

1 **Tropical Pacific Climate Variability under Solar Geoengineering: Impacts on ENSO**
2 **Extremes**

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17 **Abstract**

18 Many modelling studies suggest that the El Niño Southern Oscillation (ENSO), in interaction
19 with the tropical Pacific background climate, will change with rising atmospheric greenhouse
20 gas concentrations. Solar geoengineering (reducing the solar flux from outer space) has been
21 proposed as a means to counteract anthropogenic climate change. However, the effectiveness
22 of solar geoengineering concerning a variety of aspects of Earth's climate is uncertain. Robust
23 results are particularly challenging to obtain for ENSO because existing geoengineering
24 simulations are too short (typically ~50-year) to detect statistically significant changes in the
25 highly variable tropical Pacific background climate. We here present results from a 1000-year
26 long solar geoengineering simulation, G1, carried out with the coupled atmosphere-ocean
27 general circulation model HadCM3L. In agreement with previous studies, reducing the solar
28 irradiance (4%) to offset global mean surface warming in the model more than compensates
29 the warming in the tropical Pacific that develops in the 4×CO₂ scenario. We see an
30 overcooling of 0.3°C and a 0.23-mm day⁻¹ (5%) reduction in mean rainfall over tropical
31 Pacific relative to preindustrial conditions in the G1 simulation, owing to the different
32 latitudinal distributions of the shortwave (solar) and longwave (CO₂) forcings. The location
33 of the Intertropical Convergence Zone (ITCZ) in the tropical Pacific, which moved 7.5°
34 southwards under 4×CO₂, is restored to its preindustrial position. However, other aspects of
35 the tropical Pacific mean climate are not reset as effectively. Relative to preindustrial
36 conditions, in G1 the time-averaged zonal wind stress, zonal sea surface temperature (SST)
37 gradient, and meridional SST gradient are each statistically significantly reduced by around
38 10%, and the Pacific Walker Circulation (PWC) is consistently weakened resulting in
39 conditions conducive to increased frequency of El Niño events. The overall amplitude of
40 ENSO strengthens by 9-10% in G1, but there is a 65 % reduction in the asymmetry between
41 cold and warm events: cold events intensify more than warm events. Notably, the frequency
42 of extreme El Niño and La Niña events increases by ca. 60% and 30%, respectively, while
43 the total number of El Niño events increases by around 10%. All of these changes are

1 statistically significant either at 95 or 99% confidence level. Somewhat paradoxically, while
2 the number of total and extreme events increases, the extreme El Niño events become weaker
3 relative to the preindustrial state while the extreme La Niña events become even stronger.
4 That is, such extreme El Niño events in G1 become less intense than under preindustrial
5 conditions, but also more frequent. In contrast, extreme La Niña events become stronger in
6 G1, which is in agreement with the general overcooling of the tropical Pacific in G1 relative
7 to preindustrial conditions.

8 **1 Introduction and Background**

9 Since the industrial revolution, anthropogenic emissions of Greenhouse Gases (GHGs) have
10 led to globally increasing surface temperatures (Stocker, 2013). Higher temperatures, in turn,
11 and more generally a rapidly changing climate, can have adverse effects on humans, plants,
12 and animals through changes in various ecosystems, rising sea levels, melting glaciers, and
13 could significantly impact the frequency and intensity of extreme weather events (Moore et
14 al., 2015). Various strategies, principally a reduction of GHG emissions and enhancements of
15 carbon dioxide sinks (Pachauri et al., 2014), have been proposed to mitigate anthropogenic
16 climate change. Another group of strategies involves the intentional modification of Earth's
17 radiation balance on a global scale, known as solar geoengineering (Crutzen, 2006; Wigley,
18 2006; Curry et al., 2014). For any serious consideration of such geoengineering strategies, it
19 is essential to understand their potential perils as well as benefits. One route to study the
20 potential impacts of geoengineering on various components of Earth's climate system (e.g.,
21 atmosphere, ocean, cryosphere, etc.) is through employing state-of-the-art coupled
22 atmosphere-ocean general circulation models (AOGCMs).

23 In this context, Kravitz et al. (2011) proposed the Geoengineering Model Intercomparison
24 Project (GeoMIP), which initially consisted of a set of four experiments (viz. G1, G2, G3,
25 and G4). These experiments are designed to investigate the effects of geoengineering on the
26 regional and global climate when it is implemented to offset the annual mean global radiative
27 forcing at the top of the Earth's atmosphere introduced by GHGs. These experiments are
28 collectively called Solar Radiation Management (SRM) or solar geoengineering (Kravitz et
29 al., 2013a). In the G1 experiment, atmospheric CO₂ is instantaneously quadrupled, but the
30 global GHG-induced longwave radiative effects are offset by a simultaneous reduction in the
31 shortwave Total Solar Irradiance, TSI, (Kravitz et al., 2011). In terms of radiative forcing, the
32 quadrupling of CO₂ is similar to the year 2100 in the RCP8.5 emission scenario
33 (Representative Concentration Pathway with a radiative forcing of 8.5 W m⁻² by the year
34 2100; Schmidt et al., 2012). In this paper, we focus on the G1 experiment to investigate how
35 effectively solar geoengineering could mitigate the effects of substantial changes in
36 atmospheric CO₂ on the tropical Pacific climate.

37 The El Niño Southern Oscillation (ENSO) is an important coupled ocean-atmosphere mode
38 of interannual variability in the tropical Pacific (Park et al., 2009; Vecchi and Wittenberg
39 2010), which affects both regional and global climate (see Ropelewski and Halpert, 1987;
40 Bove et al., 1998; Malik et al., 2017). ENSO oscillates between a warm, El Niño, and a cold,
41 La Niña, phase every 2-7-year (Santoso et al., 2017). Based on Empirical Orthogonal

1 Function Analysis (EOF) of Sea Surface Temperature (SST) in the tropical Pacific (see
2 Takahashi et al., 2011), ENSO can be contrasted into two distinct modes of variability, i.e.
3 eastern and central Pacific ENSO modes (Kao and Yu, 2009; Yu and Kim, 2010; Xie and Jin,
4 2018). The eastern Pacific ENSO mode (EOF1) shows maximum SST anomaly in the eastern
5 equatorial Pacific (Niño3 region: 5° N-5° S; 150° W-90° W) whereas the central Pacific ENSO
6 mode (EOF2) indicates maximum SST anomaly in the central Pacific (Niño4 region: 5° N-5°
7 S; 160° E-150° W) (Kao and Yu, 2009; Cai et al., 2018).

8 As diagnosed from SST indices in state-of-the-art AOGCMs, there was no intermodel
9 consensus about change in frequency of ENSO events and amplitude in a warming climate
10 (Vega-Westhoff and Sriviver, 2017; Yang et al., 2018). However recently, Cai et al. (2018),
11 using SST indices based on Principal Component Analysis (PCA), showed an enhanced
12 frequency of extreme El Niño events and strengthening of ENSO amplitude under increased
13 GHG forcing. However, before that, Cai et al. (2014 and 2015b) also showed evidence of a
14 doubling of El Niño and La Niña events in the Coupled Model Intercomparison Project
15 (CMIP) phases 3 (A2 scenario) and 5 (RCP8.5) by investigating a performance-based subset
16 of models using rainfall-based ENSO indices instead of SST-based indices. Similarly, Wang
17 et al. (2017) also reported a doubling of extreme El Niño events, relative to the preindustrial
18 level, in the RCP2.6 transient scenario a century after stabilization of global mean
19 temperature. Chen et al. (2017), analysing 20 CMIP5 models (RCP8.5), found both
20 strengthening (in 6 models) and weakening (in 8 models) of ENSO amplitude. However, Cai
21 et al. (2018) later found robust evidence of a consistent increase in El Niño amplitude in the
22 subset of CMIP5 climate models, which were capable of simulating both eastern and central
23 Pacific ENSO modes. In summary, changes in ENSO characteristics such as amplitude and
24 ENSO extremes are projected in a warming climate (e.g., Cai et al., 2014, 2015b, 2018; Kim
25 et al., 2014; Wang et al., 2018).

26 Increasing GHGs have distinct effects on the tropical Pacific mean climate. In CMIP3 and
27 CMIP5 simulations, the equatorial tropical Pacific consistently shows a significant mean state
28 warming response to increased GHG forcing (van Oldenborgh et al., 2005; Collins et al.,
29 2010; Vecchi and Wittenberg 2010; Huang and Ying 2015; Luo et al., 2015). CMIP3 and
30 CMIP5 models generally show more warming on than off-equatorial tropical Pacific (Liu et
31 al., 2005; Collins et al., 2010; Cai et al., 2015a). Consistent with these warming patterns,
32 studies typically found a weakening of zonal SST gradient (ZSSTG), Pacific Walker
33 Circulation (PWC), zonal wind stress, and a shoaling of the equatorial tropical Pacific
34 thermocline (see van Oldenborgh et al., 2005; Latif et al., 2009; Park et al., 2009; Yeh et al.,
35 2009; Collins et al., 2010; Kim et al., 2014; Cai et al., 2015a; Zhou et al., 2015; Coats and
36 Karnauskas 2017; Vega-Westhoff and Sriviver 2017). Changes in the mean state of the tropical
37 Pacific can bring about variations in ENSO properties such as amplitude, frequency, and
38 spatial pattern (Collins et al., 2010; Vecchi and Wittenberg, 2010; Cai et al., 2015a).

39 We note that a previous study by Guo et al. (2018) found no statistically significant change in
40 the intensity of Walker Circulation in GeoMIP models when comparing preindustrial
41 simulations to the G1 experiment. Similarly, Gabriel and Robock (2015) found no
42 statistically significant change in frequency and amplitude of ENSO events under both global

1 warming and geoengineering scenarios in 6 GeoMIP models that captured ENSO variability
2 best. However, these authors themselves highlighted the length of their simulations (~50
3 years) as a key constraint for their studies. They suggested that long term simulations (>50
4 years) would be required to detect possible ENSO changes. Guo et al. (2018) concluded that
5 60 or more years of model simulations are required to detect changes in the PWC, while
6 Vecchi et al. (2006) and Vecchi and Soden (2007) argued that 130-yr are necessary to
7 identify any robust change in the PWC (Gabriel and Robock, 2015). Similarly, Stevenson et
8 al. (2010) estimated that 250 years are needed to detect changes in ENSO variability with a
9 statistical significance of 90%. Here we aim to address this gap in the literature and establish
10 a baseline for future studies through the analysis of long-term (1000 year) simulations of a
11 single climate model.

12 Here, we employ three 1000-year long climate model simulations (preindustrial forcing,
13 abrupt-4xCO₂ forcing, and G1) to estimate the efficacy of solar geoengineering in resetting
14 the tropical Pacific circulation. Specifically, we investigate: (1) if solar geoengineering can
15 mitigate the changes in mean tropical Pacific climate found in previous GHG warming
16 studies, and even bring it back to the preindustrial conditions; (2) if ENSO frequency and
17 amplitude are different under G1 conditions than under preindustrial simulations; and (3) if
18 the G1 experiment reduces the increase in the frequency of extreme ENSO events, as shown
19 by Cai et al. (2014, 2015b and 2018), under increased GHG forcing, relative to the
20 preindustrial state. For this purpose, we are primarily interested in the more subtle differences
21 in climate between G1 and preindustrial conditions, but also consider the profound changes
22 under 4xCO₂ where, by design, the global mean surface temperature is much higher, and thus
23 many other climate aspects vastly differ from the other two scenarios.

24 Section 2 describes the climate model HadCM3L, the data and the statistical methods used to
25 detect changes in tropical Pacific and ENSO variability. The same section also evaluates the
26 capability of HadCM3L to model ENSO. Section 3 evaluates the response of a list of metrics
27 used to understand how the mean state and ENSO variability are affected in different
28 experiments (preindustrial, 4xCO₂, G1). Section 4 elaborates on the mechanism of ENSO
29 variability under GHG forcing and solar geoengineering for the given model system. Finally,
30 Section 5 presents the discussion and conclusions.

31 **2 Data and methods**

32 **2.1 Climate model**

33 HadCM3L (Cox et al., 2000) has a horizontal resolution of 2.5° latitude × 3.75° longitude
34 (~T42) with 19 (L19) atmospheric and 20 (L20) ocean levels. HadCM3L stems from the
35 family of HadCM3 climate models; the only difference is lower ocean resolution (HadCM3:
36 1.25° × 1.25°; Valdes et al., 2017). In HadCM3L, land surface processes are simulated by the
37 MOSES-2 module (Essery and Clark, 2003; Cao et al., 2016). HadCM3L does not include an
38 interactive atmospheric chemistry scheme and thus does not consider effects of ozone
39 changes on ENSO amplitude and surface warming under 4xCO₂ (e.g., Nowack et al., 2015;
40 2017, 2018) or G1 (e.g., Nowack et al., 2016). Instead, we use preindustrial background

1 ozone climatology, prescribed on pressure levels. In section 2.4, we evaluate the ability of
2 HadCM3L to model ENSO. We acknowledge that some of our results will necessarily be
3 model-dependent, and underline the need for similar studies with other climate models. Still,
4 by using much longer simulations than used previously, our results provide statistical
5 robustness for the given model system.

6 **2.2 Simulations and observational data**

7 Here, we use HadCM3L simulations carried out by Cao et al. (2016). To achieve a quasi-
8 equilibrium preindustrial climate state, the model was spun up for 3000 years with constant
9 CO₂ concentrations (280 ppmv; parts per million by volume) and TSI (1365 W m⁻²). Then,
10 three 1000-year long experiments were carried out, starting from this preindustrial climate
11 state. These experiments are: (1) the preindustrial control (piControl) experiment with
12 constant values of CO₂ (280 ppmv) and TSI (1365 W m⁻²); (2) a quadrupled CO₂ (4×CO₂)
13 experiment in which CO₂ is suddenly increased to 1120 ppmv; and (3) sunshade
14 geoengineering (G1) experiment where the radiative effects of the instantaneously
15 quadrupled CO₂ are offset by simultaneously reducing TSI (by 4%). All experiments follow
16 the GeoMIP protocol (see Kravitz et al., 2011); the only difference being that simulations
17 were run for 1000 years (see Cao et al., 2016) instead of 50 years as in GeoMIP.

18 The monthly SST dataset from HadISST (1° latitude × 1° longitude; Rayner et al., 2003) and
19 the rainfall data from the Global Precipitation Climatology Project (GPCP; Adler et al., 2003)
20 version 2.3 (2.5° latitude × 2.5° longitude) over the period 1979-2017 are used to provide
21 observational constraints and to identify the rainfall threshold to be used for defining extreme
22 El Niño events. Further, we use ERA5 reanalysis data (Copernicus Climate Change Service
23 (C3S), 2017) covering years 1979-2019 to evaluate the capability of HadCM3L to simulate
24 ENSO variability. ERA5 has a horizontal resolution of 0.25° latitude × 0.25° longitude.
25 Specifically, we use monthly mean surface latent heat flux (lh), sensible heat flux (sh), net
26 shortwave radiation flux (sw), net longwave radiation flux (lw), ocean temperature, and zonal
27 and meridional components of wind stress.

