Eastern equatorial Pacific warming delayed by aerosols and thermostat response to CO_2 increase

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Abstract

Understanding the tropical Pacific response to global warming remains challenging. Here, we use a range of CMIP6 greenhouse warming experiments to assess the recent and future evolution of the equatorial Pacific east-west temperature gradient and corresponding Walker circulation. In abrupt CO_2-increase scenarios many models generate an initial strengthening of this gradient resembling an ocean thermostat (OT), followed by a small weakening; other models generate an immediate weakening that becomes progressively stronger establishing a pronounced eastern equatorial Pacific (EP) warming pattern. The initial response in these experiments is a strong predictor for the future EP pattern simulated in both abrupt and realistic warming scenarios, but not in historical simulations showing no multi-model trend. The likely explanation is that the recent CO_2-driven changes in the tropical Pacific are masked by aerosol effects and a potential OT-related delay, while the EP warming pattern will emerge as greenhouse gases overcome aerosol forcing.

Main

The emerging warming patterns in the tropical Pacific are key to understanding the climate response to increasing greenhouse gases [1–4]. This is true on interannual timescales, as changes in the El Niño/Southern Oscillation (ENSO) are linked to changes in the tropical Pacific mean state [5–7], and on decadal and longer timescales, as Pacific trade wind slow variations and ENSO decadal modulation can explain temporary global warming trend slowdowns [3,8–11] with links to climate sensitivity [12,13]. At these different timescales, the tropical Pacific warming affects atmospheric tropical circulation, rainfall[14], and mid-latitude teleconnections[15,16]. Therefore, it is imperative to understand and predict changes in the tropical Pacific (or more generally the Indo-Pacific), yet progress is hindered by discrepancies between models and observations[17], diverging theories[1], and uncertainty in projections[18].

Recent studies demonstrate that Pacific surface winds, and sea surface temperature (SST) and sea level pressure (SLP) gradients along the equator, have increased over the past three to four decades comprising the satellite era [14,19–21]. Additionally, the Indian Ocean has been warming at a faster rate than the Pacific since the 1950s [22,23]. Whether these trends are driven by anthropogenic climate change remains under debate.

Clement et al.[24] proposed a mechanism by which the Pacific SST gradient may increase in response to atmospheric warming, as the surface water in the western Pacific warms faster than the upwelling-cooled water in the east. This effect, the 'ocean thermostat' (OT), has been confirmed as a transient response to radiative forcing in box models [25] and ocean general circulation models (GCMs) [26,27]. In GCMs, the OT typically involves a central rather than far-eastern equatorial Pacific cooling (or suppressed warming), impacting the Indo-Pacific temperature contrast, and the effect reverses after decades to a century [27]. The importance of inter-basin warming contrasts in driving this Pacific cooling[11] is also documented in experiments imposing SST changes in the Indian and Atlantic oceans [9,23,28,29]. In addition, the wind-evaporation-SST (WES) feedback initiated off the equator in the Pacific trade wind belts may further contribute to the OT effect and lack of central and eastern equatorial Pacific warming[30].

While observations show a strengthening Pacific east-west SST gradient in recent decades, the majority of CMIP5 GCMs predict a weakening in future forcing scenarios [17,18], associated with an equatorial eastern Pacific (EP) warming pattern and weakening Walker circulation. Several mechanisms have been proposed to explain this SST gradient weakening, including enhanced evaporative damping over the warm pool [31], the slowdown of convective overturning driven by a slower precipitation rate increase (controlled by radiation) relative to low-troposphere water vapor increase (controlled by the Clausius-Clapeyron relation) [32,33], low cloud feedbacks[34] and the warming of extra-tropical regions and/or a slowdown of oceanic subtropical cells (STCs) that provide source water for equatorial upwelling [27,35]. However, the robustness of the projected zonal gradient weakening has been questioned because CMIP5 models fail to capture observed tropical Pacific trends [17,36] and exhibit a large spread in future projections [18]. Furthermore, the models show SST and wind biases along the equator [37,38] and deficiencies in simulated ENSO [39,40].

