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Yan Xia

yxia24@uci.edu

University of California, Irvine <https://orcid.org/0009-0009-1931-2083>

YONG FU LIN

Research Center for Critical Issues, Academia Sinica <https://orcid.org/0000-0002-9522-7173>

Jin-Yi Yu

University of California, Irvine <https://orcid.org/0000-0001-6156-7623>

Yunfan Chen

University of Michigan

Michael Pritchard

Complex Biological Systems, University of California, Irvine

Article

Keywords:

Posted Date: April 27th, 2026

DOI: <https://doi.org/10.21203/rs.3.rs-9489113/v1>

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Additional Declarations: There is **NO** Competing Interest.

Asymmetric Atlantic-Pacific interactions on decadal timescales underrepresented in climate models

Yan Xia^{1*}, Yong-Fu Lin², Jin-Yi Yu¹, Yunfan Chen³, Mike Pritchard^{1*}

¹Department of Earth System Science, University of California, Irvine, California, USA

²Department of Earth Sciences, National Taiwan Normal University, Taipei, Taiwan

³Earth and Environmental Sciences, University of Michigan, Ann Arbor, Ann Arbor, MI, USA.

Corresponding author: Yan Xia (yxia24@uci.edu), Mike Pritchard (mspritch@uci.edu)

Abstract

Observed tropical Pacific sea surface temperatures have exhibited a sustained La Niña-like cooling trend in the eastern Pacific since 1980, whereas coupled climate models simulate an El Niño-like warming pattern under greenhouse forcing. Atlantic warming has been proposed as a driver of this Pacific cooling through inter-basin teleconnections, yet climate models substantially underestimate this influence. Here we show that Atlantic–Pacific interactions are strongly asymmetric on decadal timescales. Lead–lag analyses of observations and CMIP6 simulations reveal two pathways linking the basins: a short-lag atmospheric bridge through which Pacific variability affects the Atlantic, and a delayed ocean-mediated pathway through which Atlantic variability drives Pacific changes. While CMIP6 models reproduce the Pacific-to-Atlantic response, they fail to sustain the delayed Atlantic-to-Pacific pathway evident in observations. This failure arises primarily from underestimated Atlantic convective sensitivity (~30% of observed), compounded by a weak Bjerknes feedback (~82% of observed), which together prevent sustained thermocline adjustment and subsurface ocean memory in the tropical Pacific. The resulting underrepresentation of the Atlantic-to-Pacific pathway helps explain why climate models struggle to reproduce the observed La Niña-like tropical Pacific trend.

Introduction

Since the 1980s, tropical Pacific sea surface temperatures have exhibited a La Niña-like trend, with warming in the west and cooling in the east^{1–3}. Most Coupled Model Intercomparison Project Phase 6 (CMIP6) simulations forced with historical greenhouse gas concentrations instead produce an El Niño-like pattern, characterized by a weakened zonal SST gradient and amplified eastern Pacific warming^{4–6}. This model-observation discrepancy has broad implications for projections of regional climate⁷, tropical cyclone activity⁸, and equilibrium climate sensitivity⁹, all of which depend on the zonal structure of tropical Pacific warming. Several mechanisms have been proposed to account for the observed cooling, including phase transitions of the Interdecadal Pacific Oscillation^{10–12}, anthropogenic aerosol forcing^{13,14}, Southern Ocean cooling^{15,16}, nonlinear ENSO rectification², and the ocean dynamical thermostat response to greenhouse gas forcing¹⁷. Among these possibilities, inter-basin influence from multidecadal Atlantic warming has received increasing attention, as pacemaker experiments and observational analyses indicate that Atlantic SST anomalies can strengthen the Pacific Walker circulation and induce eastern Pacific cooling on decadal timescales^{2,18–21}.

However, climate models appear to underestimate the strength of this Atlantic influence. With mean-state SST biases in both the tropical Atlantic and Pacific accounted for, the simulated Pacific cooling response to Atlantic warming is reduced by up to 89%^{22,23}. These mean-state SST biases largely persist from CMIP5 into CMIP6^{24,25}, suggesting that the weakened trans-basin response carries over into the current generation of models. Multi-model evaluations further show that Atlantic–Pacific interactions in both directions are systematically underestimated across interannual and decadal timescales^{25–27}, with large inter-model spread linked in part to differences in moist static energy injection from the Atlantic into the upper troposphere²⁷. More broadly, the observed sensitivity of tropical precipitation to local SST exceeds that in most climate models, a deficiency linked to the representation of convective processes at typical GCM resolution²⁸. Together, these systematic biases suggest that contemporary climate models may be missing key processes governing inter-basin interactions between the Atlantic and Pacific Oceans.