28 **2.3 Definitions and statistical tests**

29 We analyse changes in the tropical Pacific (25° N-25° S; 90° E-60° W) mean climate. We
30 present climatologies for SSTs, rainfall, Intertropical Convergence Zone (ITCZ), vertical
31 velocity averaged between 500 and 100 hPa (Omega500-100), PWC, zonal wind stress, zonal
32 and meridional SST gradients (ZSSTG and MSSTG, respectively), and thermocline depth.
33 We calculate mean climatological differences for all these variables simulated under 4×CO₂
34 and G1 relative to piControl and assess their statistical significance using non-parametric
35 Wilcoxon signed-rank and Wilcoxon rank-sum tests (Hollander and Wolfe, 1999; Gibbons
36 and Chakraborti, 2011). All analyses are performed on re-gridded (2° longitude × 2.5°
37 latitude) HadCM3L output for model years 11 to 1000 unless otherwise stated. The first ten
38 years are skipped to remove the initially significant atmospheric transient effects stemming
39 from instantaneously increasing CO₂ (see Kravitz et al., 2013b; Hong et al., 2017). Since
40 ENSO events peak in boreal winter (December-January-February; DJF; Cai et al., 2014;

1 Gabriel and Robock 2015; Santoso et al., 2017), the entire analysis is performed for DJF,
2 unless otherwise stated. Accordingly, we also analyse mean state changes in the tropical
3 Pacific during boreal winter.

4 Both rainfall and SST-based ENSO indices are used in the present study. Niño3 (5° N- 5° S;
5 150° W- 90° W) and Niño4 (5° N- 5° S; 160° E- 150° W) indices are defined by averaging SST
6 over corresponding ENSO regions. Normalised ENSO anomalies (i.e., the ENSO indices) are
7 calculated relative to piControl mean and standard deviation (s.d.) and are quadratically
8 detrended before analysis. The Niño3 index is chosen for studying the characteristics of
9 extreme El Niño events since during an extreme El Niño event, following the highest SSTs,
10 convective activity moves towards the eastern Pacific, and the ITCZ moves over the Niño3
11 region resulting in rainfall higher than 5 mm day^{-1} (Cai et al., 2014). Similar to Cai et al.
12 (2014, 2017), events with Niño3 rainfall greater than 5 mm day^{-1} are considered extreme El
13 Niño events, whereas events with Niño3 SST index greater than 0.5 s.d. and Niño3 rainfall
14 less than 5 mm day^{-1} are defined as moderate events unless otherwise stated. The Niño4 index
15 is chosen for studying the characteristics of extreme La Niña events since maximum cold
16 temperatures occur in this region (Cai et al., 2015a, 2015b). La Niña extreme ($\text{Niño4} < -1.75$
17 s.d.), moderate ($-1 > \text{Niño4} > -1.75$), and weak ($-0.5 > \text{Niño4} > -1$) events are defined
18 following Cai et al. (2015b). These definitions classify the 1988 and 1998 La Niñas in
19 observations as extreme events (see Cai et al., 2015b), and HadCM3L can capture such
20 extreme anomalies (see Sect. 3.2), which allows us to study changes in their number and
21 magnitude.

22 To understand the mechanisms responsible for changes in ENSO variability, we have
23 calculated ENSO feedbacks (e.g., Bjerkness (BJ) and heat flux (hf) feedbacks) and ocean
24 stratification. BJ feedback is a dynamical response of equatorial zonal wind stress to
25 equatorial SST anomalies. It is positive feedback that maintains the ZSSTG (Lloyd et al.,
26 2011). Here, we calculate the BJ feedback by point-wise linear regression (Bellenger et al.,
27 2014) of the zonal wind stress anomalies over the entire equatorial Pacific (5° N- 5° S; 120° E-
28 80° W; Kim and Jin 2011a; Ferret and Collins 2019) onto the eastern equatorial Pacific (5° N-
29 5° S; 180° W- 80° W; Kim and Jin 2011a; Ferret and Collins 2019) SST anomalies. We then
30 define the BJ feedback as the mean regression coefficient (Bellenger et al., 2014) over the
31 eastern equatorial Pacific region. The hf feedback is a regression coefficient calculated by
32 point-wise linearly regressing the net surface heat flux (sum of sw, lw, lh, and sh) anomalies
33 into the ocean onto the SST anomalies over the eastern equatorial Pacific (5° N- 5° S; 180° W-
34 80° W; Kim and Jin 2011a). This regression coefficient is also termed as a thermal damping
35 coefficient (Kim and Jin, 2011a). It is negative feedback in which an initial positive SST
36 anomaly causes a reduced surface net heat flux into the ocean, thus lessening the initial SST
37 anomaly (Lloyd et al., 2011). Ocean stratification is defined as the difference in the
38 volumetric average of ocean temperatures over the upper 67 m, and the temperature of a
39 single ocean layer at 95 m, both spatially averaged over the region, 5° N- 5° S; 150° E- 140°
40 W, where strong zonal wind stress anomalies also occur (see Fig. 4a and Fig. S1; Cai et al.,
41 2018).

1 Following Cai et al. (2014), the statistical significance of the change in the frequency of
2 ENSO events is tested using a bootstrap method with 10,000 realisations for the piControl
3 data. We then find the s.d. of events over these 10,000 realisations. If the difference of events
4 of piControl with 4xCO₂ and G1 is larger than 2 s.d., the change in frequency is considered
5 statistically significant. The same method is used for testing the statistical significance of a
6 change in ENSO amplitude, ZSSTG, MSSTG, ENSO amplitude asymmetry, ENSO
7 feedbacks, and ocean stratification. All changes in 4xCO₂ and G1 are described relative to
8 piControl.

9 **2.4 ENSO representation in HadCM3L**

10 Before employing HadCM3L for studying ENSO variability under 4xCO₂, and G1, we
11 evaluate its piControl simulation against present-day observational data. There is a non-linear
12 relationship between tropical Pacific SST and rainfall (Ham, 2017), which can be diagnosed
13 by Niño3 region rainfall skewness (Cai et al., 2014). Skewness is a measure of asymmetry
14 around the mean of the distribution (see eq. S1). Positive skewness means that in given data
15 distribution, the tail of the distribution is spread out towards high positive values, and vice
16 versa (Ghandi et al., 2016). The skewness criterion is used to exclude climate models
17 simulating overly wet or dry conditions over the Niño3 region (Cai et al., 2017). During
18 extreme El Niño events, the ITCZ moves equatorward, causing significant increases in
19 rainfall ($> 5 \text{ mm day}^{-1}$) over the eastern equatorial Pacific that skews the statistical
20 distribution of rainfall in the Niño3 region. Thus, for studying extreme ENSO events, the
21 model should be capable of simulating Niño3 rainfall above 5 mm day^{-1} and Niño3 rainfall
22 skewness of greater than 1 over the entire simulated period (see our Sect. 3.2.2, and Cai et al.,
23 2014 and 2015b). With a Niño3 rainfall skewness of 2.06 for piControl, HadCM3L fulfils
24 this criterion.

25 In addition, we evaluate the ENSO modelled by HadCM3L following a principal component
26 (PC) approach suggested by Cai et al. (2018). Considering distinct eastern and central Pacific
27 ENSO regimes based on EOF analysis, they found that climate models capable of simulating
28 present-day ENSO diversity show a robust increase in eastern Pacific ENSO amplitude in a
29 greenhouse warming scenario. Specifically, the approach assumes that any ENSO event can
30 be represented by performing EOF analysis on monthly SST anomalies and combining the
31 first two principal patterns (Cai et al., 2018). The first two PCs time series, PC1 and PC2,
32 show a non-linear relationship in observational datasets (Fig. S1m). Climate models that do
33 not show such a non-linear relationship cannot satisfactorily simulate ENSO diversity, and
34 hence are not sufficiently skilful for studying ENSO properties (Cai et al., 2018). Here, we
35 perform EOF analysis on quadratically detrended monthly SST and wind stress anomalies of
36 ERA5 and piControl over a consistent period of 41-year. We evaluate HadCM3L's ability to
37 simulate two distinct ENSO regimes and the non-linear relationship between the first two
38 PCs, i.e., $\text{PC2}(t) = \alpha[\text{PC1}(t)]^2 + \beta[\text{PC1}(t)] + \gamma$ (Fig. S1). From ERA5, $\alpha = -0.36$ (statistically
39 significant at 99% confidence level, hereafter “cl”) whereas in piControl $\alpha = -0.31$ (99% cl),
40 which is same as the mean $\alpha = -0.31$ value calculated by Cai et al. (2018) averaged over five
41 reanalysis datasets. The 1st and 2nd EOF patterns of monthly SST and wind stress anomalies
42 of piControl (Fig. S1 b, e) are comparable with that of ERA5 (Fig. S1 a, d). EOF1 of

1 piControl shows slightly stronger warm anomalies in the eastern equatorial Pacific, whereas
2 negative anomalies over the western Pacific are slightly weaker compared to ERA5. In
3 EOF1, the stronger wind stress anomalies occur to the west of the Niño3 region, which is a
4 characteristic feature during the eastern Pacific El Niño events (see Kim and Jin, 2011a).
5 Compared to ERA5, the spatial pattern of warm eastern Pacific anomalies is slightly stretched
6 westwards, and wind stress anomalies are relatively stronger over the equator and South
7 Pacific Convergence Zone (SPCZ). The 2nd EOF, in both ERA5 and piControl, shows warm
8 SST anomalies over the equatorial central Pacific Niño4 region. The variance distributions
9 for ERA5 and HadCM3L match well for EOF1 (ERA5: 82%, piControl: 90%) whereas a large
10 difference exist for EOF2 (ERA5: 18%, piControl: 10%).

11 The PCA is also useful for evaluating how well HadCM3L represents certain types of ENSO
12 events. Eastern and central Pacific ENSO events can be described by an E-Index $(PC1-$
13 $PC2)/\sqrt{2}$; Takahashi et al., 2011), which emphasises maximum warm anomalies in the eastern
14 Pacific region (Cai et al., 2018), and a C-Index $(PC1+PC2)/\sqrt{2}$; Takahashi et al., 2011)
15 respectively, which focuses on maximum warm anomalies in the central Pacific (Cai et al.,
16 2018). Here, we show the eastern Pacific (EP) Pattern (Fig. S1 g, h) and central Pacific (CP)
17 pattern (Fig. S1 j, k) by linear regression of mean DJF E- and C-Index, respectively, onto
18 mean DJF SST and wind stress anomalies. We find that model's EP and CP patterns agree
19 reasonably well with that of ERA5. HadCM3L underestimates the E-index skewness (1.16)
20 whereas overestimates the C-Index skewness (-0.89) compared to ERA5 (2.08 and -0.58,
21 respectively) averaged over DJF. HadCM3L's performance averaged over the entire
22 simulated period of piControl is also consistent with ERA5 (Fig. S1; α : -0.32, EOF1: 64%,
23 EOF2, 8%, E-index skewness: 1.30, C-index skewness: -0.42). In general, in HadCM3L, the
24 contrast between the E- and C-index skewness over the entire simulated period is sufficient
25 enough to differentiate relatively strong warm (cold) events in the eastern (central) equatorial
26 Pacific compared to the central (eastern) equatorial Pacific. Finally, we also evaluated the hf
27 and BJ feedbacks which, for piControl, are very similar to those of ERA5 (Table S5-6).

28 We conclude that HadCM3L has a reasonable skill for studying long-term ENSO variability
29 and its response to solar geoengineering. However, we also highlight the need for and hope to
30 motivate future modelling studies that will help identify model dependencies in the ENSO
31 response.

32 **3 Results**

33 **3.1 Changes in the tropical Pacific mean state**

34 In this section, we analyse several significant changes in the tropical Pacific mean state under
35 $4xCO_2$ and G1. In particular, we look into meridional and zonal SST changes, corresponding
36 surface wind responses, and coupled variations in the thermocline depth. Our analysis reveals
37 that this leads to significant changes in the precipitation climatology among the simulations.
38 Finally, we find consistent effects on the PWC. All these results are important not just as
39 general climatic features but also because they are mechanistically linked to changes in
40 ENSO extremes discussed in detail in Sect. 3.2.

1 **3.1.1 Sea surface temperature**

2 Tropical Pacific SSTs are spatially asymmetric along the equator. The western equatorial
3 Pacific (warm pool) is warmer on average than the eastern equatorial Pacific (cold tongue)
4 (Vecchi and Wittenberg, 2010). The piControl simulation (Fig. 1a) reasonably simulates the
5 SST asymmetry between the western and eastern equatorial Pacific well (cf. Fig 1a in Vecchi
6 and Wittenberg, 2010). Under $4\times\text{CO}_2$, the SST zonal asymmetry is significantly reduced
7 (Fig. 1b), and the entire equatorial tropical Pacific shows a warming state (e.g., Meehl and
8 Washington, 1996; Boer et al., 2004). The solar dimming in G1 largely offsets the warming
9 seen under $4\times\text{CO}_2$ and brings the tropical Pacific mean SSTs close to the preindustrial state
10 (Fig. 1c). The SPCZ, where the highest SSTs of the warm pool occur (Cai et al., 2015a; blue
11 line in Fig. 1a), moves towards the equator under $4\times\text{CO}_2$ (blue line, Fig. 1b), but returns to
12 approximately its preindustrial position in G1 (Fig. 1c).

13 The tropical Pacific is 3.90°C warmer in $4\times\text{CO}_2$ but 0.30°C colder in G1, with both
14 differences being significant at the 99% cl (see Fig. 1d-e, Table S1). The Pacific cold tongue
15 warms more rapidly than the Pacific Warm Pool under $4\times\text{CO}_2$. In contrast, in G1, a stronger
16 cooling occurs in the Pacific Warm Pool and the SPCZ than in the cold tongue region. The
17 Pacific Warm Pool is $\sim 0.4\text{-}0.6^\circ\text{C}$ colder in G1, whereas the east Pacific cools less ($\sim 0.2^\circ\text{C}$
18 in the Niño3 region), indicating a change in SST asymmetry under G1.

19 Our SST results under $4\times\text{CO}_2$ qualitatively agree with previous studies (Liu et al., 2005; van
20 Oldenborgh et al., 2005; Collins et al., 2010; Vecchi and Wittenberg, 2010; Cai et al., 2015a;
21 Huang and Ying, 2015; Luo et al., 2015; Kohyama et al., 2017; Nowack et al., 2017).
22 Overcooling of the tropics (and as such, the tropical Pacific) is a robust signal in G1
23 simulations, even short ones, simply due to the different meridional distribution of shortwave
24 and longwave forcing (Govindasamy and Caldeira, 2000; Lunt et al., 2008; Kravitz et al.,
25 2013b; Curry et al., 2014; Nowack et al., 2016). The results presented here based on a long
26 simulation not only corroborate previously published findings but also statistically
27 demonstrate that under G1, the Warm Pool and SPCZ cool faster than the cold tongue.