We address these outstanding challenges using CMIP6 models with new forcing scenarios, enabling a better assessment of historical and future trends and offering new insights into inter-model discrepancies in the tropical warming patterns with implications for future changes.

Evolution of the tropical Pacific response to global warming

We investigate the tropical Pacific and Indian Ocean warming patterns across CMIP6 experiments forced by different concentrations of greenhouse gases (GHGs) and aerosols. To capture observed and simulated trends [23,27,41], we define the equatorial Pacific east-west SST gradient as the surface temperature over the Indo-Pacific warm pool region minus that over the central-east Pacific (Methods). Changes in this gradient are coupled across models to Pacific Walker circulation changes (Extended Data Fig. 1), so strengthening or weakening of the temperature gradient implies similar strengthening or weakening of the Walker circulation.

All but one model show a long-term zonal SST gradient weakening in projections and abrupt CO_2-increase experiments (Fig. 1b-e, Fig. 2, Extended Data Fig. 2). However, in historical experiments with full radiative forcing, models show no significant trend on average (Fig. 1a) yet exhibit pronounced inter-model spread (Fig. 1a and Fig. 2b). Therefore, the projected long-term trends in the tropical Pacific appear disconnected from historical simulation changes. In addition to natural variability, this delayed SST gradient response could be caused by two factors. The first is the competition between the OT effect and the opposing atmospheric and oceanic mechanisms weakening the Pacific zonal temperature gradient in response to CO_2-

forcing. The second is aerosol emissions, which can influence tropical Pacific warming patterns[42,43] and delay the expected east-west temperature gradient weakening. We explore these separately, first examining hypothetical CO_2-only scenarios, then analyzing realistic full-forcing historical and future Shared Socioeconomic Pathway (SSP) scenarios with GHG and aerosol forcing, and finally analyzing a subset of experiments that isolate aerosol and GHG effects.

OT and EP warming pattern in CO_2-forced experiments

The abrupt 4xCO_2-increase experiments show small transient strengthening of the equatorial Pacific temperature gradient lasting about 10 years, up to 30 years in some models (Fig. 1d). The gradual 1pct CO_2 experiments show no model-mean change for the first 50 years (Fig. 1e), even though some models have a transient strengthening lasting ~100 years. In both scenarios, the initial response is followed by zonal temperature gradient weakening. Thus, depending on how abruptly the system is perturbed by CO_2, and the model in question, the long-term gradient weakening can be delayed by transient strengthening or a flat initial trend.

To understand these initial (transient) changes and model spread, we separate models into two categories based on their initial response during the first 25 years in the abrupt CO_2-forcing experiments. The OT category contains 7 models showing a strong initial increase of the temperature gradient (above 0.25 K), and the EP category contains 10 models exhibiting immediate weakening of the gradient (below -0.25 K), see Supplementary Table 1 and Fig. 2a. The rest of the models fall between these end members.

The initial response of the OT category exhibits negative temperature anomalies (cooling or suppressed warming) in the central equatorial Pacific that extend to the subtropical south-eastern Pacific (Fig. 3a) and an anomalous pressure gradient between the Indian and Pacific Oceans (Fig. 4a). The lack of eastern/central Pacific warming can be explained by upwelling of cold water that balances the CO_2-induced radiative forcing, while the western equatorial Pacific and the Indian ocean, along with the Maritime continent, warm faster. This establishes an anomalous Indo-Pacific pressure gradient, strengthening the easterly winds and causing a transient cooling [26,27] in the central and parts of the eastern equatorial Pacific (Fig. 3a). The WES feedback [27,44,45] further strengthens the trade winds south of the equator (Fig. 4a), contributing to the lack of warming here[30,46].

The initial surface warming patterns are different in the EP category, which shows a broad warming from the eastern to central Pacific (Fig. 3c) with associated low-pressure anomalies (Fig. 4c). Such changes are related to weakening of the Walker cell caused by atmospheric mechanisms discussed earlier, amplified by low-cloud feedbacks [30,34,47] that reduce the eastern Pacific low cloud cover, warming the ocean surface and further weakening the Walker circulation (Fig. 4c). A similar warming pattern also emerges in slab-ocean simulations without active ocean dynamics [27,32].