Observational and modeling studies suggest that Atlantic and Pacific SST variability influence each other on decadal timescales^{18,29–31}, through two broad classes of processes. One involves rapid atmospheric teleconnections that modify large-scale circulation, including the Walker circulation, allowing SST anomalies in one basin to influence surface winds and convection over the other³². The other involves slower ocean-mediated adjustments, in which wind-driven thermocline variations and subsurface heat redistribution transmit anomalies between basins over longer timescales³³. Although both pathways have been proposed as mechanisms linking Atlantic and Pacific variability, their relative roles in sustaining decadal-scale Atlantic–Pacific interactions have not been systematically evaluated. Moreover, despite the previously documented tendency for CMIP6 models to underestimate the strength of Atlantic–Pacific interactions on decadal

timescales^{22,23,25–27,34}, it remains unclear whether this systematic bias is associated with deficiencies in the atmospheric teleconnection pathway, the ocean-mediated pathway, or both. In particular, it remains unknown whether climate models represent the two directional components of Atlantic–Pacific interactions—Pacific-to-Atlantic and Atlantic-to-Pacific—equally well on decadal timescales.

In this study, we examine the Pacific-to-Atlantic and Atlantic-to-Pacific components of Atlantic–Pacific interactions using CMIP6 historical simulations. We show that climate models reproduce the rapid component of inter-basin interactions through which Pacific variability influences the Atlantic, but systematically underestimate the slower component through which Atlantic variability affects the Pacific. Further analysis indicates that the underestimation of equatorial Atlantic convective sensitivity—partially traceable to mean-state SST biases—plays a primary role in weakening the Walker circulation response to Atlantic SST anomalies. In addition, a weaker-than-observed Bjerknes feedback in the equatorial Pacific further limits the oceanic adjustment to this already-attenuated wind forcing. These results reveal a pronounced asymmetry in simulated Atlantic–Pacific interactions and suggest that the compounding suppression of atmospheric convection and oceanic coupling may contribute to model biases in decadal tropical Pacific SST variability.

Asymmetric Atlantic–Pacific Interaction in Observations and CMIP6 Models

Lead–lag correlations between the indices of Tropical Pacific Decadal Variability (TPDV) and Tropical Atlantic Multidecadal Variability (TAMV) (Methods) over 1891–2014 reveal contrasting structures of Atlantic–Pacific interactions in observations and CMIP6 simulations (Fig. 1).

Observational datasets (HadISST, ERSST, COBE; Fig. 1a) consistently exhibit an asymmetric correlation pattern with two distinct peaks of opposite sign, although with varying magnitudes across products. The observational mean shows a positive correlation ($r \approx 0.39$) at lags of 0–2 years with TPDV leading, and a negative correlation ($r \approx -0.42$) when TAMV leads by approximately 15 years. In contrast, the CMIP6 multi-model mean (Fig. 1b) shows a symmetric, unimodal structure centered near zero lag, with peak correlation ($r \approx 0.63$) at +1 year. Models reproduce the short-lag Pacific-to-Atlantic relationship, albeit with enhanced amplitude, but fail to reproduce the observed negative correlation when TAMV leads, and the small inter-model spread indicates that this behavior is consistent across the ensemble. These results suggest that while CMIP6 models capture the near-synchronous Atlantic response associated with Pacific variability, they fail to reproduce the delayed Pacific response to Atlantic variability evident in observations.

To examine the spatiotemporal structure of Atlantic–Pacific interactions, we apply lead–lag partial regression of SST anomalies onto the TPDV and TAMV indices, controlling for covariability between the two basins (Fig. 2; Methods). At concurrent timescales (i.e., zero lag),

the CMIP6 ensemble reproduces the bidirectional SST response with broadly realistic spatial structures. Under positive TPDV forcing, observations show equatorial Atlantic warming (~ 0.25 K; Fig. 2m); models capture this pattern but with reduced amplitude (~ 0.15 – 0.20 K; Fig. 2n). Under positive TAMV forcing, observations exhibit central–eastern Pacific cooling (~ 0.4 K over 150°W – 100°W ; Fig. 2o); models reproduce the sign and approximate magnitude but with a systematic southward displacement of the cooling center to 5° – 10°S (Fig. 2p). Despite these spatial biases, the concurrent SST response is reasonably reproduced in both directions.