28 **3.1.2 Precipitation**

29 In the tropical Pacific, there are three dominant bands of rainfall activity: one in the western
30 Pacific Warm Pool, one in the SPCZ, and the last one along the ITCZ situated at around 8°N
31 and 150°W - 90°W . Further, the eastern equatorial Pacific is relatively dry compared with
32 these three rainy bands (cf. Fig. 2a Sun et al. 2020). Under piControl, HadCM3L simulates
33 these spatial rainfall patterns well, with maxima of $\sim 6\text{-}8$, $\sim 12\text{-}14$, and $\sim 8\text{-}10\text{ mm day}^{-1}$ over
34 the Pacific Warm Pool, the SPCZ, and the ITCZ, respectively (Fig. 2a). Under $4\times\text{CO}_2$, the
35 spatial rainfall pattern changes significantly. The ITCZ moves equatorward, and the SPCZ
36 becomes zonally oriented (blue line, Fig. 2b). The rainfall asymmetry between the western
37 and eastern equatorial Pacific decreases under $4\times\text{CO}_2$. Precipitation migrates from the west
38 Pacific to the Niño3 region, with maximum rainfall at $\sim 145^\circ\text{W}$. The reduced zonal
39 asymmetry in the rainfall between western and eastern Pacific is effectively restored to the
40 preindustrial state in G1 (Fig. 2c).

1 A statistically significant (99% cl) overall precipitation increase of 0.21 mm day^{-1} (+5%) is
2 seen over the tropical Pacific under $4\times\text{CO}_2$ (Fig. 2d). In contrast, the mean rainfall in G1
3 decreases by 0.23 mm day^{-1} (-5%; Fig. 2e), consistent with the simulated reduction in
4 temperature ($-0.30 \text{ }^\circ\text{C}$) over the tropical Pacific. However, there is a strong regional structure:
5 under $4\times\text{CO}_2$, rainfall decreases to a maximum of $\sim 3 \text{ mm day}^{-1}$ over parts of the Pacific
6 Warm Pool and off-equatorial regions, whereas a significant increase of $\sim 15\text{-}18 \text{ mm day}^{-1}$
7 develops over the Niño3 region. An overall increase in mean rainfall under the GHG
8 warming scenario has also been reported in many previous studies (e.g., Watanabe et al.,
9 2012; Power et al., 2013; Chung et al., 2014; Nowack et al., 2016). Under G1, rainfall
10 decreases over the Pacific Warm Pool, SPCZ, and ITCZ regions. In contrast, rainfall
11 increases significantly over most parts of central and eastern equatorial Pacific, with a
12 maximum ($\sim 1.5\text{-}2 \text{ mm day}^{-1}$) centred at $\sim 150^\circ \text{ W}$ (Fig. 2e). Kravitz et al. (2013b) reported a
13 decrease of 0.2 mm day^{-1} over the tropical regions. Under G1, the magnitude of the lapse rate
14 decreases, resulting in increased atmospheric stability and hence suppressed convection,
15 which leads to an overall reduction of rainfall over the tropics (Bala et al., 2008; Kravitz et
16 al., 2013b).

17 The position of the ITCZ over the tropical Pacific ($25^\circ \text{ N}\text{-}25^\circ \text{ S}$; $90^\circ \text{ E}\text{-}60^\circ \text{ W}$) is calculated by
18 finding the latitude of maximum rainfall (blue lines, Fig. 2a-e). The median position of this
19 maximum ITCZ (from $154^\circ \text{ W}\text{-}82^\circ \text{ W}$) is 7.5° N , 0° , and 7.5° N under piControl, $4\times\text{CO}_2$, and
20 G1, respectively. Thus, under $4\times\text{CO}_2$, the ITCZ mean position shifts over the equator and is
21 positioned within the Niño3 region. G1 restores the ITCZ and SPCZ to their preindustrial
22 orientations. Still, differences in the magnitude of rainfall persist over these regions, as well
23 as over the Pacific Warm Pool (Fig. 2a, c, e). That is, while the relative additional rainfall
24 asymmetry between the western and eastern Pacific in $4\times\text{CO}_2$ is mostly resolved in G1, the
25 tropical Pacific is overall wetter under $4\times\text{CO}_2$ but drier in G1.

26 **3.1.3 Zonal wind stress**

27 Changes in zonal wind stress are directly dependent on and interact with ENSO amplitude
28 (Guilyardi, 2006), ENSO period (Zelle et al., 2005; Capotondi et al., 2006), and ZSSTG (Hu
29 and Fedorov, 2016). A positive feedback loop between zonal wind stress, SST, and
30 thermocline depth influences the evolution of ENSO (Philip and van Oldenborgh, 2006). A
31 decrease in the strength of the trade winds is concurrent with a flattening of the thermocline,
32 a reduction of upwelling in the eastern Pacific, and increased SST in the eastern relative to
33 the western equatorial Pacific, thus resulting in further weakening of the trade winds (Collins
34 et al., 2010). We use the zonal wind stress index, Westerly Wind Bursts (WWBs), and
35 Easterly Wind Bursts (EWBs) to study the wind stress over the tropical Pacific. The zonal
36 wind stress index is defined as the wind stress averaged over the equatorial tropical Pacific
37 ($5^\circ \text{ N}\text{-}5^\circ \text{ S}$; $120^\circ \text{ E}\text{-}80^\circ \text{ W}$). Although here not explicitly diagnosed through daily data, WWBs
38 and EWBs are contained respectively in the positive and negative values of this wind stress
39 index (see Hu and Fedorov, 2016). As the duration of WWBs is 5 to 40 days (Gebbie et al.,
40 2007), the monthly mean data of westerly wind stress includes a monthly average of these
41 bursts.

1 We find that the zonal wind stress is significantly reduced over most parts of the tropical
2 Pacific, especially over the Niño3 region in both 4×CO₂ and G1 (Fig. 3a-e), in agreement
3 with the reduced zonal SST gradients in both scenarios (Fig. 1). The zonal wind stress
4 weakens by 31% and 10% in 4×CO₂ and G1 (statistically significant at 99% cl; Fig. 4a),
5 respectively. We also see a considerable weakening of zonal wind stress over the Niño3
6 region, both under 4×CO₂ and G1. The strength of WWBs increases by 13% under G1
7 relative to piControl (99 % cl), while the EWBs decrease in strength by 7% (99% cl). In
8 comparison, the strength of both the WWBs and EWBs is reduced (99% cl) under 4×CO₂, by
9 33% and 28%, respectively. The strong WWBs are more closely linked to positive SST
10 anomalies than negative SST anomalies (Cai et al., 2015a) and thus are likely to increase the
11 frequency of extreme El Niño events (Hu and Fedorov 2016) in G1, which is important with
12 regards to the mechanistic interpretation of the ENSO changes below.

13 **3.1.4 Zonal and meridional sea surface temperature gradients**

14 The ZSSTG between western and eastern equatorial Pacific is one of the characteristic
15 features of the equatorial tropical Pacific. The ZSSTG is weak during an El Niño and strong
16 during La Niña events (Latif et al., 2009). The ZSSTG is calculated as the difference between
17 SST in the western Pacific Warm Pool (5° N-5° S; 100° E-126° E) and eastern equatorial
18 Pacific (Niño3 region: 5° N-5° S; 160° E-150° W). The zonal SST gradient is reduced both in
19 4xCO₂ and G1 (Fig. 4b, 99% cl), but the reduction is smaller in G1 (11%) than in 4xCO₂
20 (62%). The reduced zonal SST asymmetry in 4×CO₂ and G1 is consistent with the weakening
21 of the trade winds and zonal wind stress, as noted in Sect. 3.1.3. The weakening of trade
22 winds can result in reduced upwelling in the eastern equatorial Pacific, and east to west
23 surface currents (Collins et al., 2010), leading to an increase in El Niño events. Our results
24 under 4xCO₂ are in agreement with Coats and Karnauskas (2017), who using several climate
25 models found a weakening of the ZSSTG under the RCP8.5 scenario.

26 MSSTG is calculated as the SST averaged over the off-equatorial region (5° N-10° N; 150°
27 W-90° W) minus SST averaged over the equatorial region (2.5° N-2.5° S; 150° W-90° W) (Cai
28 et al., 2014). Reversal of sign or weakening of the MSSTG has been observed during extreme
29 El Niño events, as the ITCZ moves over the equator (e.g., Cai et al., 2014). Overall there is a
30 change in sign and reduction of MSSTG in 4×CO₂ (~-111%, 99% cl) and only a slight
31 decrease in G1 (~-9%, 99% cl) (Fig. S3, and Table S2). The decrease in strength of MSSTG
32 is an indication that extreme El Niño events are expected to increase (Cai et al., 2014) under
33 solar geoengineering. The weakening of the MSSTG is qualitatively in agreement with
34 previous studies under increased GHG forcings (e.g., Cai et al., 2014; Wang et al., 2017).

35 **3.1.5 Thermocline**

36 Previous studies (e.g., Vecchi and Soden, 2007; Yeh et al., 2009) revealed shoaling as well as
37 a reduction in the east-west tilt of the equatorial Pacific thermocline under increased GHG
38 scenarios. A decrease in thermocline depth and slope is a dynamical response to reduced
39 zonal wind stress. Shoaling of the equatorial Pacific thermocline can result in positive SST

1 anomalies in the eastern tropical Pacific, which in turn can affect the formation of El Niño
2 (Collins et al., 2010).

3 Thermocline depth here is defined as the depth of the 20 °C (for piControl and G1), and 24 °C
4 (for 4×CO₂) isotherms averaged between 5° N and 5° S, following Phillip and van
5 Oldenborgh, (2006). Due to surface warming in GHG scenarios, the 20 °C isotherm deepens
6 (Yang and Wang et al., 2009), and this must be compensated by using a warmer isotherm (24
7 °C) as a metric in the 4×CO₂ case.

8 In 4xCO₂, the tropical Pacific thermocline depth (24 °C isotherm) shoals by 22% (99% cl,
9 Fig. 4c), as expected from similar experiments (Vecchi and Soden, 2007; Yeh et al., 2009).
10 However, there is no statistically significant change in the mean thermocline depth in G1. In
11 4xCO₂, most likely the weakened easterlies (as noticed in Sect. 3.1.3; e.g., Yeh et al., 2009,
12 Wang et al., 2017) and greater ocean temperature stratification due to increased surface
13 warming (see Sect. 4 and Cai et al., 2018) lead to a significant shoaling of the thermocline
14 across the western and central equatorial Pacific. In contrast, relatively little change takes
15 place between 130° W and 90° W. In a CMIP3 multimodel (SRESA1B scenario) ensemble,
16 Yeh et al. (2009) found a more profound deepening of the thermocline in this part of the
17 eastern equatorial Pacific; however, for example, Nowack et al. (2017) did not find such
18 changes under 4xCO₂ (cf. their Fig. S9). One possible explanation for this behaviour is the
19 competing effects of upper-ocean warming (which deepens the thermocline) and the
20 weakening of westerly zonal wind stress, causing thermocline shoaling (see Kim et al.
21 2011a).

22 **3.1.6 Vertical velocity and Walker circulation**

23 Under normal conditions, there is strong atmospheric upwelling over the western equatorial
24 Pacific, SPCZ, and ITCZ. In contrast, the relatively cold and dry eastern Pacific is dominated
25 by atmospheric downwelling. This process, as simulated in HadCM3L, can be seen in maps
26 of Omega500-100 (Fig. 5a). The region of ascent over the SPCZ and ITCZ moves
27 equatorward in 4×CO₂ (Fig 5b), consistent with the increase in SST and precipitation over the
28 equatorial region (Fig. 1d and 2d). The convective centre also moves towards the Niño3
29 region and centres at ~150° W. While these changes in spatial patterns of atmospheric
30 divergence and convergence are found to be corrected for G1 (Fig. 5c), significant
31 differences in the strength of the atmospheric circulation remain, which in turn are coupled to
32 the aforementioned changes in atmospheric stability. Specifically, both for 4×CO₂ and G1,
33 upwelling decreases over the Warm Pool, but increases in the central Pacific and the eastern
34 part of the Niño3 region (Fig. 5d-e). This picture is consistent with changes in the spatial
35 extent and a weakening of the tropical PWC (Fig. 6a-c). In 4xCO₂, the weakening and
36 shifting of circulation patterns are consistent with multimodel results reported by Bayr et al.
37 (2014) under GHG forcing. While mitigated, the PWC weakening found in G1 remains
38 highly statistically significant (99% cl; Fig. 6d-e).

39

40

1 **3.2 ENSO amplitude and frequency**

2 In Sect. 3.1, we described a variety of coupled, and highly significant changes in the tropical
3 Pacific mean state, such as the weakening of zonal and meridional SST gradients, zonal wind
4 stress, and PWC. It is well-known that such changes can affect ENSO variability. This
5 section discusses various metrics used to characterise ENSO variability and unfolds how they
6 change in 4xCO₂ and G1. Specifically, we investigate the amplitude of ENSO, changes in
7 amplitude asymmetry between El Niño and La Niña events, and ENSO frequency.

8 **3.2.1 ENSO amplitude**

9 To characterise changes in ENSO, this study uses two separate indices for two different
10 regions, because extreme warm and cold events are not mirror images of each other (Cai et
11 al., 2015b). The Niño3 (Niño4) index is employed for studying characteristics of El Niño (La
12 Niña) events in the eastern (central) Pacific region. ENSO amplitude is defined as the
13 standard deviation of SST anomalies in a given ENSO region (e.g., Philip and van
14 Oldenborgh 2006; Nowack et al., 2017). The maximum amplitude of warm events is defined
15 as the maximum positive ENSO anomaly during the entire time series analysed (Gabriel and
16 Robock, 2015). Cold events are defined similarly, but using the maximum negative ENSO
17 anomaly.

18 In 4×CO₂, both eastern and central Pacific ENSO amplitudes undergo a statistically
19 significant decrease (47 and 64%, respectively, at 99% cl, Table 1-2). The maximum
20 amplitude of warm events in the eastern Pacific and cold events in the central Pacific are also
21 significantly reduced (57% and 36% at 99% cl, respectively; Table 3-4). Previous studies
22 found that climate models produced mixed responses (both increases and decreases in
23 amplitude) in terms of how ENSO amplitude change with global warming (see Latif et al.,
24 2009; Collins et al., 2010; Vega-Westhoff and Srivier, 2017). However, Cai et al. (2018)
25 found an intermodel consensus, for models capable of simulating ENSO diversity, for
26 strengthening of ENSO amplitude under A2, RCP4.5, and RPC8.5 transient scenarios. In
27 contrast, in G1, the eastern Pacific ENSO amplitude gets strengthened (9% at 99% cl), and no
28 statistically significant change is noticed in the central Pacific ENSO amplitude.

29 Further, the maximum amplitude of cold events is strengthened in the central Pacific (20% at
30 99% cl), but no statistically significant change occurs in the eastern Pacific. A validation of
31 these changes in ENSO amplitude using the E- and C-indices, as these indices represent SST
32 anomalies similar to those of Niño3 and Niño4 index (Cai et al., 2015a), yields indeed very
33 identical results (see Table 1-4). Thus, our simulations imply that significant changes can
34 occur in ENSO events under solar geoengineering. Mechanistically, it is self-evident that
35 these changes might be linked to the tropical Pacific SST overcooling of ca. 0.30 °C and the
36 substantial SST gradient changes under G1 relative to piControl.

37 However, the use of standard deviations to define ENSO amplitude is suboptimal, because
38 amplitudes of El Niño and La Niña events are asymmetric, i.e., in general, El Niño events are
39 stronger than La Niña events (An and Jin, 2004; Schopf and Burgman, 2006; Ohba and Ueda,
40 2009; Ham, 2017). The relative strength of ENSO warm and cold events can be measured by

1 the skewness of SST over the ENSO regions (Vega-Westhoff and Sriver, 2017). Following
2 Ham (2017), we investigate the asymmetry in the amplitude of El Niño and La Niña events
3 by comparing the skewness of detrended Niño3 SST anomalies in piControl with 4×CO₂ and
4 G1.