The two model categories eventually converge to a warming pattern showing enhanced eastern equatorial warming [30,48], with sharp meridional contrast especially in the southern hemisphere

(Fig. 3b and 3d). The transition from cooling to warming (in OT models) or continuous warming (in EP models) in the eastern and central equatorial Pacific can be explained by the gradual warming of the upper ocean, which reduces the OT effect. This allows other competing mechanisms, like decreased convective mass flux over the warm pool [32], stronger evaporative damping in the west[31], and on longer timescales enhanced extra-tropical warming/ STC slowdown [27,49], to become dominant, leading to a Walker circulation slowdown. In the EP category, these mechanisms, amplified by cloud feedbacks, allow for a greater Walker circulation slowdown and broader equatorial warming compared to the OT models.

The structure, magnitude and time of emergence of the Pacific warming pattern in the idealized CO_2-forcing experiments are therefore set by competing OT and EP warming effects [27,48]. The balance appears to be controlled by differences between model mean states – on average, the OT models are colder, including the warm pool but excluding the Southeast Pacific, but have weaker easterly winds in the equatorial band (Extended Data Fig. 3). A colder warm pool may reduce evaporative damping and Walker circulation slowdown, decreasing the EP response**,** while weaker mean winds maintain a shallower thermocline in the central equatorial Pacific, likely facilitating the OT effect. Compared to the OT category, the EP category has a smaller tropical SST bias relative to observations but higher SLP and equatorial zonal wind stress biases (Extended Data Figs. 4 and 5).

While most models exhibit transient temperature gradient strengthening in the first decades of the abrupt CO_2-increase experiments (Fig. 1a), only several show a strong change, exceeding

+0.25 K and lasting up to 25 years (Figs. 2a and 5a). These models have a particularly weak long-term response. By contrast, models that do not display a transient OT effect have stronger long-term EP warming. Overall, initial and long-term changes in the temperature gradient in the abrupt 4xCO_2 experiments are highly correlated (Fig. 5a). Furthermore, long-term changes in the gradual 1pct-CO_2 simulations are also highly correlated with the initial changes in the abrupt experiments (Fig. 5b). In the 1pct-CO_2 experiments, the OT models also show the longest delay (up to 80 years) before the zonal temperature gradient starts weakening.

Thus, despite some inter-model spread, the initial response in the abrupt CO_2 forcing experiments is a good predictor for the strength of the long-term EP warming across idealized CO_2 scenarios, the GHG-only forced historical simulations and the long-term trends in SSP experiments.

Historical versus future projections

Similar to the 1pct CO_2-only experiments, realistically forced SSPs show high correlation between the long-term response of the zonal temperature gradient and the initial response in 4xCO_2 abrupt experiments (Fig. 5e,f). This is true for SSP5-8.5, in which aerosol emissions decline rapidly, and SSP3-7.0, which maintains aerosol emissions close to 2000-2010 levels [46]. Consequently, in the long-term tropical Pacific response, CO_2 and other GHGs dominate aerosols, and again the initial response in abrupt 4xCO_2 experiments is a good predictor for long-term changes.

In contrast, we find almost no correlation between the initial response in abrupt-forcing experiments and historical simulations (Fig. 5d). In principle, this could result from relatively small historical GHG changes, compared to 4xCO_2 or 1pct CO_2, and hence the dominant role of natural variability [51]. However, historical experiments with at least 5 models (CESM2-FV2, HadGEM3-GC31-LL, NESM3, MPI-ESM-1-2-HAM, TaiESM1) simulate significant mean zonal SST gradient strengthening in recent decades (Fig. 2b), which contradicts the initial response in abrupt CO_2 experiment in those same models (Fig. 2). Furthermore, for 14 models with GHG-only historical runs available, the correlation between the simulated historical trends and the initial response in the 4xCO_2 experiments is strong (0.72 versus 0.25 for the full historical forcing, Fig. 5c). That is, without aerosols, these models would have predicted trends for the past 60 years – from flat to a substantial gradient weakening – consistent with their initial response in the 4xCO_2 experiments. This suggests aerosol effects can temporarily strengthen the SST gradient or delay its weakening, which, together with natural variability, may diminish the correlation between historical simulations and abrupt CO_2-only experiments.