At longer decadal lags, however, systematic deficiencies emerge in both directions. For the Pacific-to-Atlantic pathway, observations show tropical North Atlantic warming at concurrent time (Fig. 2m) that expands across the broader North Atlantic as the TPDV lead increases to 10 and 15 years (Fig. 2e, a). The observed patterns appear in the CMIP6 ensemble mean only at shorter lags and with reduced amplitude (Fig. 2n, j); at 10- and 15-year leads, models display partially reversed Atlantic SST anomalies (Fig. 2f, b). Model deficiencies are more pronounced for the Atlantic-to-Pacific pathway. When TAMV leads, observations show western Pacific warming and central–eastern Pacific cooling, with the anomaly patterns strengthening from concurrent time toward 5-, 10-, and 15-year lags (Fig. 2o, k, g, c), indicating an intensifying La Niña–like response in the Pacific. In contrast, models produce basin-wide Pacific warming that intensifies from the 5-year toward the 10-year lag (Fig. 2l, h). The observed La Niña–like pattern is simulated by the models only at concurrent time (Fig. 2p). The reversed SST anomalies at decadal leads indicate that the processes required to sustain slow Atlantic–Pacific interactions along the Atlantic-to-Pacific pathway are largely absent in current-generation climate models.

The contrasting model performance at concurrent versus longer decadal lags is consistent with the asymmetric versus symmetric lead–lag TPDV–TAMV correlations revealed in Fig. 1a, b, where observations show a significant positive correlation at short decadal lags with TPDV leading and a significant negative correlation when TAMV leads at longer decadal lags. In contrast, the models reproduce the short-lag correlation but fail to simulate the longer-lag decadal correlation. This suggests that current-generation models have the greatest difficulty simulating the Atlantic-to-Pacific pathway of the Atlantic–Pacific interactions, which primarily operates at longer decadal lags.

Power spectral analysis of both indices provides additional support for this interpretation (Fig. 1c, d). To examine variability across the full range of timescales, the TPDV and TAMV indices in this analysis were calculated from unfiltered SST anomalies. The results indicate that CMIP6 models reproduce the essential spectral characteristics of both TPDV and TAMV. Observations show that TPDV exhibits a spectral peak at 3–6 years, whereas TAMV shows pronounced multidecadal variability with a spectral maximum near 55 years. The CMIP6 ensemble reproduces these two dominant spectral bands, although the simulated multidecadal peak is weaker and does not exceed the 95% significance threshold. Given that the intrinsic variability timescales are reasonably represented, the muted decadal Atlantic–Pacific interactions in models

likely arise from deficiencies in the interaction mechanisms rather than from missing basin variability.

Mechanism diagnostics in the following section require atmospheric and oceanic reanalysis fields available only for 1958–2014. Over this shorter period, the observed lead–lag correlation structure — positive when TPDV leads and negative when TAMV leads — is preserved, while the CMIP6 ensemble mean remains symmetric and centered near zero lag (Fig. S1). Spatial regression patterns of SST and 850 hPa wind anomalies (Fig. S2) are consistent with the full-period SST results (Fig. 2), though anomaly magnitudes are amplified at shorter lead times. Under TAMV forcing, the observed wind field exhibits equatorial easterly anomalies over the Pacific that strengthen from concurrent time toward longer lags, corroborating the La Niña–like SST response; models produce no coherent equatorial wind response at decadal lags. These results confirm that the model biases are not sensitive to the choice of analysis period.

Atmospheric Bridge and Oceanic Interaction Mechanisms

To quantify the directionality and temporal evolution of Atlantic–Pacific atmospheric interactions, we analyze the Trans-Basin Walker Circulation Gradient (G_{WC} ; Methods; Figs. S3–S6), defined as the inter-basin difference in column-averaged tropospheric vertical velocity (Fig. 3a, b). We denote G_{WC}^{ATL} when the regression is performed onto the TAMV index and G_{WC}^{PAC} when performed onto the TPDV index, so that the superscript identifies the forcing basin whose influence is being evaluated. Because the atmosphere adjusts rapidly to SST forcing through the Matsuno–Gill response to diabatic heating anomalies³⁵, G_{WC} primarily reflects the fast component of Atlantic–Pacific interactions transmitted through the atmospheric bridge. A positive value of the gradient indicates anomalous ascent over the Pacific and subsidence over the Atlantic, corresponding to a Pacific-to-Atlantic interaction pathway. Conversely, a negative value of the gradient indicates anomalous ascent over the Atlantic and subsidence over the Pacific, corresponding to an Atlantic-to-Pacific interaction pathway. Persistent anomalies in this gradient at longer decadal lags therefore indicate sustained atmospheric forcing capable of initiating the slower oceanic component of Atlantic–Pacific interactions.