5 We find that, relative to piControl, the Niño3 SST skewness is reduced both in 4×CO₂ (190%
6 at 99% cl) and G1 (65% at 99% cl) (Table 5). The E-Index also indicates reduced skewness
7 under both 4×CO₂ (85%) and G1 (28%) at 99% cl. The reduced skewness is further
8 illustrated in maps showing differences in skewness between 4×CO₂ and G1 with piControl
9 (Fig. S4). Over the eastern equatorial Pacific, the SSTs are transformed from positively to
10 negatively skewed under 4×CO₂ (Fig. S4b). Our results qualitatively agree with Ham (2017),
11 who found a 40% reduction in ENSO amplitude asymmetry using several CMIP5 models in
12 the RCP4.5 scenario. In G1 (Fig. S4e), the skewness of SSTs is reduced over the eastern
13 equatorial Pacific, whereas it strengthens over the central equatorial Pacific region (at 99%
14 cl). The strengthening of skewness over the central equatorial Pacific is also consistent with
15 increased C-Index skewness (66% at 99% cl) under G1 relative to piControl. Thus, due to the
16 concurrent strengthening of the maximum amplitude of cold events and reduction in the
17 asymmetry of SST skewness, the intensity of cold events is predicted to increase compared to
18 warm events under solar geoengineering.

19 **3.2.2 El Niño frequency**

20 We choose a threshold value of rainfall for defining extreme El Niño events based on the
21 work of Cai et al., (2014, 2017), who chose averaged DJF Niño3 total rainfall exceeding 5
22 mm day⁻¹ for this threshold based on observations. However, as pointed out by Cai et al.
23 (2017), trends in Niño3 rainfall are mainly driven by two factors: (1) the change in the mean
24 state of the tropical Pacific and (2) the change in frequency of extreme El Niño events.
25 Therefore, since we want to focus on the changes in the extremes, we need to remove
26 contribution (1) from the raw Niño3 time series. We, therefore, fit a quadratic polynomial to
27 the time series of rainfall data from which all extreme El Niño events (DJF total rainfall > 5
28 mm day⁻¹) have been excluded and then subtract this trend from the raw Niño3 rainfall time
29 series. Linearly detrending the rainfall time series produces similar results. Note that under
30 piControl (observations), total rainfall of 5 mm day⁻¹ is ~85th (~93rd) percentile in detrended
31 Niño3 rainfall time series. Wang et al. (2020) termed events with rainfall > 5 mm day⁻¹ as
32 extreme convective El Niño events.

33 With detrended Niño3 total rainfall exceeding 5 mm day⁻¹ as an extreme, three extreme and
34 seven moderate El Niño events can be identified from the historical record between 1979 and
35 2017 (Fig. 7a). A statistically significant increase of 526% (99% cl) in extreme El Niño
36 events can be seen under 4×CO₂ (939 events) relative to piControl (150 events) (Fig. 7b-c).
37 The geoengineering of climate (G1) largely offsets the increase in extreme El Niño frequency
38 under 4×CO₂ (Fig. 7d), however, compared to piControl, still a 17% increase in extremes and
39 a 12% increase in the total number of El Niño events (moderate plus extreme) can be seen at
40 95% cl. Thus, an El Niño event occurring every ~3.3-yr under preindustrial conditions occurs
41 every ~2.9-yr under solar geoengineered conditions.

1 A threshold of detrended Niño3 total rainfall of 5 mm day^{-1} recognises events as extremes
2 even when the MSSTG is positive and stronger, especially under $4\times\text{CO}_2$, which plausibly
3 means that ITCZ might not shift over the equator for strong convection to occur during such
4 extremes. The El Niño event of 2015 is a typical example of such events. We test our results
5 with a more strict criterion by choosing only those events as extremes, which have
6 characteristics similar to that of 1982 and 1997 El Niño events (i.e., Niño3 rainfall $> 5 \text{ mm}$
7 day^{-1} and MSSTG < 0). We declare events having characteristics similar to that of the 2015
8 event as moderate El Niño events (Fig. S5). Based on this method, we find a robust increase
9 in the number of extreme El Niño events both in $4\times\text{CO}_2$ (924%) and G1 (61%) at 99% cl. We
10 also performed the same analysis by linearly detrending the rainfall time series and find
11 similar results (Fig. S6).

12 An alternative approach to quantifying extreme El Niño events is based on Niño3 SST index
13 > 1.75 s.d. as an extreme event threshold (Cai et al., 2014). We note that using this definition,
14 no statistically significant change in the number of extreme El Niño events is detected in G1
15 (61 events), whereas they reduced from 57 in piControl to zero events in $4\times\text{CO}_2$ highlighting
16 the dependency of specific results on the precise definition of El Niño events used. However,
17 relative to piControl, Niño3 SST index indicates a statistically significant increase (decrease)
18 of 12% (46%) in the frequency of the total number of El Niño events (Niño3 SST index > 0.5
19 s.d.) (Table S3) in G1 ($4\times\text{CO}_2$). Further, we examine the change in extreme El Niño events
20 using E-Index > 1.5 s.d. (see Cai et al., 2018) as a threshold. The SST based E-Index
21 identifies 79, 147, and 93 extreme El Niño events in piControl, $4\times\text{CO}_2$, and G1, respectively.
22 Thus using E-Index, extreme El Niño events increase by 86% (99% cl) and 17% (missing
23 95% cl by three events) in $4\times\text{CO}_2$ and G1, respectively. Based on the E-index definition, we
24 see a statistically significant increase in the total number of El Niño events in $4\times\text{CO}_2$ (107%)
25 and no statistically significant change in G1 (Table S3). Note that Wang et al. (2020) showed
26 that extreme El Niño events having E-Index > 1.5 s.d. can still happen even if the Niño3
27 rainfall is not greater than 5 mm day^{-1} (cf. Figure 2 in Wang et al., 2020).

28 We highlight that both in $4\times\text{CO}_2$ and solar geoengineered climate, more weak and reversed
29 MSSTG events occur relative to piControl (Fig. S3). More frequent reversals of MSSTG
30 result in a more frequent establishment of strong convection in the eastern equatorial Pacific.
31 According to Cai et al. (2014), more frequent convection over the eastern tropical Pacific
32 increases the sensitivity of rainfall by 25% to positive SST anomalies. Further, in Sect. 3.1.3,
33 we found that WWBs (EWBs) are 13% (7%) stronger (weaker) than in piControl, which also
34 favours a higher frequency of El Niño events in G1. Thus, we conclude that changes in the
35 tropical Pacific mean state; in particular weakening of temperature gradients (MSSTG and
36 ZSSTG), changes in zonal wind stress, and convection over the tropical Pacific (and
37 consistent weakening of the PWC) are the plausible causes of increased frequency of extreme
38 El Niño events under G1.

39 3.2.3 La Niña frequency

40 During La Niña events, the ZSSTG, the PWC, and atmospheric convection in the western
41 Pacific are stronger than on average. Here, we present plots of Niño4 vs ZSSTG for

1 piControl, 4×CO₂, and G1 (Fig. 8a-c). In 4×CO₂, extreme La Nina events are reduced to zero
2 relative to piControl, and a statistically significant (99% cl) decrease occurs in moderate,
3 weak, and total number (sum of extreme, moderate and weak events) of La Niña events. Our
4 findings are inconsistent to those of Cai et al. (2015b) who found nearly doubling of extreme
5 La Nina events under increased GHG forcing. We see a statistically significant (95% cl)
6 increase in extreme La Niña events in G1. The number of extreme La Niña events increases
7 by 32% (61 events) in G1 relative to piControl (46 events). Thus, an extreme La Niña event
8 occurs every ~22 years in piControl and every ~16 years in G1.

9 The increased number of extreme El Niño events provides a possible mechanism for
10 increased frequency of La Niña events, as they result in more heat discharge events causing
11 cooling, hence providing conducive conditions for increased occurrence of La Niña events
12 (Cai et al., 2015a, 2015b). In addition, the ocean becomes 4% more stratified under G1
13 relative to piControl (Fig. 15e, Table S7). The increased vertical ocean stratification in the
14 central equatorial Pacific steers cooling in the Niño4 region and, hence, can cause more
15 frequent strong positive ZSSTG anomalies (Fig. S9c and S10b) resulting in an increased
16 number of extreme La Niña events (see also Cai et al., 2015b).

17 **3.3 Spatial characteristics of ENSO**

18 In Sect. 3.2, we showed that overall and maximum ENSO event amplitudes generally
19 strengthened under G1, while the amplitude asymmetry between warm and cold events is
20 significantly reduced. In this section, we present composite anomalies, i.e. the average
21 patterns of all El Niño and La Niña events. These composites provide process-based evidence
22 for the strengthening (weakening) of extreme La Niña (El Niño) events in G1. We show that
23 the PWC, SST, and composite rainfall anomalies are strengthened for extreme La Niña
24 events, while they are weakened for extreme El Niño events under G1. For composite
25 analysis, extreme El Niño events are selected with Niño3 rainfall > 5 mm day⁻¹ and MSSTG
26 < 0 (Fig. S5) because it gives a more robust estimate as all events show a reversal of MSSTG
27 and more vigorous convection.

28 **3.3.1 Weakening of extreme El Niño events in G1**

29 The broad spatial patterns of composite SST (Fig. 9), rainfall (Fig. 10), and PWC (Fig. 11)
30 anomalies for the extreme and total number of El Niño events in G1 are very similar to those
31 of piControl. During extreme El Niño events, in G1, we find reduced SST (Fig. 9e) and
32 rainfall anomalies (Fig. 10e) over the eastern and western equatorial Pacific with a consistent
33 weakening of the eastern and western branch of PWC (Fig. 11e). We also note reduced SST
34 (Fig. 9f) and rainfall (Fig. 10f) anomalies over the western Pacific in agreement with a
35 weakening of western branch of PWC (Fig. 11f) for the total number of El Niño events in G1.
36 Thus, in general, extreme El Niño events tend to be weaker in G1 than in piControl. We
37 conclude that, in our simulations, extreme El Niño events are more frequent but slightly less
38 intense in a solar geoengineered climate than in preindustrial conditions. We further confirm
39 this with a histogram of detrended Niño3 SST anomalies (Fig. S7a). Though more frequent
40 positive Niño3 SST anomalies occur under G1 (between 1 and 3 °C), the mean Niño3 SST

1 anomaly is weaker in G1 (1.95 °C) than in piControl (2.23 °C) at 99% cl. Thus, the strength
2 of extreme El Niño events is reduced by ~12% in G1 compared to piControl. However, no
3 statistically significant shift in histograms of Niño3 SST anomalies is detected for the total
4 number of El Niño events (Fig. S7b).

5 **3.3.2 Strengthening of La Niña events in G1**

6 The broad spatial patterns of composite SST (Fig 12a-d), rainfall (Fig. 13a-d) and PWC (14a-
7 d) anomalies for the extreme and total number of La Niña events are similar under G1 and
8 piControl. During the extreme and total number of La Niña events, the negative SST and
9 rainfall anomalies, and both east and west branch of PWC are strengthened indicating an
10 overall intensification of La Niña events in G1 relative piControl. We note that most of the
11 stronger negative SST anomalies occur over the eastern equatorial Pacific. We confirm
12 strengthening of La Niña events by plotting histograms of detrended Niño3 SST anomalies
13 for the extreme (piControl: -1.45 °C; G1: -1.68 °C) and the total number of La Niña events
14 (piControl: -1.03 °C; G1: -1.22 °C) based on the Niño4 SST index (Fig. S7c-d). Thus, we
15 conclude that the strength of extreme (total number of) La Niña events is increased by ~16%
16 (~18%) in G1 compared to piControl.

17 **4 Mechanisms behind the changes in ENSO variability**

18 **4.1 Under greenhouse gas forcing**

19 The reduced ENSO amplitude under 4×CO₂ is mainly caused by stronger hf and weaker BJ
20 feedback relative to piControl (Fig. 15a-b, and Table S5-6). More rapid warming over the
21 eastern than western equatorial Pacific regions reduces the SST asymmetry between western
22 and eastern Pacific (Fig. 1d), resulting in the weakening of ZSSTG (Fig. 4b) that significantly
23 weakens the zonal winds stress (Fig. 4a) and hence PWC (Fig. 6b, d, see Bayr et al., 2014).
24 The overall reduction of zonal wind stress reduces the BJ feedback, which, in turn, can
25 weaken the ENSO amplitude. Climate models show an inverse relationship between hf
26 feedback and ENSO amplitude (Lloyd et al., 2009, 2011; Kim and Jin, 2011b). The increased
27 hf feedback might be the result of enhanced clouds due to strengthened convection (Fig. 5b,
28 d) and stronger evaporative cooling in response to enhanced SSTs under 4×CO₂ (Knutson
29 and Manabe, 1994; Kim and Jin, 2011b). Kim and Jin (2011a, b) found intermodel consensus
30 on the strengthening of hf feedback in CMIP3 models under enhanced GHG warming
31 scenario (Ferret and Collins, 2019). Further, we see increased ocean stratification under
32 4×CO₂ (Fig. 15d and Table S7). A more stratified ocean is associated with an increase in both
33 the El Niño events and amplitude in the eastern Pacific (Wang et al., 2020). It can also
34 modify the balance between feedback processes (Dewitte et al., 2013). Enhanced
35 stratification may also cause negative temperature anomalies in the central to the western
36 Pacific through changes in thermocline tilt (Dewitte et al., 2013). Since the overall ENSO
37 amplitude decreases in our 4xCO₂ simulation, we, thus, conclude that the ocean stratification
38 mechanisms cannot be the dominant factor here, but that hf and BJ feedbacks must more than
39 cancel out the effect of ocean stratification on ENSO amplitude. Bjerknes feedback is a
40 multi-component process (e.g., Kim and Jin, 2011a), where some components may increase

1 and some may decrease under the influence of external forcing. For instance, increased upper
2 ocean stratification tends to enhance the Bjerknes feedback, likely through coupling between
3 the wind and thermocline. However, this study represents the Bjerknes feedback solely on the
4 coupling between wind and SST, a caveat of this analysis.

5 The increased frequency of extreme El Niño events under $4\times\text{CO}_2$ is due to change in the
6 mean position of the ITCZ (Fig. S2), causing frequent reversals of MSSTG (Fig. S3), and
7 eastward extension of the western branch of PWC (Fig. 6), which both result in increased
8 rainfall over the eastern Pacific (see Wang et al., 2020). This is due to greater east equatorial
9 than off-equatorial Pacific warming (see Cai et al., 2020), which shifts the mean position of
10 ITCZ towards the equator (Fig. S2). Simultaneously more rapid warming of the eastern than
11 western equatorial Pacific reduces the ZSSTG, and hence zonal wind stress, as also evident
12 from the weakening and shift of the PWC (Fig. 6) and increased instances of negative ZSSTG
13 anomalies (Fig. S9). Ultimately, this leads to more frequent vigorous convection over the
14 Niño3 region (Fig. 5d), and enhanced rainfall (Fig. 2d, S8). Therefore, despite the weakening
15 of the ENSO amplitude under $4\times\text{CO}_2$, rapid warming of the eastern equatorial Pacific causes
16 frequent reversals of meridional and zonal SST gradients, resulting in an increased frequency
17 of extreme El Niño events (see also Cai et al., 2014; Wang et al., 2020).