Opposing effects of aerosol and GHG

We investigate the role of aerosols using a subset of 12 models for which historical GHG-only and aerosol-only experiments are available. For these models, we compare mean surface temperature and pressure pattern changes between the full-forcing historical simulations and these hypothetical partial-forcing experiments, focusing on changes since the 1950s for which reliable radiative forcing and ocean temperature measurements are available.

The GHG-only historical simulations generate a model-mean ocean warming similar to the EP warming in abrupt 4xCO_2 experiments, with equatorial warming in the eastern and central Pacific and anomalous wind convergence towards a negative SLP anomaly off the South American west coast (Fig. 6b). The full-forcing historical simulations show enhanced warming over land, stronger at mid-to-high latitudes and weaker over South Asia (Fig. 6a). However, the patterns of change over the ocean are more muted (Fig. 6a). Likewise, while the SLP and wind anomalies show pronounced changes in the extra-tropical Pacific and Indian Oceans (the latter effect is likely related to aerosol emissions over Asia), the anomalies are small in the equatorial Pacific.

In contrast, the aerosol-only historical simulations show a model-mean equatorial cooling signal (Fig. 6c, Extended Data Fig. 6c) similar to the EP pattern but with opposite sign. Unlike the OT-type response, this cooling is not accompanied by low pressure anomalies over the Indian ocean. Instead, low pressure anomalies over the southern Pacific and western Atlantic and weak positive anomalies develop along the equator cause anomalous equatorial divergence of winds, especially in the southern hemisphere.

In the full historical simulations, the aerosol and GHG forcings produce opposite model-mean trends in the tropical Pacific and largely cancel, generating a muted warming pattern. Nevertheless, individual models display large differences in the tropics. These include differences in the overall temperature sensitivity to aerosol and GHG forcings, the extent to which aerosols induce patterned versus uniform cooling, and whether the responses to aerosols and GHG forcing are linearly additive or involve nonlinear interactions. We analyze three illustrative models — HadGEM3-GC31-LL, MIROC6 and CNRM-CM6 — in more depth (Extended Data Fig. 7). HadGEM3-GC31-LL exhibits a stark contrast between the full-forcing historical experiment, showing central equatorial Pacific cooling, and the GHG-only historical experiment showing EP warming (Extended Data Fig. 8). This contrast is probably driven by nonlinear interactions between the aerosol and GHG-driven effects, since a linear superposition of the two simulations does not produce equatorial cooling.

Several other models, notably CESM2-FV2, TaiESM1 and UKESM1, generate a strong SST gradient strengthening in the historical simulations, but do not show an OT-type response in the abrupt CO_2scenario (Fig. 2). Similar to HadGEM3-GC31-LL, these trends might be driven by aerosol forcing and nonlinear aerosol-GHG interactions, rather than the OT mechanism.

Two other models, MIROC6 and CNRM-CM6, show similar warming patterns between the historical full-forcing and GHG-only simulations, and are less sensitive to aerosol forcing than HadGEM3-GC31-LL (Extended Data Fig. 8). Yet, CNRM-CM6 develops a modest equatorial cooling in both the GHG-only and full-forcing scenarios (Extended Data Fig. 9), while MIROC6 has a slight EP warming trend in the full-forcing simulation, and a stronger EP warming trend in the GHG-only simulation (Extended Data Fig. 10). Thus, CNRM-CM6 is representative of the models with zonal temperature gradient strengthening or no trend in the historical simulations, and hence a delay of EP warming, caused likely by an OT-type response with aerosol effects superimposed. MIROC6 may represent models that show a slight weakening of the temperature gradient in the full historical experiments since they do not have strong aerosol or OT-type effects, and thus generate historical trends opposite to observations.