The lag partial regression of G_{WC} onto the TPDV and TAMV indices indicates that at concurrent timescales, observations reveal asymmetric interaction strengths between the two forcing directions. Under TPDV forcing (Fig. 3a), the positive G_{WC}^{PAC} response peaks at zero lag and gradually weakens toward longer lags, whereas under TAMV forcing (Fig. 3b) the negative G_{WC}^{ATL} response at zero lag reaches only 74% of the TPDV-forced magnitude, indicating that Atlantic SST anomalies generate weaker cross-basin Walker circulation responses than comparable Pacific SST anomalies. The CMIP6 ensemble captures the observed dominance of Pacific-driven Atlantic–Pacific interactions at concurrent timescales, although their amplitudes are substantially underestimated: the modeled response reaches 53% of the observed magnitude under TPDV forcing but only 37% under TAMV forcing, demonstrating that the Atlantic-to-Pacific

atmospheric pathway is far more attenuated than the Pacific-to-Atlantic pathway even at concurrent timescales.

The lag partial regression of G_{WC} shown in Fig. 3a, b reveals an even stronger asymmetry between the two forcing directions at longer lags. Under TPDV forcing, the observed G_{WC}^{PAC} response persists until approximately an 8-year lag before returning to zero thereafter. Under TAMV forcing, however, observations exhibit a progressively intensifying negative G_{WC}^{ATL} response that peaks when TAMV leads by about 6-8 years. These results indicate that the slow component of the Atlantic–Pacific interactions is dominated by the Atlantic-to-Pacific pathway in observations, in contrast to the dominance of the Pacific-to-Atlantic pathway for the short-lag component of the interactions. However, CMIP6 models fail to reproduce this asymmetric dominance of the Atlantic-to-Pacific pathway for the slow component of the Atlantic–Pacific interactions. The modeled response to TAMV forcing peaks weakly at a 0-year lag and vanishes by about year 4. Thus, although CMIP6 models capture the fast Walker circulation response associated with TAMV forcing, they cannot sustain Atlantic-driven circulation anomalies beyond concurrent timescales.

To understand why models fail to reproduce the slow Atlantic-to-Pacific component of Atlantic–Pacific interactions, we diagnose the oceanic processes responsible for the delayed Pacific response to Atlantic variability. We examine the temporal evolution of equatorial zonal wind stress, thermocline depth, and SST anomalies along the equatorial Pacific under TAMV forcing (Fig. 3c–h). Regressions of zonal wind stress reveal coherent easterly anomalies confined to the western–central equatorial Pacific (140°E–160°W) when TAMV leads by 8–12 years, with peak values of -0.006 to -0.008 Pa σ^{-1} near 160°E–180° and gradually weakening toward zero lag (Fig. 3c). In CMIP6 models, this signal is reversed in sign and reduced to amplitudes below ± 0.002 Pa σ^{-1} across all lags and longitudes, without coherent spatial structure (Fig. 3d).

The resulting thermocline adjustment reflects the Bjerknes feedback³⁶ response to this sustained easterly forcing. Observations show a strengthened zonal tilt of the D20 isotherm, with western deepening of $+6$ to $+8$ m σ^{-1} near 140°E–180° and eastern shoaling of -0 to -2 m σ^{-1} near 120°W–100°W, emerging when TAMV leads by approximately 12 years and peaking TAMV leads by 8–10 years (Fig. 3e). In contrast, CMIP6 models fail to reproduce this structure, instead showing a weak reversed tilt of less than ± 2 m σ^{-1} and lacking coherent spatial organization (Fig. 3f).

These subsurface adjustments subsequently influence the surface temperature response. Observations show persistent cooling across the central-to-eastern Pacific under TAMV forcing from concurrent time through 15-year leads, with peak cooling exceeding -0.8 °C σ^{-1} near 100°W when TAMV leads by approximately 5 years and the cooling signal extending westward to 180°–140°W when TAMV leads by 15 years (Fig. 3g). By contrast, the CMIP6 SST response is reversed in sign across most lags, with basin-wide warming reaching $+0.4$ °C σ^{-1} near 100°W when TAMV leads by 6-8 years, and only weaker central–to-eastern Pacific cooling appearing

during the final three years before zero lag (Fig. 3h). This reversed SST pattern is consistent with the absence of sustained wind forcing and thermocline adjustment shown in Figs. 3d and 3f.

The CMIP6 model deficiencies identified above are not sensitive to the choice of analysis period. Comparison of the CMIP6 MMM diagnostics over 1891–2014 and 1958–2014 (Fig. S8) shows consistent results across both periods: G_{WC}^{ATL} remains near zero at all leads (Fig. S8a, b), equatorial wind stress lacks coherent easterly anomalies (Fig. S8c, d), and the thermocline response is weak and reversed (Fig. S8e, f). The SST response shows basin-wide warming at decadal leads in both periods, although the cooling response at short lead times is weaker over 1891–2014 (Fig. S8g) than over 1958–2014 (Fig. S8h), consistent with the reduced anomaly amplitudes in the full-period SST regression maps (Fig. 2o, k). These results confirm that the mechanism diagnostics presented over 1958–2014 reflect systematic model deficiencies rather than artifacts of the shorter analysis window.

Convective sensitivity deficiency and cascading effects on Atlantic-to-Pacific interactions

To identify the cause of model deficiencies in the Atlantic-to-Pacific pathway, we examine the relationship between local convective sensitivity and cross-basin interactions. Convective sensitivity is defined as the pointwise regression of precipitation onto SST anomalies (Methods), quantifying the efficiency with which SST anomalies generate diabatic heating and associated vertical motion. Averaged over the equatorial Pacific (EP; 180°–280°E, 5°S–5°N) and equatorial Atlantic (EA; 280°E–340°E, 5°S–5°N), convective sensitivity exhibits significant inter-model relationships with the EA–EP SST gradient (EP: $R = -0.51$, $p = 0.004$; EA: $R = 0.39$, $p = 0.034$; Fig. 4a, b). Models with stronger EP convective sensitivity tend to produce a reduced Atlantic–Pacific SST contrast, whereas stronger EA sensitivity is associated with an enhanced contrast. Although the CMIP6 ensemble captures the observed EA–EP gradient with only modest bias, the underlying convective sensitivities are distributed asymmetrically. EP convective sensitivity aligns well with observations, consistent with the relatively realistic fast Pacific-to-Atlantic atmospheric response. In contrast, EA convective sensitivity is markedly underestimated, with the observational estimate lying well outside the model spread.

The link between local convective sensitivity and the fast atmospheric bridge is confirmed by the inter-model relationship between convective sensitivity and the concurrent Walker circulation response. In the Pacific, EP convective sensitivity is significantly correlated with $G_{WC}^{PAC}(0)$ ($R = 0.59$, $p = 0.001$; Fig. 4c): models with stronger convection generate a stronger Pacific-to-Atlantic circulation response, though the CMIP6 ensemble slightly underestimates $G_{WC}^{PAC}(0)$ relative to observations. In the Atlantic, models and observations occupy distinct regimes (Fig. 4d). The observational estimate shows high EA convective sensitivity paired with a strong $G_{WC}^{ATL}(0)$ response, whereas the CMIP6 ensemble clusters in a region of substantially suppressed sensitivity and near-zero circulation response; the inter-model relationship is statistically insignificant ($R = -0.23$, $p = 0.246$). This displacement indicates that models fail to generate

sufficient deep convective responses and the associated vertical motion per unit Atlantic warming.

The substantial underestimation of EA convective sensitivity is partially traceable to mean-state SST biases. The tropical Atlantic (TA; 280°–340°E, 20°S–20°N) exhibits a cold bias of approximately -0.27 K in the MMM, with individual models reaching -1.5 K (Fig. 4e), while the tropical Pacific (TP; 180°–280°E, 20°S–20°N) exhibits a complementary warm bias of approximately $+0.21$ K (Fig. 4f). Lower saturation humidity, set by Clausius–Clapeyron scaling, reduces the boundary-layer moisture available for convection per unit SST anomaly; a smaller departure of SST from the convective threshold places a smaller fraction of SST variability within the regime of efficient deep convection³⁷. A warmer mean state enhances both. Consistent with these effects, the warm-biased tropical Pacific produces convective sensitivity with little bias, whereas the cold-biased tropical Atlantic substantially underestimates it (Fig. 4g).

The CMIP6 ensemble underestimates EA convective sensitivity, with the MMM reproducing only approximately 30% of the observed value (0.79 versus 2.68 mm day⁻¹ K⁻¹; Fig. 4g), consistent with the widespread underestimation of tropical precipitation sensitivity to local SST documented across climate models²⁸. Grouping models by the sign of their TA bias reveals a partial but systematic link to mean-state SST. Models with a warm TA bias (red markers) produce EA convective sensitivity closer to observations — 73% of warm-bias models exceed the MMM value, compared with 42% of cold-bias models. However, cold-bias models show no systematic relationship between their SST bias magnitude and convective sensitivity, and even warm-bias models remain far below the observed value. These results suggest that the mean-state SST bias contributes to, but does not fully account for, the suppressed Atlantic convective response. The residual underestimation may reflect deficiencies in convection parameterization at typical GCM resolutions, a well-documented source of uncertainty in simulated tropical precipitation^{28,38}.

The suppressed convective sensitivity substantially attenuates the concurrent Walker circulation response. The simulated Atlantic-to-Pacific Walker circulation response in the CMIP6 MMM ($G_{WC}^{ATL}(0)$; Fig. 4h) reproduces only about 37% of the observed magnitude. TA warm-bias models show stronger $G_{WC}^{ATL}(0)$, consistent with the direct link between convective sensitivity and the circulation response, whereas TA cold-bias models show no systematic improvement. Beyond this attenuation of the Atlantic-to-Pacific atmospheric bridge, the Pacific oceanic adjustment to the weakened wind stress is further limited by a reduced Bjerknes feedback efficiency. The simulated Pacific Bjerknes feedback parameter (μ ; Fig. 4i) — quantifying the coupling between equatorial wind stress and thermocline displacement over the equatorial central–eastern Pacific—is approximately 18% weaker than observed in the MMM (approximately -6.0×10^{-4} Pa m⁻¹ versus the observed -7.3×10^{-4} Pa m⁻¹). TA warm- and cold-bias models show no systematic difference in μ (Fig. 4i), as expected for an oceanic process independent of Atlantic mean state. This reduction limits the amplification available through the

wind–thermocline coupled feedback, reducing the conversion of wind forcing into thermocline displacement.

These compounding deficiencies culminate in the near-total suppression of the delayed thermocline response within the tropical Pacific to Atlantic forcing. The thermocline gradient index G_{D20} in the equatorial Pacific is evaluated at a 6-year lag, corresponding to the lag of peak observed G_{WC}^{ATL} under TAMV forcing (see Fig. 3b), and therefore representing the period of maximum sustained Atlantic-induced atmospheric forcing acting on the tropical Pacific Ocean. Observations exhibit a strong delayed equatorial Pacific thermocline gradient ($G_{D20} \approx +4.2 \text{ m } \sigma^{-1}$ at a 6-year lag (Fig. 4j), while the delayed CMIP6 MMM response is slightly negative ($-0.55 \text{ m } \sigma^{-1}$). This sign reversal is consistent with the reversed thermocline tilt and the absence of eastward-propagating subsurface signals in the tropical Pacific (Fig. 3e–f), indicating that the suppression of both atmospheric forcing and oceanic coupling prevents the buildup of subsurface ocean memory in the tropical Pacific required for delayed Pacific SST responses to TAMV forcing (Fig. 3g–h). TA warm-bias models produce G_{D20} closer to observations (Fig. 4j), reflecting the accumulated benefit of stronger atmospheric forcing acting through an oceanic pathway that is itself unaffected by the TA bias.