18 We note that under GHG forcing, HadCM3L does not simulate an increase in the frequency
19 of extreme La Niña events as found by Cai et al. (2015b) using CMIP5 models. However, it
20 does show an increase in the total number of La Niña events (Table S4). In a multimodel
21 ensemble mean, Cai et al. (2015b) found that the western Pacific warms more rapidly than
22 the central Pacific under increased GHG forcing, resulting in strengthening of the zonal SST
23 gradient between these two regions. Strengthening of this zonal SST gradient and increased
24 vertical upper ocean stratification provide conducive conditions for increased frequency of
25 extreme La Niña events (Cai et al., 2015b). One reason why we do not see an increase in the
26 frequency of central Pacific extreme La Niña events might be that HadCM3L does not
27 simulate more rapid warming of the western Pacific compared to the central Pacific as
28 noticed by Cai et al. (2015b) (compare our Fig. 1d with Fig. 3b in Cai et al., 2015b), hence, as
29 stronger zonal SST gradient does not develop, across the equatorial Pacific, as needed for
30 extreme La Niña events to occur (see Fig. S9a, c and S10).

31 **4.2 Under solar geoengineering**

32 G1 over cools the upper ocean layers, whereas the GHG-induced warming in the lower ocean
33 layers is not entirely offset, thus increasing ocean stratification (Fig. 15). The increased
34 stratification boosts atmosphere-ocean coupling (see Cai et al., 2018), which favours
35 enhanced westerly wind bursts (Fig. 4a) (e.g., Capotondi et al., 2018) to generate stronger
36 SST anomalies over the eastern Pacific (Wang et al., 2020). The larger cooling of the western
37 Pacific than the eastern Pacific can also enhance westerly wind bursts reinforcing the BJ
38 feedback and hence SST anomalies in the eastern Pacific. We conclude that increased ocean
39 stratification, along with stronger BJ feedback, is the most likely mechanism behind the
40 overall strengthening of ENSO amplitude under G1.

1 The increased frequency of extreme El Niño events under G1 can be linked to the changes in
2 MSSTG and ZSSTG (see Cai et al., 2014, and Fig. S3, S9). The eastern off-equatorial Pacific
3 cools more than the eastern equatorial regions, providing relatively more conducive
4 conditions for convection to occur through a shift of ITCZ over to the Niño3 region (Fig. 1e).
5 At the same time, the larger cooling of the western equatorial Pacific than of the eastern
6 equatorial Pacific reduces the ZSSTG and convective activity over the western Pacific, which
7 leads to a weakening of the western branch of PWC (Fig. 6e). Hence we see reduced rainfall
8 over the western Pacific and enhanced rainfall from the Niño3 to the central Pacific region
9 (Fig 2e). These mean state changes, strengthening of convection between $\sim 140^\circ$ W and $\sim 150^\circ$
10 E, and more reversals of the MSSTG and ZSSTG (Fig. S3) result in an increased number of
11 extreme El Niño events in G1 than in piControl (Fig. 7).

12 **5 Discussion and conclusions**

13 In this paper, we have analysed the impact of abruptly increased GHG forcing ($4\times\text{CO}_2$), and
14 solar geoengineering (G1), on the tropical Pacific mean climate and ENSO extremes.
15 Previous solar geoengineering studies did not show any statistically significant change in the
16 PWC (e.g., Guo et al., 2018) or ENSO frequency and amplitude (e.g., Gabriel and Robock
17 2015). However, those results were strongly limited by the length of the respective
18 simulations, which made changes challenging to detect, given the high tropical Pacific
19 climate variability. This limitation has been overcome here by using long (1000-year) climate
20 model simulations, carried out with HadCM3L. The longer record makes it possible to detect
21 even relatively small changes between the preindustrial and G1 scenarios within the chosen
22 model system.

23 To conclude, solar geoengineering can compensate many of the GHG-induced changes in the
24 tropical Pacific, but, importantly, not all of them. In particular, controlling the downward
25 shortwave flux cannot correct one of the climate system's most dominant modes of
26 variability, i.e., ENSO, wholly back to preindustrial conditions. The ENSO feedbacks
27 (Bjerkness and heat flux) and more stratified ocean temperatures may induce ENSO to
28 behave differently under G1 than under piControl and $4\times\text{CO}_2$. Different meridional
29 distributions of shortwave and longwave forcings (e.g., Nowack et al., 2016) resulting in the
30 surface ocean overcooling, and residual warming of the deep ocean are the plausible reasons
31 for the solar geoengineered climate not reverting entirely to the preindustrial state.

32 The changes in ENSO feedbacks and more stratified ocean temperatures under both $4\times\text{CO}_2$
33 and G1 can also affect the eastern and central Pacific ENSO variability differently. For
34 instance, more stratified ocean and enhanced BJ feedback in G1 strengthens the eastern
35 Pacific ENSO amplitude but not central Pacific ENSO amplitude (Table 1-2). Similarly, the
36 enhanced hf and weaker BJ feedback in $4\times\text{CO}_2$ results in a more substantial reduction in
37 central Pacific ENSO amplitude than eastern Pacific ENSO amplitude (Table 1-2). In the
38 current model system, we expect that changes in tropical Pacific mean state and feedback
39 process, both under $4\times\text{CO}_2$ and G1, may impact the occurrence ratio of central Pacific El
40 Niño (La Niña) to eastern Pacific El Niño (La Niña) (e.g., Yeh et al., 2009), which requires
41 further detailed analysis.

1 Finally, we note that this is a single model study, and more studies are needed to show the
2 robustness and model-dependence of any results discussed here, e.g. using long-term
3 multimodel ensembles from GeoMIP6 (Kravitz et al., 2015), once the data are released. The
4 long-term Stratospheric Aerosol Geoengineering Large Ensemble (GLENS; Tilmes et al.,
5 2018) data can also be explored to investigate ENSO variability under geoengineering.

6 We summarise our key findings as follows:

- 7 1. The warming over the tropical Pacific under increased GHG forcing ($4\times\text{CO}_2$) is
8 overcompensated under solar sunshade geoengineering (G1), resulting, by design, in
9 tropical mean overcooling of approximately $0.3\text{ }^\circ\text{C}$. This overcooling is more
10 pronounced in the western tropical Pacific and SPCZ than in the eastern Pacific under
11 the G1 scenario.
- 12 2. The reduced SST and rainfall asymmetry between the warm pool and the cold tongue,
13 seen under $4\times\text{CO}_2$, is mostly corrected in G1, but regionally important differences
14 remain relative to preindustrial conditions. The tropical Pacific is 5% wetter in
15 $4\times\text{CO}_2$, whereas it is 5% drier in G1 relative to piControl. In particular, solar
16 geoengineering results in decreased rainfall over the warm pool, SPCZ, and ITCZ and
17 increased rainfall over the central and eastern equatorial Pacific.
- 18 3. The preindustrial median position of ITCZ ($154^\circ\text{ W}-82^\circ\text{ W}; 7.5^\circ\text{ N}$) changes
19 significantly under $4\times\text{CO}_2$ and moves over the equator ($154^\circ\text{ W}-82^\circ\text{ W}; 0^\circ$). G1
20 restores the ITCZ to its preindustrial position ($154^\circ\text{ W}-82^\circ\text{ W}; 7.5^\circ\text{ N}$).
- 21 4. The increased GHG forcing results in 31% reduction in zonal wind stress over the
22 tropical Pacific. G1 fails to compensate this reduction entirely and results in
23 weakening the zonal wind stress by 10% with a 13% (7%) increase (decrease) in
24 WWBs (EWBs), thus providing more conducive conditions for El Niño extremes.
- 25 5. Under solar geoengineering, both ZSSTG and MSSTG are reduced by 11% and 9%,
26 respectively. More frequent reversal of MSSTG occurs in G1 relative to piControl.
- 27 6. In $4\times\text{CO}_2$, the thermocline flattens over the tropical Pacific, and G1 recovers its
28 preindustrial condition.
- 29 7. The PWC becomes weaker both under $4\times\text{CO}_2$ and G1 scenarios.
- 30 8. The increased GHG forcing results in a weakening of ENSO amplitude, whereas solar
31 geoengineering strengthens it relative to preindustrial climate. The maximum
32 amplitude of cold events is enhanced under G1.
- 33 9. The reduced ENSO amplitude under $4\times\text{CO}_2$ is mainly due to enhanced hf feedback,
34 whereas the increase under G1 is mainly caused by enhanced BJ feedback and ocean
35 stratification.
- 36 10. The ENSO amplitude asymmetry between warm and cold events is reduced under G1
37 relative to piControl.
- 38 11. The frequency of extreme El Niño events increases by 61% in G1 relative to
39 piControl. Further, the frequency of the total number of El Niño events also increases
40 by 12%. Thus, an El Niño event occurring every $\sim 3.3\text{-yr}$ under preindustrial
41 conditions occurs every $\sim 2.9\text{-yr}$ under solar geoengineered climate. The reason for the

1 occurrence of more extreme El Niño events under G1 is more frequent reversals of
2 MSSTG compared to piControl.

3 12. The frequency of extreme La Niña events increases by 32% under G1 relative to
4 piControl. Thus, an extreme La Niña event occurring every ~22-yr in piControl
5 occurs every ~16-yr in G1.

6 **Author contribution.** Long Cao developed the model code and performed the simulations.
7 Abdul Malik formulated the research questions, defined the methodology with the help of all
8 co-authors, and performed the scientific analysis. Abdul Malik prepared the manuscript with
9 contributions from all co-authors.

10 **Competing interests.** The authors declare that they have no conflict of interest.

11 **Data availability.** Data are available upon request from Long Cao (longcao@zju.edu.cn).
12

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20 <https://cds.climate.copernicus.eu/cdsapp#!/home>.
21

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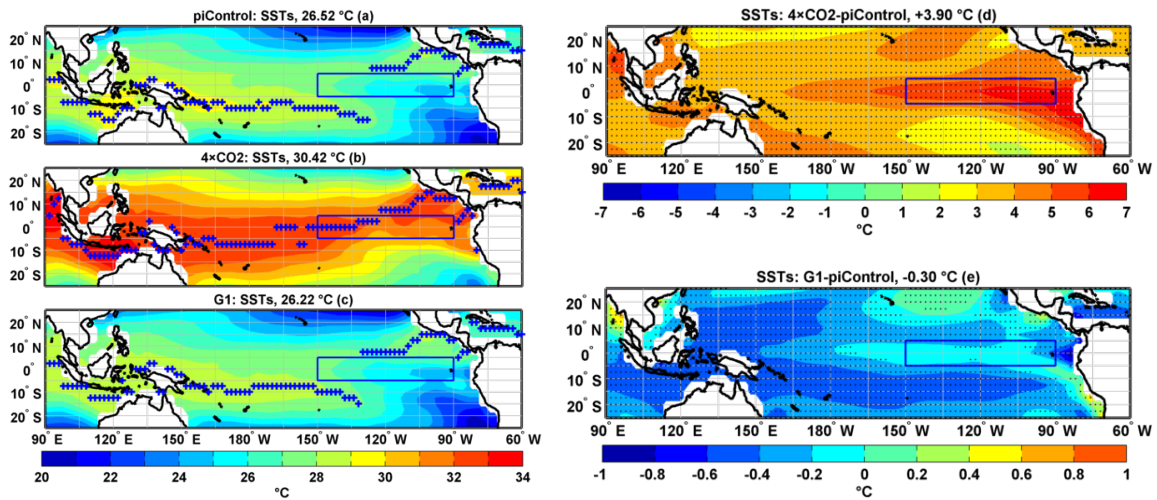
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1 **Figures and Figure Captions**

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4 **Figure 1.** Tropical Pacific SST mean DJF climatology (a) piControl (b) $4\times\text{CO}_2$ (c) G1 (d)
 5 difference $4\times\text{CO}_2$ -piControl and (e) difference G1-piControl. The blue plus sign in a-c
 6 indicates latitudes with maximum SSTs. Stipples indicate grid points where the difference is
 7 statistically significant at 99% cl using a non-parametric Wilcoxon rank-sum test. The box in
 8 the eastern Pacific identifies the Niño3 region. The numbers in a-c represent a mean
 9 temperature in the corresponding simulation, and numbers in d-e represent an area-averaged
 10 difference of piControl with $4\times\text{CO}_2$ and G1, respectively, in the tropical Pacific region (25°
 11 N- 25° S; 90° E- 60° W).

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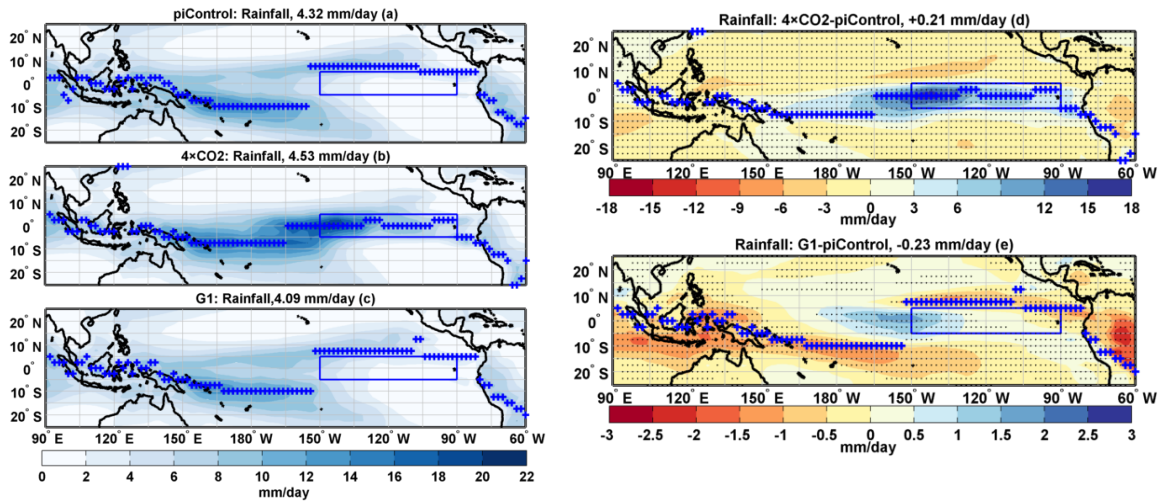
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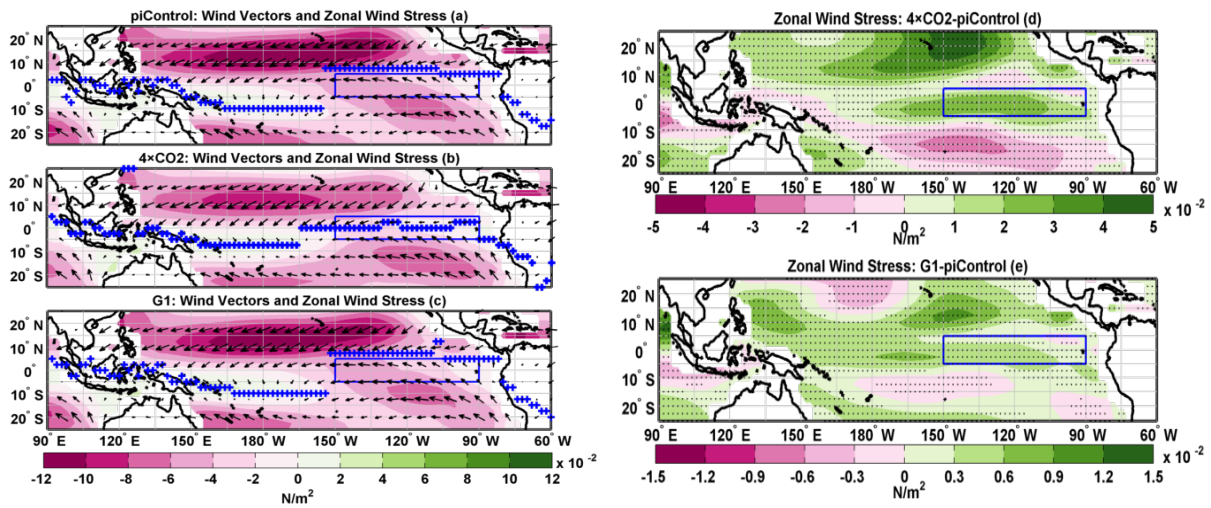
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Figure 2. Tropical Pacific rainfall mean DJF climatology (a) piControl (b) $4\times\text{CO}_2$ (c) G1 (d) difference: $4\times\text{CO}_2$ -piControl; the blue plus signs indicate the position of ITCZ under $4\times\text{CO}_2$ and (e) difference: G1-piControl; the blue plus signs indicate the position of ITCZ under G1. In a-c, the blue plus signs indicate the position of ITCZ for the corresponding experiment. Stipples indicate grid points where the difference is statistically significant at 99% cl using a non-parametric Wilcoxon rank-sum test. The numbers in a-c represent mean rainfall in the corresponding simulation, and numbers in d-e represent an area-averaged difference of piControl with $4\times\text{CO}_2$ and G1, respectively, in the tropical Pacific region (25°N - 25°S ; 90°E - 60°W).



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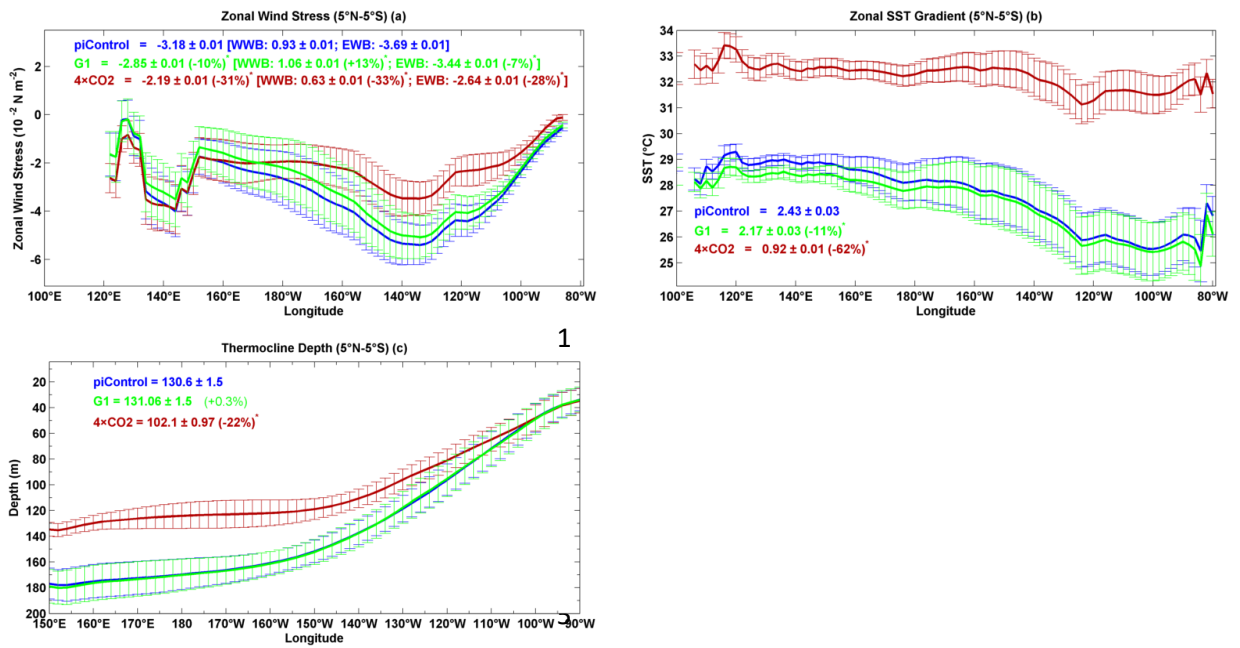
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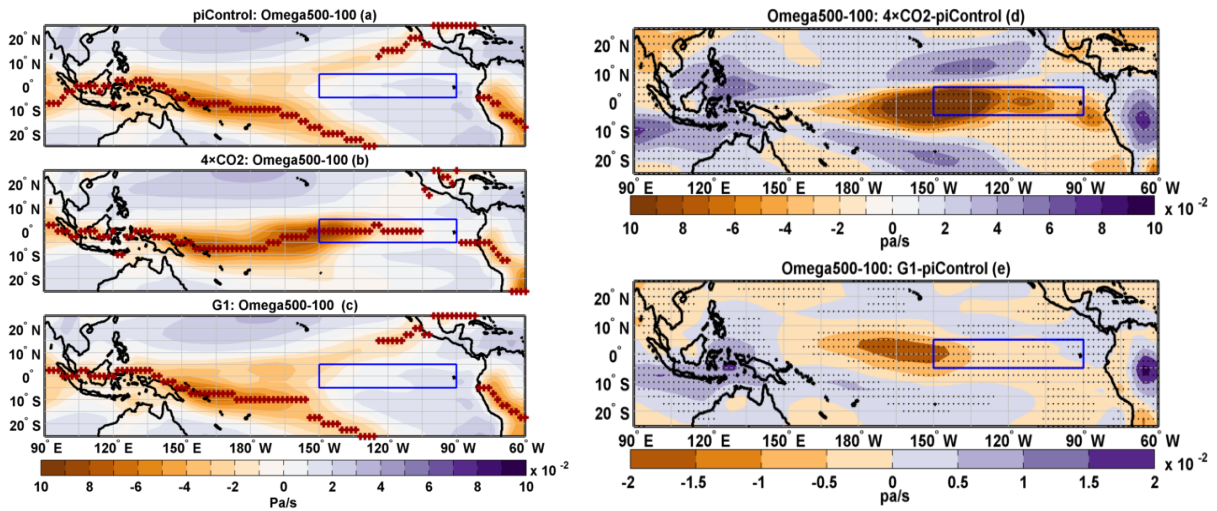
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Figure 3. Tropical Pacific zonal wind stress mean DJF climatology (a) piControl (b) $4\times\text{CO}_2$ (c) G1 (d) difference: $4\times\text{CO}_2$ -piControl and (e) difference: G1-piControl. Black arrows indicate the direction of 10 m wind. The blue plus sign in a-c indicates latitudes with maximum rainfall. Stipples indicate grid points where the difference is statistically significant at 99% cl using a non-parametric Wilcoxon rank-sum test.



6 **Figure 4.** DJF mean climatology of (a) zonal wind stress, (b) zonal SST gradient, and (c)
 7 thermocline depth. Error bars indicate ± 1 s.d. calculated over the simulated period. Numbers
 8 with an asterisk indicate that the percentage change is statistically significant at 99% cl.

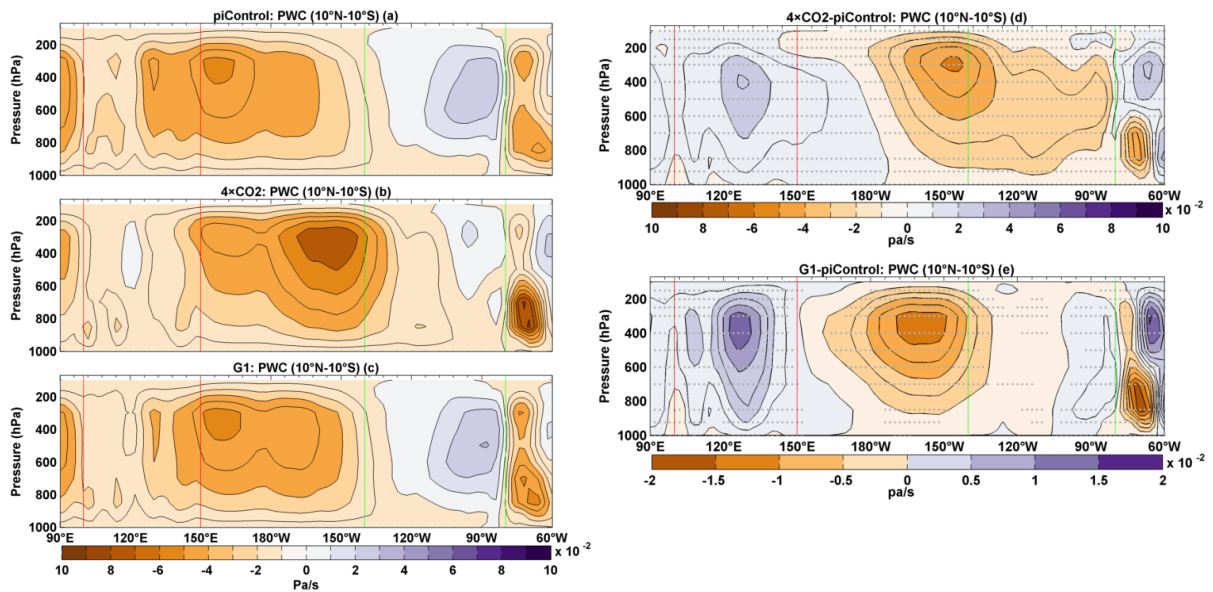
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11 **Figure 5.** Tropical Pacific mean DJF climatology of vertical velocity averaged between 500-
 12 and 100-hPa (Omega500-100) (a) piControl (b) 4xCO₂ (c) G1 (d) difference: 4xCO₂-
 13 piControl and (e) difference: G1-piControl. In a-c, the brown plus sign indicates latitudes
 14 where maximum upwelling occurs. Stipples indicate grid points where the difference is
 15 statistically significant at 99% cl using a non-parametric Wilcoxon rank-sum test.

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2 **Figure 6.** Mean DJF climatology of tropical Pacific Walker Circulation averaged over 90° E-
 3 60° W and 10° N-10° S (a) piControl (b) 4×CO₂ (c) G1 (d) difference: 4×CO₂-piControl and
 4 (e) difference: G1-piControl. Green (red) vertical lines show the longitudinal spread of the
 5 eastern (western) Pacific. Stipples indicate grid points where the difference is statistically
 6 significant at 99% cl using a non-parametric Wilcoxon rank-sum test.

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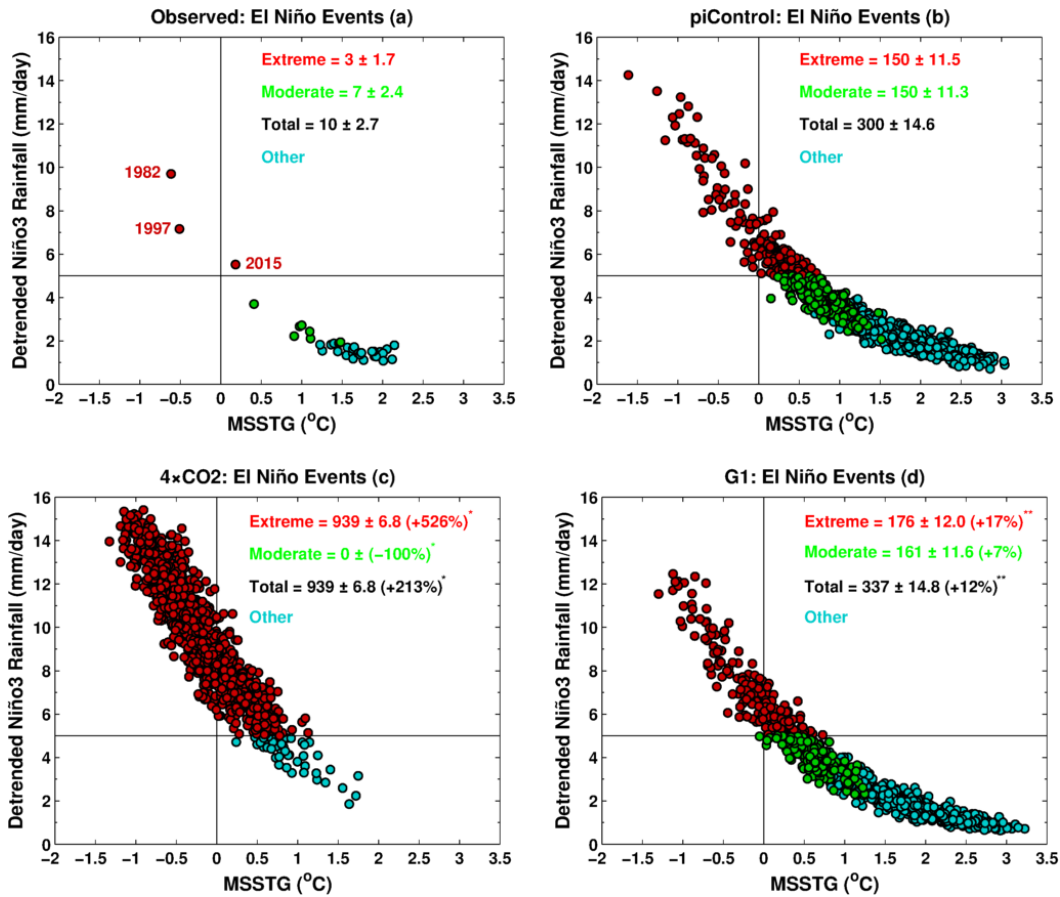
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3 **Figure 7.** Relationship between MSSTG and Niño3 rainfall for (a) observations (b) piControl
 4 (c) 4×CO₂, and (d) G1. A solid black horizontal line indicates a threshold value of 5 mm day⁻¹.
 5 See text for the definition of extreme, moderate, and total El Niño events. A single (double)
 6 asterisk indicates that the change in frequency, relative to piControl, is statistically significant
 7 at 99% (95%) c.l. Numbers with a ± symbol indicate s.d. calculated with 10,000 bootstrap
 8 realisations. Following Cai et al. (2014), a non-ENSO related trend has been removed from
 9 the rainfall time series.

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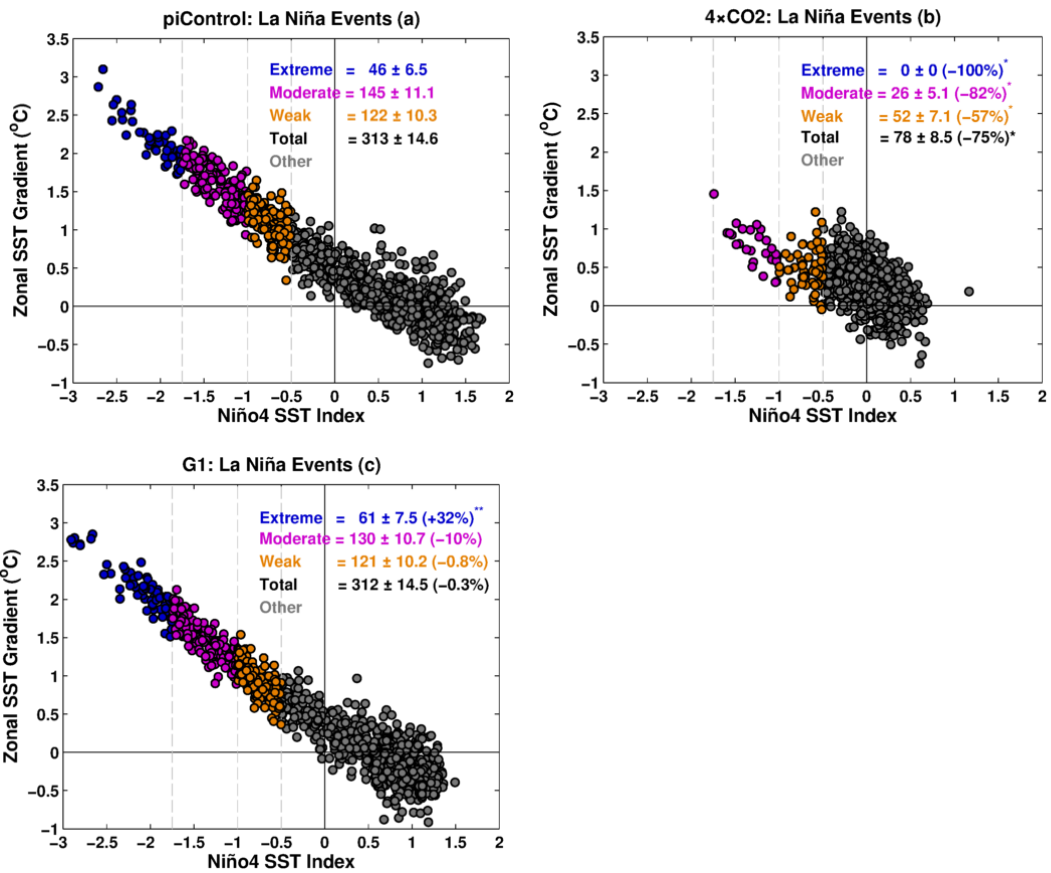
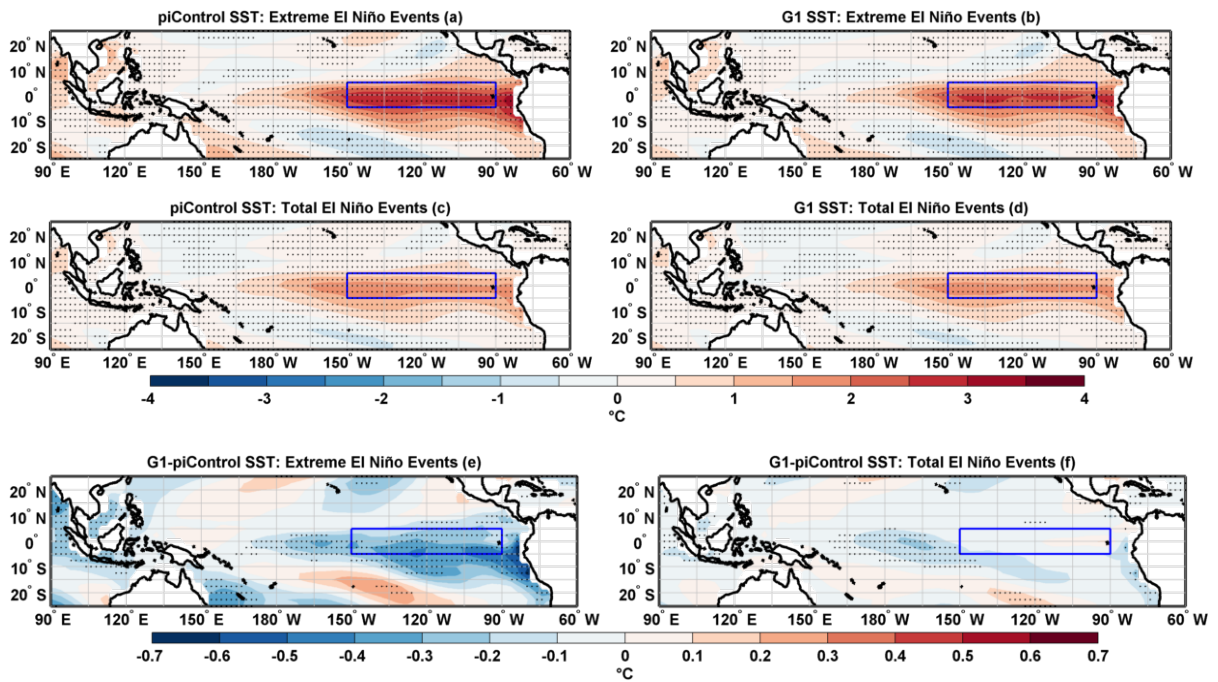


Figure 8. Relationship between ZSSTG and Niño4 SST index for (a) piControl (b) 4xCO₂ and (c) G1. Dashed grey vertical lines indicate threshold values of -1.75, -1, and -0.5 s.d. See text for the definition of extreme, moderate, weak, and total La Niña events. A single (double) asterisk indicates that the change in frequency is statistically significant at 99% (95%) cl. Numbers with a ± symbol indicate s.d. calculated with 10,000 bootstrap realisations.



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3 **Figure 9.** Composites of SST anomalies for extreme El Niño events in (a) piControl and (b)
 4 G1. Composites of SST anomalies for the total number of El Niño events in (c) piControl and
 5 (d) G1. Composite differences (G1-piControl) of SST anomalies for (e) extreme El Niño
 6 events and (f) total number of El Niño events. Stipples indicate grid points with statistical
 7 significance at 99% cl using a non-parametric Wilcoxon rank-sum test. The blue box in the
 8 eastern Pacific identifies the Niño3 region.

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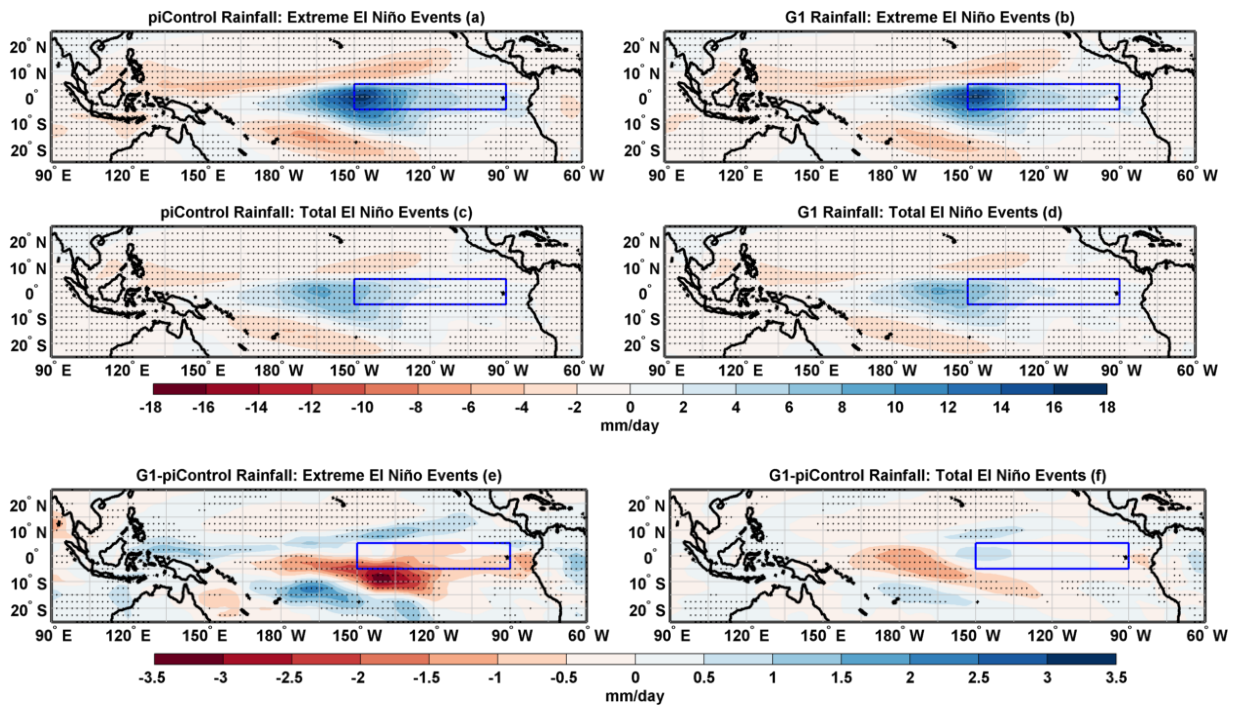
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3 **Figure 10.** Composites of rainfall anomalies for extreme El Niño events in (a) piControl and
 4 (b) G1. Composites of rainfall anomalies for the total number of El Niño events in (c)
 5 piControl and (d) G1. Composite differences (G1-piControl) of rainfall anomalies for (e)
 6 extreme El Niño events and (f) total number of El Niño events. Stipples in a-d and f (e)
 7 indicate grid points with statistical significance at 99 (95) % cl using a non-parametric
 8 Wilcoxon rank-sum test. The blue box in the eastern Pacific identifies the Niño3 region.

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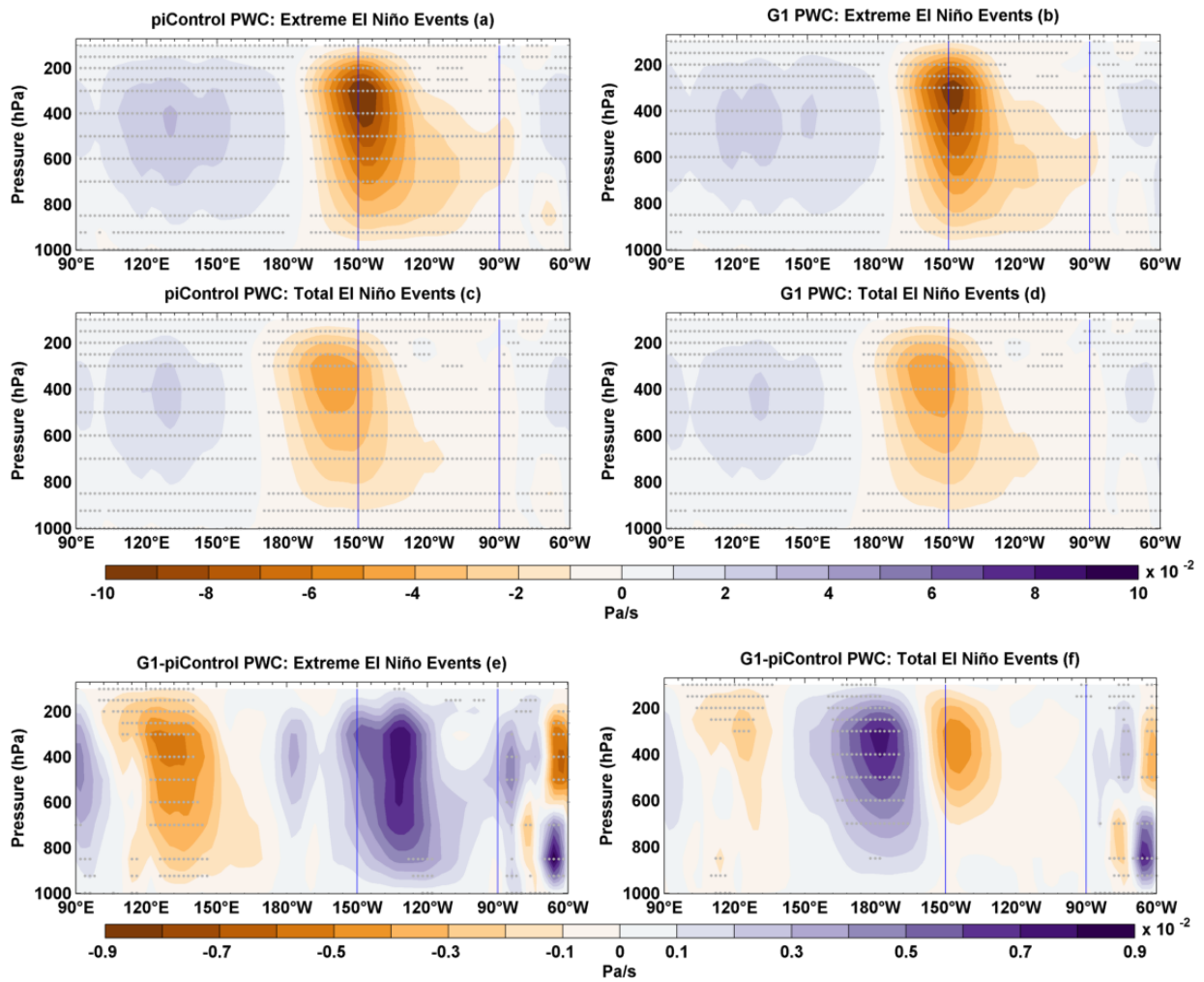
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3 **Figure 11.** Composites of PWC anomalies for extreme El Niño events in (a) piControl and
 4 (b) G1. Composites of PWC anomalies for the total number of El Niño events in (c)
 5 piControl and (d) G1. Composite differences (G1-piControl) of PWC for (e) extreme El Niño
 6 events and (f) total number of El Niño events. Stipples indicate grid points with statistical
 7 significance at 99% CI using a non-parametric Wilcoxon rank-sum test. The blue vertical lines
 8 indicate the Niño3 region.

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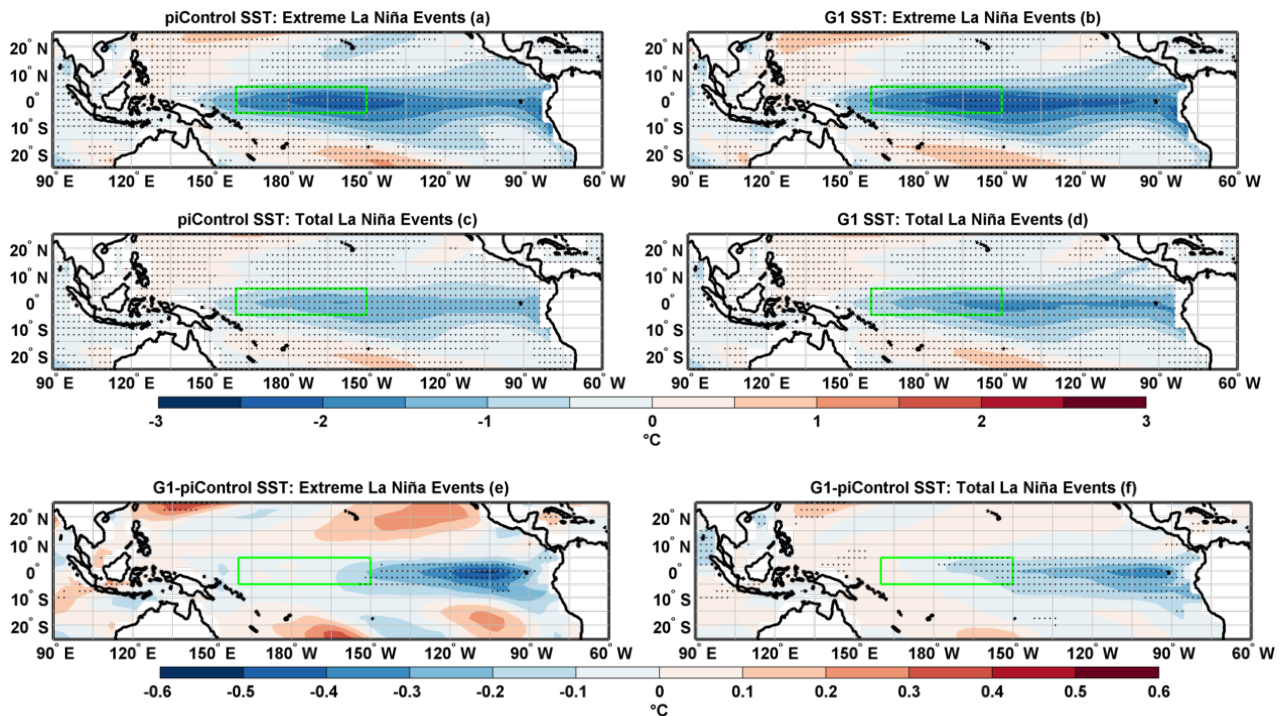
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3 **Figure 12.** Composites of SST anomalies for extreme La Niña events in (a) piControl and (b)
 4 G1. Composites of SST for the total number of La Niña events in (c) piControl and (d) G1.
 5 Composite differences (G1-piControl) of SST for (e) extreme La Niña events and (f) the total
 6 number of La Niña events. Stipples indicate grid points with statistical significance at 99% cl
 7 using a non-parametric Wilcoxon rank-sum test. The green box indicates the Niño4 region.

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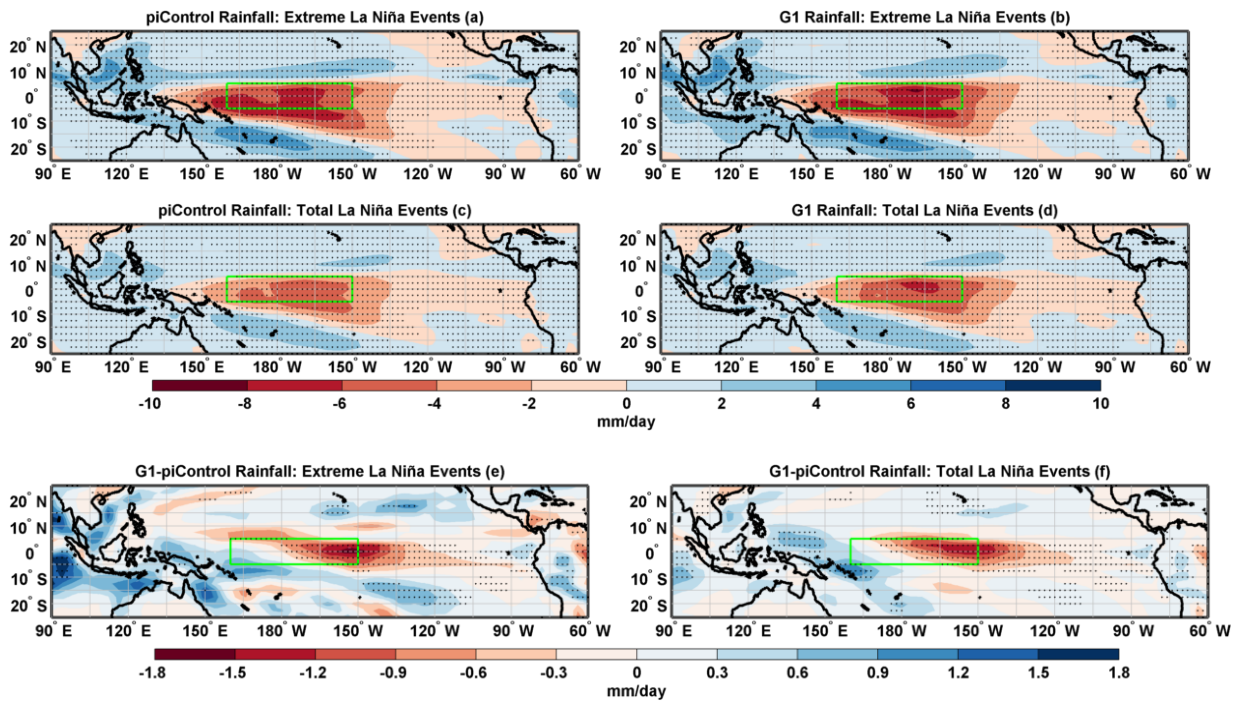
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3 **Figure 13.** Composites of rainfall anomalies for extreme La Niña events in (a) piControl and
 4 (b) G1. Composites of rainfall anomalies for the total number of La Niña events in (c)
 5 piControl and (d) G1. Composite differences (G1-piControl) of rainfall for (e) extreme La
 6 Niña events and (f) the total number of La Niña events. Stipples indicate grid points with
 7 statistical significance at 99% cl using a non-parametric Wilcoxon rank-sum test. The green
 8 box indicates the Niño4 region.

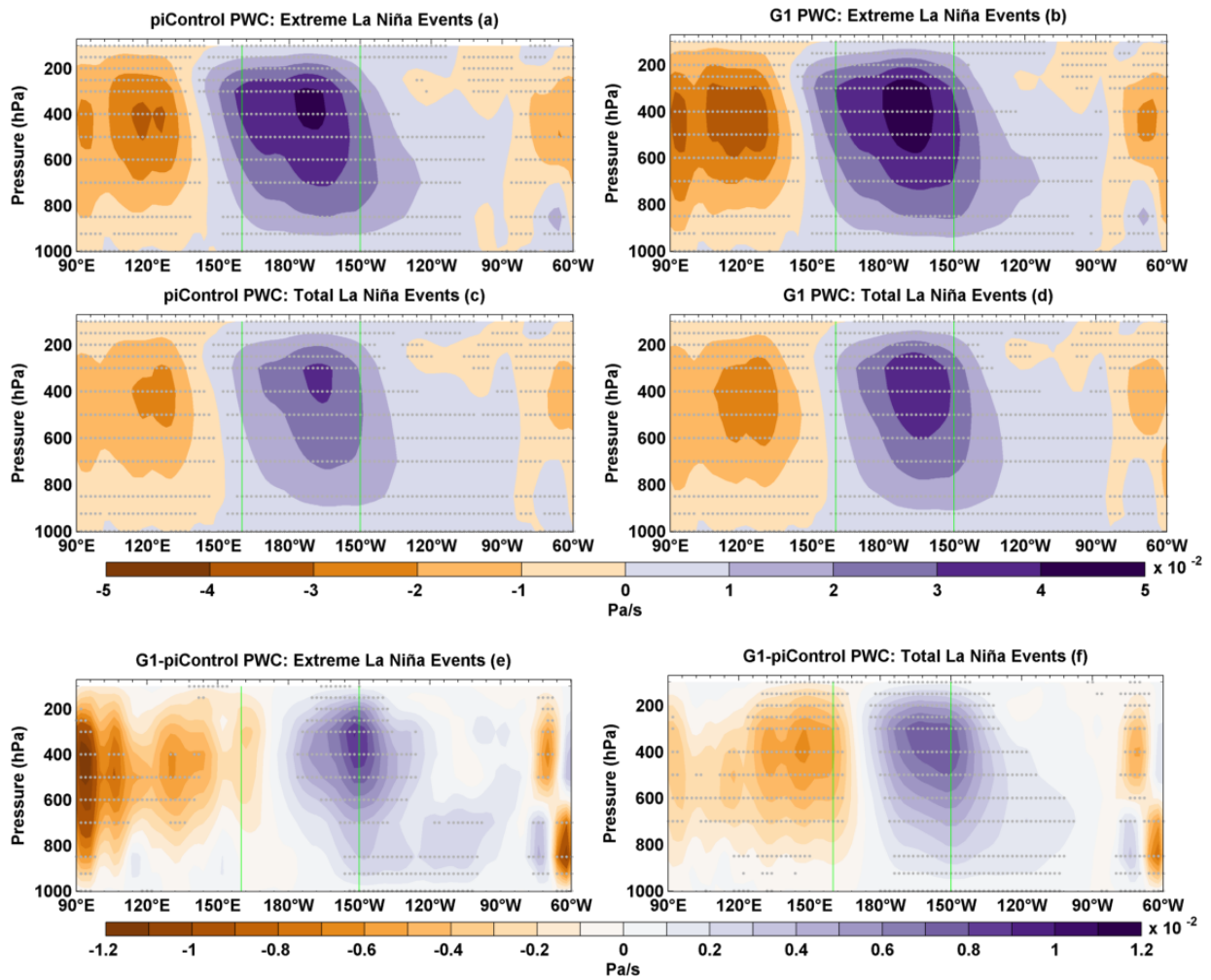
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3 **Figure 14.** Composites of PWC anomalies for extreme La Niña events in (a) piControl and
 4 (b) G1. Composites of PWC for the total number of La Niña events in (c) piControl and (d)
 5 G1. Composite differences (G1-piControl) of PWC anomalies for (e) extreme La Niña events
 6 and (f) the total number of La Niña events. Stipples indicate grid points with statistical
 7 significance at 99% cl using a non-parametric Wilcoxon rank-sum test. The green vertical
 8 lines indicate the Niño4 region.

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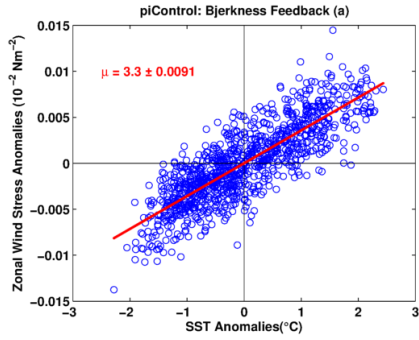
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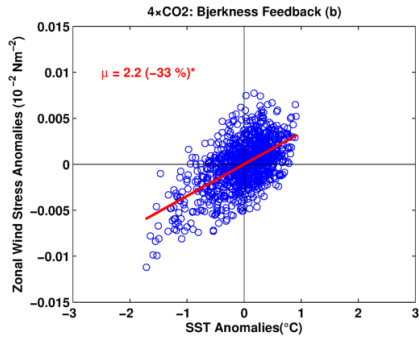
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16

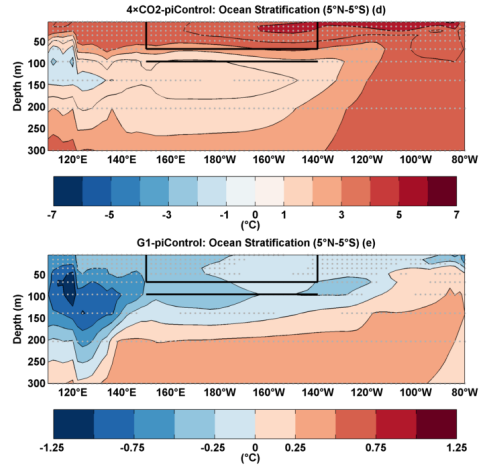
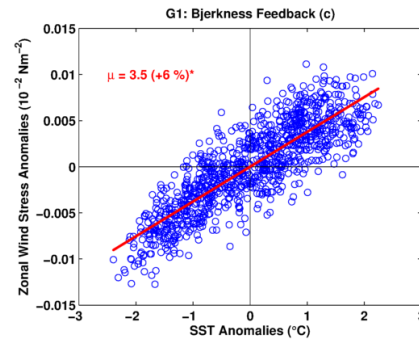
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4 **Figure 15.** BJ feedback (μ ; $10^{-2} \text{ Nm}^{-2}/^{\circ}\text{C}$) for (a) piControl (b) $4\times\text{CO}_2$, and (c) G1. The value
 5 with \pm sign indicates s.d. of μ after 10,000 bootstrap realisations. An asterisk indicates
 6 statistical significance at 99% cl. Mean change in ocean temperature, (d) $4\times\text{CO}_2$ -piControl,
 7 and (e) G1-piControl. The black box shows the area averaging region for upper ocean
 8 temperature, and the black line shows the lower layer used for calculation of stratification as
 9 a difference of upper and lower layer. Stipples indicate grid points with statistical
 10 significance at 99% cl using a non-parametric Wilcoxon rank-sum test.

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1 **Tables and Table Captions**

2 **Table 1.** Eastern Pacific ENSO amplitude

Experiment	Amplitude (°C)	Difference w.r.t. piControl (°C)	Std. Dev. 10,000 Realizations (°C)	~ Change w.r.t. piControl (%)
piControl	1.04 [1.03]		0.0213 [0.03]	
4×CO ₂	0.55 [0.85]	-0.49 [-0.18]		-47* [-17*]
G1	1.13 [1.13]	0.09 [0.1]		+9* [+10**]

3 Key: Niño3 [E-Index]; *99% cl; **95% cl

4

5 **Table 2.** Central Pacific ENSO amplitude

Experiment	Amplitude (°C)	Difference w.r.t. piControl (°C)	Std. Dev. 10,000 Realizations (°C)	~ Change w.r.t. piControl (%)
piControl	(0.78) [0.85]		(0.0132) [0.0167]	
4×CO ₂	(0.28) [0.53]	(-0.50) [-0.32]		(-64*) [-38*]
G1	(0.79) [0.83]	(0.01) [0.03]		(+1) [-3]

6 Key: (Niño4) [C-Index]; *99% cl; **95% cl

7

8 **Table 3.** Maximum amplitude of warm events

Experiment	Amplitude (°C)	Difference w.r.t. piControl (°C)	Std. Dev. 10,000 Realizations (°C)	~ Change w.r.t. piControl (%)
piControl	2.97 [4.59]		0.0687 [0.2342]	
4×CO ₂	1.29 [3.65]	-1.68 [-0.94]		-57* [-21*]
G1	2.85 [4.33]	-0.12 [-0.26]		-4 [-6]

9 Key: Niño3 [E-Index]; *99% cl; **95% cl

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11 **Table 4.** Maximum amplitude of cold events

Experiment	Amplitude (°C)	Difference w.r.t. piControl (°C)	Std. Dev. 10,000 Realizations (°C)	~ Change w.r.t. piControl (%)
piControl	(-2.13) [-2.47]		(0.0459) [0.1452]	
4×CO ₂	(-1.37) [-2.17]	(-0.76) [-0.30]		(-36*) [-12*]
G1	(-2.55) [-2.90]	(0.42) [0.43]		(+20*) [+17*]

12 Key: (Niño4) [C-Index]; *99% cl; **95% cl

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14 **Table 5.** Niño3 SST skewness

Experiment	Skewness	Difference w.r.t. piControl	Std. Dev. 10,000 Realizations	~ Change w.r.t. piControl (%)
piControl	0.52*		0.0542	
4×CO ₂	-0.47*	-0.99		-190*
G1	0.18*	-0.34		-65*

15 Key: *99% cl; **95% cl