Discussion

We have investigated the tropical Pacific response to radiative forcing across a range of CMIP6 warming simulations. In abrupt 4xCO_2 experiments, we distinguish two types of behavior based on initial decadal changes: (1) models with a relatively strong OT-like initial response followed by a weak and delayed EP warming and (2) models in which EP warming starts to develop within the first decade. Eventually, most CO_2-onfly experiments develop EP warming, but of vastly different magnitudes and somewhat different spatial structures, and the initial response in abrupt 4xCO_2 experiments is a good predictor for the eventual EP pattern strength. Accordingly, the initial OT-like response is typically replaced in the long-term by a weak EP pattern, while the initial, relatively weak EP pattern develops into a strong EP pattern. Across the models, the correlation between the initial temperature gradient changes in the 4xCO_2 experiments and longer-term changes in CO_2-rise experiments varies from 0.72 to 0.86, confirming the utility of the abrupt 4xCO_2 experiments in elucidating the physics of the response and highlighting the OT and EP mechanisms.

Despite the strong correlation between the initial and long-term responses across CO_2-only experiments, we find poor correlation between initial changes in the 4xCO_2 experiments and the realistic (full-forcing) historical simulations. Isolating the effects of aerosols and GHGs shows that aerosols may explain this poor correlation by countering the GHG warming pattern and thus suppressing or delaying the EP warming response in the full-forcing experiments. A strong OT effect in some of the models may also contribute to this delay.

In realistic future scenarios (SSP3-7.0, SSP5-8.5), even with continuing aerosol emissions, the equatorial EP warming becomes more pronounced and it evolves to correlate well with the $4xCO_2$ initial response, as the GHG-driven tropical response overcomes aerosol effects. Thus, our results suggest that any anthropogenic component of the currently observed trends in the tropical Pacific may be of transient nature and caused by either aerosol effects or an OT response to GHG-forcing, or a combination of both, acting to delay the EP warming. Consequently, as future atmospheric GHGs increase and aerosol emissions decrease, an enhanced warming of the eastern equatorial Pacific and associated Walker circulation weakening can be expected (Fig. 5, Extended Data Fig. 1 and Supplementary Fig. 1).

Several major questions remain. First, the large spread in long-term tropical Pacific projections remains an issue, caused largely by differences in how models respond to CO_2 rather than aerosol effects or natural variability. The competition among various mechanisms that either strengthen or weaken the Pacific temperature gradient, including the transient OT effect, plays out differently in each model, and how these differences are related to model resolution, biases and sub-grid parameterizations remains a challenging question to address. A generally colder tropical Pacific seems to favor the OT effect, but with large variations across the models. Second, the models differ strongly in their representation of tropical natural variability, and future changes in modes of climate variability may contribute to inter-model spread, which particularly concerns model representation of ENSO[52] and its change.

Third, few models capture the observed historical trend, whether aerosol-driven or not, and differences in the controlling mechanisms and in projections are large. If aerosols are driving the observed lack of warming in the equatorial central and eastern Pacific, one may expect a strong reversal once aerosol emissions level off and GHG effects dominate. However, if the observed trends in the Pacific are driven primarily by an OT-type response, perhaps in combination with multi-decadal climate variability, we may expect a delayed and weaker EP warming pattern. The uncertainty in what drives the observed trend thus makes it difficult to obtain more robust future projections, warranting more scrutiny of the role of aerosols in these models and constraints on the transient ocean thermostat.

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Methods

Experiments

We analyze three types of experiments from the Climate Model Intercomparison Project Phase 6 (CMIP6) archive. The first type consists of two hypothetical CO_2-only experiments: "abrupt-4xCO_2" rise and "1pctCO_2" gradual CO_2 rise (1% per year) where CO_2-increases are relative to a pre-industrial level of 280 ppm (here quotation marks refer to labelling of experiments in the CMIP6 archive). The second type has full-forcing experiments: "historical" simulations using the observed greenhouse gas (GHG) concentrations and aerosol emissions, as well as two future scenarios, "SSP5-8.5" and "SSP3-7.0" which follow projected changes in GHG and aerosols. The third type of experiments consists of modified historical runs with either historical GHG concentrations prescribed and no aerosols ("hist-GHG"), or prescribed historical aerosol emissions with a constant preindustrial GHG level ("hist-aer"). Note that for UKESM1 and MPI-ESM-1-2-HAM, we used their "hist-piaer" (historical with pre-industrial aerosol emission) simulation rather than the "hist-GHG" simulation as the latter was not available. The "hist-piaer" simulation, however, is similar to "hist-GHG" for our purposes since it keeps aerosol emissions constant at the pre-industrial level.

Models and observational datasets

We use a total of 40 CMIP6 models in our analysis. The criterion for including a given model is whether it has at least one ensemble member and 150 years of simulation available for the following experiments: “abrupt-4xCO_2”, “1pctCO_2”, “piControl”, and “historical” (at the time the data was downloaded during a period from March to September 2020). Some models do not have future projected scenarios (Fig. 2b), and these models are excluded from that part of the analysis. For the CO_2-only simulations, given the strong forcing signal, we only use one ensemble member per model, but for the historical simulation and future projections, for which the forcing is smaller and therefore natural variability plays a greater role, we include all available ensemble members, whose numbers range from 1 to 32 (Supplementary Table 1). For Fig 1, however, we use only one ensemble member per model to give equal weight to each model. In addition, a subset of 12 models is used for the second part of the analysis to separate the effects of greenhouse gases and aerosols by comparing the GHG-only, aerosols-only and full-forcing historical simulations. The selection of these 12 models is based on the criterion of having at least one ensemble member available for each of the three types of historical simulations – realistic and partial-forcing. Lastly, for SST data we use Extended Reconstructed Sea Surface Temperature, version 4 (ERSST v4) [54], and for winds and sea level pressure we use NCEP/NCAR Reanalysis 1[55].

Metrics

We define the Pacific zonal temperature gradient as the surface temperature difference between two regions along the equator – the region of the Indo-Pacific warm pool encompassing the western equatorial Pacific, the Maritime continent and the eastern Indian ocean (80ºE to 150ºE, 5ºS to 5ºN, including some land areas) minus the central and eastern equatorial Pacific (180ºE to 280ºE, 5ºS to 5ºN), see Supplementary Fig. 1a. We will refer to this gradient as the Pacific zonal temperature or SST gradient since the inclusion of land has only a minor effect, and since this gradient is dominated by the Pacific east-west SST contrast. The strength of the Walker circulation is defined as a sea level pressure (SLP) difference between the same regions used for defining the zonal temperature gradient, but with a minus sign. Anomalies are calculated relative to a preindustrial Control ("piControl") time mean of at least 150 years for the hypothetical CO_2-only experiments and relative to a historical baseline of 1950-1970 for the historical and SSP scenarios. We chose this particular baseline because historical observations become more reliable after the 1950s.

Data availability

CMIP6 data is available at: https://esgf-node.llnl.gov/search/cmip6/. Individual dataset used in this study are available upon request in the event they are temporarily unavailable for download at the above directory. ERSST v4 is available at: https://www.ncdc.noaa.gov/data-access/marineocean-data/extended-reconstructed-sea-surface-temperature-ersst-v4 and NCEP/NCAR Reanalysis is available at: https://psl.noaa.gov/data/gridded/data.ncep.reanalysis.html#citations

Code Availability

All plots and analysis are carried out using Python v. 3.4 including the following packages: xarray, numpy, xesmf, pandas, os, matplotlib and cartopy. The majority of code used for analysis is publicly available at: https://github.com/ubbu36/CMIP6_pacific_analysis (DOI: 343968992)[4]. All code files are available upon reasonable request.

Methods references

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Author contributions

U.K.H and A.V.F. contributed equally to designing the research. U.K.H performed the data analysis and, together with A.V.F., interpreted the results. U.K.H wrote the manuscript and edited it together with A.V.F.

Competing interests

The authors declare no competing interests.

Figure captions

Figure 1. Multi-model-mean temporal evolution of the Pacific zonal surface temperature gradient in different experiments. *(a) Historical simulations and changes in the observed SST gradient (red line, errst v4 data); both are relative to the 1950-1970 baseline. (b,c) Future projection scenarios SSP5-8.5 and SSP3-7.0 relative to the 2015-2025 baseline. (d,e) Abrupt 4xCO_2 and gradual 1pct CO_2-rise experiments relative to the piControl level. In order to weight each model equally, only one ensemble member per model is included. Shading indicates model spread. A ten-year running mean is applied. Negative values indicate the weakening of the temperature gradient.*

***Figure 2**. **Changes in the Pacific zonal surface temperature gradient in different experiments for each model**. These bar charts show zonal temperature gradient changes for (a) the CO_2-only experiments and (b) the full-forcing historical and SSP5-8.5 simulations. Anomalies in the CO_2-only experiments are measured relative to piControl. Anomalies in the historical and SSP5-8.5 scenarios are calculated relative to the 1950-1970 baseline and compared to the observed changes (the red bar). Error bars, when provided, indicate ensemble spread (one standard deviation) for models that include at least 3 ensemble members.*

Figure 3. Initial and long-term SST anomaly patterns at low latitudes developing in the abrupt 4xCO_2 experiments. *The panels show multi-model-mean anomalies relative to the area mean warming for two model categories based on the structure of the initial response. The ocean thermostat (OT) category contains models with a relative cooling (or lack of warming) in the central and eastern equatorial Pacific in the first 25 years. The eastern equatorial Pacific warming (EP) category contains models with a clear warming in the central and eastern equatorial Pacific during the same years. In the long-term response (right panels) both categories show a pronounced EP warming pattern but whose strength depends on the type and strength of the initial response. The models assigned to each category are listed in the legend of Fig. 5. We used 7 and 13 models for the OT and EP categories, respectively. The remaining models fall in between these end-members. To highlight the patterns of change, anomalies are computed with respect to piControl with the ocean mean warming (60 ^{o}E - 60 ^{o}W, 40 ^{o}S - 40^{o}N) subtracted. Hatching indicates that 80% of the models agree on the sign of the anomaly.*

Figure 4. Initial and long-term sea level pressure (SLP) and surface wind anomaly patterns in the abrupt 4xCO_2 experiments. *The plot shows SLP anomalies relative to the mean area change (within 60^{o}E - 60^{o}W, 30^{o}S - 30^{o}N) and wind anomalies (arrows) for the OT and EP model categories and the same times as in Fig 2. Note the long-term reduction in the east-west SLP gradient and westerly wind anomalies along the equator in panels (b) and (d), indicating the weakening of the Walker circulation. This contrasts the strengthening of the Walker circulation in the OT category in panel (a) during the first decades of the experiments.*

Figure 5. Changes in the Pacific zonal surface temperature gradient in different experiments versus the 4xCO_2 initial response: *(a) long-term anomalies in the abrupt 4xCO_2 experiments for years 100-150; (b) anomalies in the 1pct CO_2 experiments for years 20-80; (c) anomalies in the GHG-only historical simulations for years 1990-2014 relative to piControl; (d) anomalies in the full-forcing historical simulations for years 1990-2014 relative to piControl; (e,f) long-term anomalies, defined as 2080-2100 minus piControl, for the SSP3-7.0 and SSP5-8.5 scenarios. Each marker+color combination represents one model as described in the legend below the panels. Models marked as OT or EP, respectively, have a clear initial strengthening or weakening of the Pacific zonal temperature gradient in the first 25 years of the 4xCO_2 simulations. Models that have partial-forcing experiments are denoted with an asterisk. Negative values indicate the weakening of the zonal temperature gradient. Dashed vertical lines indicate the thresholds for the OT and EP categories respectively. The R and p values denote the correlation coefficients and statistical significance of Pearson's correlation.*

Figure 6. Anomalies in surface temperature, sea level pressure (SLP) and surface winds in experiments with full or partial historical forcing*. The maps show multi-model-mean anomalies in surface temperature (left) and SLP and wind (right) for years 2000-2014 relative to the 1950-1970 baseline for three historical experiments: (a,b) with full forcing, and (c,d) with GHG-only and (e,f) aerosols-only forcings. Note that the colorbar for panel (e) is saturated at -0.5 and 0.5, which highlights the pattern of change over the ocean. A subset of 12 models that have all three types of simulations is used (these models are included in Fig. 5c).*