The results above reveal a coherent physical chain linking Atlantic convective sensitivity deficiency to the suppressed Atlantic–Pacific interactions in CMIP6 models. In observations, Atlantic warming induces strong convection, sustaining Walker circulation anomalies over decadal timescales. This generates persistent easterly wind stress in the tropical Pacific, forcing thermocline adjustment through Bjerknes feedback and producing eastward-propagating subsurface signals that culminate in delayed Pacific SST cooling about 8–15 years later. In contrast, CMIP6 models underestimate Atlantic convective sensitivity, a bias that may reflect both mean-state SST biases and convective parameterization deficiencies, producing transient Walker circulation anomalies that decay within about 1–3 years. The resulting weak Pacific wind forcing prevents thermocline adjustment, eliminating oceanic memory and confining the Atlantic–Pacific interaction to concurrent timescales only. The Atlantic convective sensitivity deficiency therefore disrupts the coupled atmospheric–oceanic processes required to sustain the slow Atlantic-to-Pacific interaction pathway, explaining the systematic underrepresentation of the decadal Atlantic–Pacific interaction in CMIP6 models.

Discussion

In summary, our results reveal a strongly asymmetric and timescale-dependent structure of Atlantic–Pacific interactions. Observations exhibit a short-lag Pacific-to-Atlantic atmospheric pathway and a delayed Atlantic-to-Pacific pathway sustained by wind-driven oceanic adjustment on decadal timescales. In contrast, CMIP6 models reproduce the short-lag Pacific-to-Atlantic response but fail to sustain Atlantic-driven circulation anomalies beyond a few years. This selective breakdown arises from compounding deficiencies in two processes. (1) The primary

deficiency is the substantial underestimation of equatorial Atlantic convective sensitivity, which weakens the Walker circulation response to Atlantic SST anomalies. (2) This atmospheric bias is further compounded by a weaker-than-observed Bjerknes feedback in the equatorial Pacific, which limits the amplification of the already-attenuated wind forcing into thermocline displacement. Together, these deficiencies prevent sustained wind forcing from being converted into thermocline adjustment and subsurface ocean memory, eliminating the delayed Pacific response to Atlantic forcing. The compounding suppression of atmospheric forcing and oceanic coupling therefore emerges as the primary constraint on the representation of decadal Atlantic-to-Pacific interactions in current climate models.

The suppressed Atlantic-to-Pacific coupling identified here has important implications for interpreting recent tropical Pacific climate trends (Fig. 5c, d). Since 1980, observations show persistent eastern Pacific cooling and a strengthened zonal SST gradient^{1,3,11}, whereas CMIP6 simulations produce enhanced eastern Pacific warming under the same forcing^{22,23,34,39}. The tropical Atlantic has warmed since the late 1990s, and our results suggest that such Atlantic warming can reinforce eastern Pacific cooling through the delayed Atlantic-to-Pacific interaction pathway identified in this study. However, the convective sensitivity deficiency in CMIP6 models weakens the atmospheric and oceanic responses to Atlantic SST anomalies, producing transient Walker circulation anomalies that cannot sustain the wind forcing required for delayed Pacific thermocline adjustment. The muted representation of this pathway in climate models suggests that projections of tropical Pacific climate and related regional impacts remain subject to substantial uncertainty.

The convective sensitivity deficiency extends beyond the equatorial Atlantic. CMIP6 models also underestimate convective sensitivity over the western Pacific warm pool (Fig. 5a, b), where previous work¹⁹ showed that Atlantic warming induces secondary convection that strengthens the Walker circulation from the western to the eastern Pacific. The muted convective response in this region suggests that the indirect Atlantic–Indian Ocean–Pacific pathway may also be weakened in current models. Taken together, these deficiencies across multiple tropical basins point to systematic limitations in representing convection at typical GCM resolutions. Recent kilometer-scale simulations¹⁵ demonstrate that resolving mesoscale ocean eddies and stratocumulus cloud feedbacks can reproduce observed Southern Ocean and eastern Pacific cooling, a mechanism complementary to the Atlantic pathway examined here. It would be worth analyzing whether the Atlantic SST–precipitation relationship is stronger in the kilometer-scale coupled simulations than in CMIP6 models, which would directly test the mechanism proposed here.

Our conclusions are based on lead–lag partial regression analyses, which isolate basin-scale relationships but cannot fully establish causality; targeted pacemaker experiments would provide more direct evidence. A comparison of diagnostics over 1958–2014 and 1891–2014 shows that the interaction biases are amplified in the more recent period (Fig. S8), when anthropogenic warming has intensified and Atlantic SSTs have risen substantially. Whether this amplification will continue under future greenhouse forcing remains an important question for near-term

climate prediction. Improving the representation of tropical convection and equatorial ocean dynamics, together with expanded observations of subsurface processes, will therefore be essential for reducing these systematic biases and improving confidence in projections of decadal tropical climate variability.

Methods

Observational, