>
<th>Observation</th>
<th>CMIP6 historical</th>
<th>CMIP6 SSP585</th>
</tr>
</thead>
<tbody>
<tr>
<td>El Niño–induced variability over TP (JJA)</td>
<td>0.60</td>
<td>0.48</td>
<td>0.58</td>
</tr>
<tr>
<td>La Niña–induced variability over TP (JJA)</td>
<td>$-0.38$</td>
<td>$-0.38$</td>
<td>$-0.50$</td>
</tr>
<tr>
<td>El Niño–induced variability over TP (DJF)</td>
<td>1.38</td>
<td>1.11</td>
<td>1.47</td>
</tr>
<tr>
<td>La Niña–induced variability over TP (DJF)</td>
<td>$-1.06$</td>
<td>$-0.95$</td>
<td>$-1.29$</td>
</tr>
<tr>
<td>El Niño–induced variability over CMC (JJA)</td>
<td>$-0.70$</td>
<td>$-0.33$</td>
<td>$-0.67$</td>
</tr>
<tr>
<td>La Niña–induced variability over CMC (JJA)</td>
<td>0.49</td>
<td>0.32</td>
<td>0.55</td>
</tr>
<tr>
<td>El Niño–induced variability over CMC (OND)</td>
<td>$-1.41$</td>
<td>$-0.62$</td>
<td>$-0.72$</td>
</tr>
<tr>
<td>La Niña–induced variability over CMC (OND)</td>
<td>0.94</td>
<td>0.51</td>
<td>0.59</td>
</tr>
</tbody>
</table>
warming. It may be difficult to determine if hot models are indeed too hot considering that ongoing warming pathways are uncertain. Here, we at least communicate the projections from three groups to account for the model uncertainty.

4. Impacts of changes in ENSO teleconnection over the MC

The intensity of ENSO teleconnection could mean two things. One is the intensity of the ENSO-induced precipitation variability. The other one is the intensity of ENSO–rainfall correlation. Here, we analyze both aspects and show that both the ENSO–rainfall covariance \[\text{cov (ts\_N34, pr\_CMC)}\] from 20.26 to 20.59 mm day\(^{-1}\) at JJA, from 0.46 to 0.62 mm day\(^{-1}\) at DJF (Fig. 9b) and correlation \[\text{corr (ts\_N34, pr\_CMC)}\] from 0.35 to 0.61 at JJA, from 0.55 to 0.62 at DJF (Fig. 20a) agree that future ENSO teleconnection will enhance over the CMC. On the contrary, positive ENSO teleconnection over the EMC shows a decrease due to eastward shift of the dipole structure. The decrease can be seen in both covariance (Fig. 14b) \[\text{cov (ts\_N34, pr\_EMC)}\] from 1.51 to 0.32 mm day\(^{-1}\) at JJA \[\text{and correlation (Fig. 20b)}\] \[\text{corr (ts\_N34, pr\_EMC)}\] from 0.51 to 0.12 at JJA.

Projected enhancement and eastward shift of the ENSO–rainfall teleconnection under warming could have potential impacts across the MC region. For example, the summer transboundary haze across the MC is one major air pollution that has threatened human health in recent years, often enhanced during extended drought periods (Supari et al. 2018). Based on our CMIP6 model projections, El Niño–induced drought could strengthen during El Niño years, which may escalate the haze situation. On the other hand, the MC is wetter than normal during La Niña years, and it could become even wetter under warming and lead to more flooding.

Moreover, changes in ENSO–rainfall teleconnection may have impacts on the seasonal predictability of the MC rainfall under warming. If one region’s rainfall is strongly correlated with ENSO, it often indicates that such a region receives more direct ENSO modulations and is thus more predictable. The temporal stationarity of the teleconnection is important (Coats et al. 2013). If the ENSO–rainfall correlation is stable

<table>
<thead>
<tr>
<th>ENSO–rainfall teleconnection zero crossing longitude</th>
<th>Observation</th>
<th>CMIP6 historical</th>
<th>CMIP6 SSP585</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-cross for cov (ts_N34, pr_eq) in JJA</td>
<td>142°E</td>
<td>137°E</td>
<td>146°E</td>
</tr>
<tr>
<td>0-cross for cov (ts_N34, pr_eq) in DJF</td>
<td>154°E</td>
<td>137°E</td>
<td>135°E</td>
</tr>
<tr>
<td>0-cross for El Niño–induced variability in JJA</td>
<td>142°E</td>
<td>139°E</td>
<td>150°E</td>
</tr>
<tr>
<td>0-cross for La Niña–induced variability in JJA</td>
<td>140°E</td>
<td>138°E</td>
<td>148°E</td>
</tr>
<tr>
<td>0-cross for El Niño–induced variability in DJF</td>
<td>157°E</td>
<td>132°E</td>
<td>134°E</td>
</tr>
<tr>
<td>0-cross for La Niña–induced variability in DJF</td>
<td>149°E</td>
<td>130°E</td>
<td>131°E</td>
</tr>
</tbody>
</table>

![Fig. 12. Summer (JJA) zonal dipole-like ENSO–rainfall teleconnection shifts eastward under warming. (a) Equatorial precipitation (5°S–5°N averaged) covariance with Niño-3.4 sea surface temperature. Observations (GPCP; black curve), the multimodel mean of 32 CMIP6 models for the historical period (blue curve), and the multimodel mean for the SSP585 scenario (red curve) are shown. The shading indicates the 95% model range. (b) Zero crossing longitude of the precipitation covariance across 32 CMIP6 models. The observation (black line), the model mean for the historical period (blue line), and the model mean for the future SSP585 scenario (red line) are shown. Model results are ordered by their values in the historical period. Model agreement on the future change is shown at the top.](image-url)
and even stronger under global warming, it will surely be helpful for higher predictability. Here, the enhanced negative ENSO–rainfall correlation over the CMC (Fig. 20a) suggests better predictability for the local rainfall.

We also note the spatial inhomogeneity as to future changes. Within the MC, CMC is drier than normal during El Niño. Instead, the EMC is wetter than normal during El Niño due to the fact that EMC is a part of the western Pacific on the positive variability side of the zonal dipole (Fig. 1). Under warming, the CMC is projected to enhance the negative rainfall variability, while the EMC is projected to reduce the positive rainfall variability along with the eastward shift of the ENSO teleconnection (Fig. 20b). These changes could trigger spatially inhomogeneous impacts on agriculture and other climate-relevant sectors in society (Anderson et al. 2017).

5. Discussion
a. CMIP6 model performance and biases on teleconnection

CMIP6 model biases over the MC can be traced back to the fact that the modeled TP teleconnection extends too far west into the MC (Fig. 3). We show that the zero crossing longitude of the zonal rainfall variability in the model is west of the observation, and the bias is more notable during winter than summer. The bias is more notable during winter than summer [lon_0-cross obs = 142°E for JJA, 154°E for DJF; lon_0-cross_model = 137°E for JJA (5° bias), 137°E for DJF (17° bias)]. Physically, the biased rainfall zonal structure can be attributed to factors such as the westward position of the rising branch of the Walker circulation as in CMIP3/5 (e.g., Cai et al. 2009), a westward far-reaching positive ts anomaly during El Niño.
linked to the cold tongue bias in modeled ts (Bayr et al. 2020; Li and Xie 2012), and excessive ITCZ biases (Zhou et al. 2020). Although such systematic biases still exist in CMIP6 models, there has been a large reduction of biases in the tropical cold tongue and ENSO teleconnections in the Indo-Australian monsoon region (Langenbrunner and Neelin 2013; Fasullo et al. 2020).

As to future changes, we show that individual projections are not largely sensitive to the historical bias of that model. For each quantity analyzed, we rank the results of each model based on their values for the historical period (e.g., Figs. 7a, 8a, 12b, 14a, and 16b). We do see a couple of model outliers which simulate unrealistic features and project differently (see the appendix), but the majority of models do provide meaningful projections. We show that models with larger or smaller biases do not give distinctly different future changes compared with the rest of the models, which indicates that future projections (at least on the sign of change) are not largely sensitive to the models’ historical biases.

Power et al. (2013) noticed that one main problem for the CMIP3/5 modeled teleconnection is that the positive loading in Pacific teleconnection tends to extend too far to the west. Therefore, Power et al. (2013) did tests to correct the model bias for future projections by shifting each model’s teleconnection patterns eastward to give the best match to the observed pattern. This simple correction leads to similar projected changes to those raw projections, with even larger changes and greater agreement between models. With this understanding, we consider our raw CMIP6 projections useful to tell the correct direction of change, even though its magnitude may be underestimated or overestimated.

b. CMIP6 projected changes in ENSO teleconnection over the TP consistent with CMIP3/5

Here, using CMIP6 models, we show that ENSO teleconnection over the TP enhances and shifts eastward under warming. These conclusions agree with previous studies based on CMIP3 and CMIP5. The consistency between the results we have obtained with those obtained previously using CMIP5 models and SST-forced AGCMs increases confidence. Power et al. (2013) projected reduced El Niño–induced rainfall variability in the western Pacific Ocean and enhanced rainfall variability in the central and eastern equatorial Pacific. Both El Niño–driven (Chung and Power 2015, 2016) and La Niña–driven (Chung et al. 2014; Chung and Power 2016) precipitation anomalies in the Pacific will increase. Wang et al. (2016) showed that the western Pacific Ocean has already warmed to levels that are unprecedented in the historical record. Intermittent disruptions to rainfall patterns and intensity over the Pacific Ocean (largely induced by ENSO) have major impacts on severe weather and agriculture in the tropics and beyond.

Power et al. (2017) show (using CMIP5 models) that humans may have contributed to the major disruption that occurred during the late twentieth century and suggest that elevated risk may remain for the twenty-first century. Power and Delage (2018) showed that ENSO-driven variability in the TP increases, which further
drives enhanced rainfall variability in 18 of the 25 regions during JJA and in 19 regions during DJF.

c. ENSO teleconnection changes compared with ENSO changes in CMIP6

In our study, we note that the strengthening of summer ENSO teleconnection across models (28/32 models agreed) is more robust than the increase in the Niño-3.4 ts variability (24/32 models agreed; Fig. 7b). Tropical ts variability changes are complex given oceanic and atmospheric processes, and the net effect of diverging feedbacks could possibly give less robust changes to the ts variability and a low model agreement. Previous findings showed that ENSO responses under warming are uncertain across varying emission scenarios and idealized simulations (Chen et al. 2017; Callahan et al. 2021; Cai et al. 2021; Brown et al. 2020). On the contrary, the ENSO-induced precipitation variability over the TP strengthens robustly. It involves mean-state changes beyond the bonds by ENSO itself, for example, the tropical rainfall variability is strongly related to mean atmospheric changes associated with the Clausius–Clapeyron relationship (Hu et al. 2021).

Previous studies (Power et al. 2013; Chung and Power 2014, 2015, 2016; Chung et al. 2014) conducted idealized SST-forced AGCM experiments and showed that the intensification of El Niño–driven anomalies in the Indo-Pacific Ocean primarily arises from a nonlinear interaction between background warming and El Niño–driven SST anomalies. The dominance of the nonlinear response and its consistency across scenarios explains why there can be a consensus among models and scenarios on projected changes in ENSO-driven precipitation variability in the absence of a consensus on changes in ENSO-driven SST variability.

On top of the background warming, ENSO amplitude does play a role in driving changes in the rainfall variability. In this study, we show that ENSO-induced rainfall variability is correlated with the ENSO amplitude across individual models (Figs. 5c,d). That is, models with strong ENSO teleconnection are those models with strong SST variability. We also show that future changes in ENSO teleconnection scale with...
changes in ENSO amplitude across models (Fig. 7b). It is consistent with previous results in the observation that the enhancement of ENSO teleconnection is modulated by the ENSO amplitude (Wang et al. 2020). Our results also agree with the understanding from the AGCM experiments in previous studies (Power et al. 2013; Chung and Power 2014, 2015, 2016; Chung et al. 2014). Their results showed that the impact of global warming on the ENSO-induced rainfall variability depends strongly on the amplitude of the El Niño SSTA. In all experiments, global warming enhances the tropical Pacific precipitation variability and shifts the longitude of the maximum variability toward the east. As the ENSO amplitude increases, the precipitation variability increases and the maximum response shifts further east. Beyond the tropics, Yeh et al. (2022) showed that future projections of ENSO temperature teleconnection over North America also have diversity due to confounding effects from climate change and the ENSO amplitude, and the role of ENSO intensity in temperature teleconnection diversity tends to be weaker for the higher-emission scenarios like SSP585 compared with the SSP126 in CMIP6.

d. CMIP6-informed understanding of the changes in ENSO teleconnection

Power et al. (2013) proposed a diagram (their Fig. 4) to illustrate the nonlinear interaction that contributes to the robust changes in ENSO-driven rainfall variability. They showed that, even without changes in ENSO amplitude, robust changes in background SST due to global warming (typical “El Niño like” warming with higher warming toward the east) could give rise to the changes in the ENSO-induced rainfall variability in the Pacific. They further diagnosed the

FIG. 17. ENSO–rainfall teleconnection zero crossing longitude and the rainfall center longitude. (a) Longitude of precipitation center against the zero crossing longitude of the precipitation covariance during the DJF season in the observation (black star), 32 CMIP6 model historical (blue) and future SSP585 scenario (red). Correlation between two axes and the p value are shown. (b) As in (a), but for the JJA season. (c) Future change (SSP585 scenario minus the historical period) in longitude of precipitation center against change in the zero crossing longitude of the precipitation covariance during the DJF season in the observation (black star), 32 CMIP6 model historical (blue) and future SSP585 scenario (red). (d) As in (c), but showing the JJA season.
component of rainfall changes and showed that the zonal shift (reflected in the reduced rainfall variability in the western Pacific and increased rainfall variability in the central Pacific) is mainly due to changes in the mean circulation dynamics (i.e., Walker circulation–related changes). These agree with the understanding that the eastward shift of the Walker circulation under warming is mainly associated with the El Niño–like warming pattern (Bayr et al. 2014). In addition to the sea surface temperature influences on the circulation changes, other studies (e.g., Yim et al. 2017) have shown that land–sea thermal contrast also contributes to changes in the Walker circulation simulated in atmospheric general circulation models.

Here in this study, even though our focus is the MC domain, we try to put information together to improve the understanding of the changes in the ENSO teleconnection over the MC and TP as a whole. We present a schematic diagram of the physical processes that contribute to future changes in the ENSO–rainfall teleconnection (Fig. 21). In the present
climate, the Walker circulation rising branch (also where the equatorial rainfall is centered) is located above the warm pool at EMC. When El Niño takes place, a weakened trade wind releases hot water from the warm pool to the east, creating anomalous warming at the central equatorial Pacific (CEP). Walker circulation moves to the east, creating the precipitation anomaly on the east and west side of the motion, that is, wetter in the CEP and drier in the EMC. Such rainfall variability creates a zonal dipole pattern with the zero crossing around the EMC.

Under warming, the equatorial Pacific is projected to have an “El Niño-like” warming pattern, with higher warming in the eastern equatorial Pacific (EEP). This SST pattern is associated with eastward-shifting Walker circulation (Bayr et al. 2014). With higher SST, the tropical precipitation is expected to enhance (governed by the Clausius–Clapeyron relationship), with
the center migrating to the east along with the eastward Walker circulation. When El Niño takes place, the physical processes for ENSO teleconnection are the same, but with increased mean tropical SST and rainfall. The nonlinear interactions result in both intensification of the ENSO-induced rainfall variability and eastward shift of the zero crossing longitude. Note that the eastward Walker circulation contributed via the component of mean circulation dynamics (Power et al. 2013).

e. Seasonal variations of ENSO teleconnection over the MC

In this study, we show that ENSO teleconnection over the MC varies across seasons (Fig. 2). The southern part of the MC is affected significantly during JJA, while the other areas are more pronounced during the DJF season. It is consistent with the finding of previous studies (Juneng and Tangang 2005; Supari et al. 2018) that El Niño–induced MC rainfall variability evolves northeastward, modulated by the two AC systems over the SIO and WNP.

We show that the El Niño–related negative precipitation variability over the CMC appears weaker after SON (Figs. 10b and 11b). It is consistent with the understanding that a background monsoon plays a role (Hendon 2003). During El Niño events, eastward shift of the convection center leads to anomalous subsidence in the MC (resulting in below average rainfall). The relationship is strongest during the period June through November, when anomalous surface easterlies across the MC enhance the evaporative cooling in the surrounding ocean. Toward the wet season in the boreal winter, the climatological winds shift northwesterly, reduce in speed, and dampen the local surface temperature cooling. Meanwhile, a patch of weak positive correlation appears within the CMC near Sumatra–Malay Peninsula (Fig. 2a) due to the anomalous anticyclone circulation near the Philippines (Li et al. 2017; Chang et al. 2004).

As to future changes, we note that ENSO teleconnection over the MC appears to change most during the summer compared with winter, especially the eastward shift of the zero crossing longitude (Fig. 13a) (for JJA, from 137° to 146°E shifting 9°; for DJF, not much change). With the eastward shift, ENSO-induced positive rainfall variability over the EMC is largely reduced (from 1.51 to 0.32 mm day$^{-1}$, reducing by 1.19 mm day$^{-1}$, compared with the insignificant change during DJF) (Fig. 14b). Why ENSO-induced rainfall variability shifts more during summer is a very interesting topic. Here, we simply diagnose the Walker circulation using the warm pool rainfall center. We find that the warm pool rainfall center also shifts further during summer (Fig. 16d) (for JJA, from 132° to 127°E shifting 5°; for DJF, insignificant change). It matches well with the seasonal reduction of the rainfall teleconnection over EMC (Fig. 14b) and the shift in the ENSO

### Table 5. Changes in ENSO–rainfall teleconnection-related quantities in hot, median, and cold models. The change is calculated using the value for 2065–99 in the SSP585 scenario minus that for 1980–2014 in the historical period. The second column is the future change based on the 32-CMIP6-model mean. The third column is the 15-hot-model mean. The fourth column is the 13-median-model mean. The fifth column is the 4-cold-model mean.

<table>
<thead>
<tr>
<th>Future changes in ENSO–rainfall teleconnection-related quantities in JJA</th>
<th>32-model mean</th>
<th>15-hot-model mean</th>
<th>13-median-model mean</th>
<th>4-cold-model mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>Increase in mean (ts$_{N34}$) (°C)</td>
<td>3.2</td>
<td>3.83</td>
<td>2.77</td>
<td>2.27</td>
</tr>
<tr>
<td>Increase in std (ts$_{N34}$) (°C)</td>
<td>0.14</td>
<td>0.19</td>
<td>0.1</td>
<td>0.05</td>
</tr>
<tr>
<td>Enhance in cov (ts$<em>{N34}$, pr$</em>{TP}$) (mm day$^{-1}$)</td>
<td>0.22</td>
<td>0.29</td>
<td>0.16</td>
<td>0.14</td>
</tr>
<tr>
<td>Enhance in cov (ts$<em>{N34}$, pr$</em>{CMC}$) (mm day$^{-1}$)</td>
<td>-0.33</td>
<td>-0.47</td>
<td>-0.2</td>
<td>-0.23</td>
</tr>
<tr>
<td>Eastward shift in cov (ts$<em>{N34}$, pr$</em>{eq}$) 0-cross lon (°)</td>
<td>9.75</td>
<td>11.1</td>
<td>9.23</td>
<td>6.38</td>
</tr>
<tr>
<td>Eastward shift in rainfall center lon (°)</td>
<td>5.4</td>
<td>7.3</td>
<td>3.6</td>
<td>4.5</td>
</tr>
</tbody>
</table>

**Figure 20. Intensification of ENSO–rainfall teleconnection reflected in correlation metrics.** (a) 12-month ENSO–precipitation correlation across the CMC. Observations (GPCP, black curve), the multimodel mean of 32 CMIP6 models for the historical period (blue curve), and the multimodel mean for the SSP585 scenario (red curve) are shown. The shading indicates the 95% model range. (b) As in (a), but for the EMC domain.
teleconnection (Fig. 13). Detailed investigations into the circulation will be left for further studies.

f. ENSO diversity influences on teleconnection

In this study, we show that El Niño– and La Niña–induced teleconnection are very similar in temporal evolution (Fig. 11) and spatial pattern (Fig. 15). In the observation, El Niño–induced rainfall variability is higher than La Niña–induced variability over the TP (Fig. 11). For the DJF peak season, La Niña–induced variability over the TP is 77% of El Niño–induced rainfall variability (Table 3). It is also true for the CMC domain at the OND peak season (La Niña induced variability is 67% of El Niño induced rainfall variability). In CMIP6 models, El Niño–induced rainfall variability is also higher than La Niña–induced variability over the TP and CMC.

As to the intensification under warming, we see a slight discrepancy between El Niño and La Niña (Fig. 11). Over the TP, future changes in the El Niño–induced rainfall variabilities at the DJF peak season (1.11–1.47 mm day$^{-1}$, 0.36 mm day$^{-1}$ increase) are slightly larger than changes in the La Niña–induced variabilities (−0.95 to −1.29 mm day$^{-1}$, 0.34 mm day$^{-1}$ enhancement of negative value). Over the CMC, changes in the El Niño–induced rainfall variabilities at OND of ENSO year 0 (from −0.62 to −0.72 mm day$^{-1}$, 0.1 mm day$^{-1}$ enhancement) are also slightly larger than changes in La Niña–induced rainfall variabilities (from 0.51 to 0.59 mm day$^{-1}$, 0.08 mm day$^{-1}$ enhancement).

We show that El Niño– and La Niña–induced zonal dipole structures are very similar (Fig. 15). As to the zero crossing longitude, we show that El Niño’s location is slightly more east than La Niña in the observation and in the models. For the eastward shifts during summer (JJA), El Niño teleconnection appears to shift more (139° to 150°E, shifting 11°) than La Niña (138° to 148°E, shifting 10°). For the eastward shifts during DJF, El Niño
teleconnection also appears to shift more (132° to 134'E, shifting 2°) than La Niña (130° to 131'E, shifting 1°). But note that the winter shift is not robust and within the noise range.

Besides the El Niño–La Niña asymmetry we addressed in this study, there are other aspects of ENSO diversity (Chen et al. 2017). ENSO events with varying magnitude and peak location could induce varying teleconnection impacts. For example, Salimun et al. (2014) showed that conventional and El Niño Modoki exerted different impact patterns on winter mean precipitation over Malaysia. Tangang et al. (2017) showed ENSO impacts on precipitation extremes over Malaysia depend on the ENSO intensity. Strong or moderate El Niño events show similar impacts. However, during DJF, strong and moderate La Niña events show opposite impacts, that is, moderate La Niña events enhance precipitation extremes over Malaysia (canonical teleconnection), while strong La Niña events reduce precipitation extremes. In our study, we mainly analyze 35-yr time snippets, which is short to give a robust measure of ENSO diversity. Meanwhile, CMIP6 models may not show reliable performance on modeled ENSO diversity. Therefore, investigation into ENSO teleconnection changes considering the ENSO diversity will be left for future studies.

g. Indian Ocean influences on the MC rainfall variability

The MC is located in between the Indian Ocean and the Pacific Ocean. On the one hand, the MC rainfall variability can be associated with the ENSO events in the Pacific Ocean (our study addressed this). On the other hand, the Indian Ocean dipole (IOD) is also an important driver. Like the ENSO, the IOD causes negative impacts on the MC rainfall (Amirudin et al. 2020). When the positive IOD takes place, the MC is drier than normal. Rainfall variability starts in the southern part of the MC during the JJA season and propagates northeastward during SON. When both IOD and ENSO co-occur, the negative impacts over the MC are more significant, especially over the southern part of Southeast Asia during the JJA and SON seasons. Changes in MC rainfall variability due to combined impacts of ENSO and IOD under warming are interesting topics for future studies.

6. Conclusions

Across the Indo-Pacific Ocean, ENSO induces a zonal dipole pattern of precipitation variability, that is, positive variability in the TP and horseshoe-shaped negative variability toward the MC (Figs. 1 and 2). In this study based on CMIP6 models, we address how ENSO influences the MC rainfall via teleconnection under warming. We focus on changes in the magnitude and spatial adjustment.

We first evaluate CMIP6 models in the historical period, and we show that models can produce reasonably realistic spatial patterns of the ENSO teleconnection (Fig. 3). But models underestimate the amplitude of ENSO-induced rainfall variability over the CMC (Fig. 4); this bias can be traced to the modeled TP teleconnection extending too far west into the MC, thus making the positive rainfall variability over the CMC overestimated. Most models are good overall, except for a few outliers. Models with weak ENSO teleconnection are models with weak ENSO amplitude (Fig. 5). It agrees with the understanding that the model performance in ENSO teleconnection is linked to the performance of ENSO itself (Cai et al. 2009).

Next, we analyze CMIP6 models under the SSP585 scenario (Fig. 6). Models suggest the teleconnection over the TP will strengthen with strong model agreement (Fig. 7a). Models projecting a stronger ENSO teleconnection over the TP are those models projecting a stronger SST variability (Fig. 7b). We also show that the teleconnection over the CMC will strengthen together with the teleconnection over the TP (Fig. 8), which indicates the zonal dipole teleconnection will strengthen as a whole (Fig. 9). Over the CMC, we show that the intensification of the ENSO teleconnection applies to both land and sea rainfall variability (Fig. 10). Composite analysis shows that both El Niño– and La Niña–induced rainfall variability will enhance across the ENSO cycle, and changes in the El Niño teleconnection are slightly larger than changes in La Niña teleconnection (Fig. 11).

Except for the intensification, we find that ENSO teleconnection will shift eastward under warming, with robust model agreement (Fig. 12). That is, the zero crossing of the dipole-shaped rainfall variability moves eastward. We also note that eastward shift of the ENSO teleconnection is more notable during summer than winter (Fig. 13), which results in the summer reduction of the positive ENSO teleconnection over the EMC (Fig. 14). Composite analysis shows that the El Niño teleconnection shifts slightly further than the changes in the La Niña teleconnection (Fig. 15). We further show that the eastward shift of teleconnection is seasonally linked to the shift in the mean rainfall center (indicative of the Walker circulation) (Fig. 16). Models with the teleconnection zero crossing toward the east are often models with the mean rainfall center toward the east (Fig. 17a). In addition, models projecting a further eastward shift in the ENSO teleconnection are generally those projecting a farther-eastward shift of the rainfall center (Fig. 17b). By merging the historical and SSP585 run into a long record, we create running time snippets to analyze the trend of changes across time. We show that the magnitude changes and the spatial changes in the teleconnection scale up with the warming trend despite natural variations (Fig. 18).

Note that we compute multimodel means to assess the average model behavior. We find that models with larger or smaller biases do not give distinctly different future changes compared with the rest of the models. It indicates that future projections are not largely sensitive to the models' historical biases. We also communicate the projections from three groups of models (“hot,” “median,” and “cold”) to account for the model uncertainty. We show that all three groups of models agree on the sign of change, which suggests that our conclusions are robust and meaningful (Fig. 19). We also note that hot models project a faster warming, a stronger ENSO, a stronger enhancement of ENSO teleconnection over the TP and CMC, a larger eastward shift of the ENSO teleconnection zero crossing, and a larger eastward shift of the warm pool.
FIG. A1. Summer (JJA) ENSO–precipitation correlation in the historical period. (top left) Observations and (top right) 32-model mean. The dashed black box indicates the domain of the CMC, and the dashed blue box indicates the domain of the TP.
These robust future changes in ENSO teleconnection over the MC have potential impacts. Stronger ENSO-induced rainfall variability could escalate drought across the CMC in an El Niño year and enhance flooding in a La Niña year under warming. A stronger correlation between CMC rainfall and ENSO also suggests an increase in the predictability of CMC rainfall (Fig. 20a). We also note the spatial inhomogeneity as to future changes. The EMC is wetter than normal during El Niño. Under warming, the EMC is projected to reduce the positive rainfall variability (Fig. 14b) and correlation.

**FIG. A2.** Summer (JJA) ENSO–precipitation covariance across the Indo-Pacific Ocean. The black curve is the observation, the blue curve is the 32-CMIP6-model mean for the historical period, and the red curve is the SSP585 scenario. The dashed line indicates the zero crossing longitude of the precipitation covariance.

---

rainfall center. These agree with our conclusions that changes scale with warming.
(Fig. 20b) along with the eastward shift of the ENSO teleconnection. These changes could trigger spatially inhomogeneous impacts on climate-relevant sectors.

Acknowledgments. The authors thank Bin Wang and Scott Power for insightful discussions and very helpful suggestions. This work was performed with the HPC facility at The National Supercomputing Centre (NSCC) Singapore. We acknowledge the World Climate Research Programme, which, through its Working Group on Coupled Modelling, coordinated and promoted CMIP5 and CMIP6. We thank the climate modeling groups for producing and making available their model output, the Earth System Grid Federation (ESGF) for archiving the data and providing access, and the multiple funding agencies who support CMIP and ESGF.

Data availability statement. All CMIP6 model data used in this study are available at the CMIP6 archive at https://pcmdi.llnl.gov/CMIP6/. Surface temperature observations from the HadISST dataset (1870–2017) are available at https://www.metoffice.gov.uk/hadobs/hadisst/. Precipitation observations from the GPCP dataset (1979–2017) are available at http://eagle1.umd.edu/GPCP_ICDR/. ERA5 datasets are available at (https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era5).

APPENDIX

Model Diversity and Sensitivity on Detrending

We show multimodel results in the main results. Here, in the appendix, we show diversity from individual models. Across 32 models, a couple of model outliers have unrealistic historical simulations and future projections different from the rest of the models. These models are MCM UA-1-0 (model 23, shown to have unrealistic ENSO responses in the tropics; Fig. A1) that projected weakened precipitation variability in both TP and CMC (Fig. 8b), and INM-CM4-8 (model 19, shown to have little ENSO-rainfall teleconnection; Fig. A2) that projected a westward shift in ENSO teleconnection (Fig. 12b). Setting aside these model outliers, there is enough agreement among models (with acceptable biases) to indicate a robust and meaningful future change.

We merge the historical and SSP585 runs to create a long record (1850–2099, 250 years). We use a 35-yr running window with a 1-yr shift to create 216 time snippets. For each time snippet, we conduct de-climatology and cubic-polynomial detrending. Here, we test detrending for the merged 1850–2099 period in the preprocessing before sliding time snippets. The results (Fig. A3) show that with 250-yr detrending, the summer rainfall teleconnection over TP changes from 0.59 to 0.81 mm day\(^{-1}\) in the warming period, which is very similar to the 35-yr detrending result (from 0.58 to 0.8 mm day\(^{-1}\)). Summer rainfall teleconnection over the CMC changes from \(-0.26\) to \(-0.58\) mm day\(^{-1}\) in the warming period, which is very similar to the 35-yr detrending result (from \(-0.26\) to \(-0.59\) mm day\(^{-1}\)). The comparison indicates that both ways of detrending give almost the same results and our conclusions still hold.

REFERENCES


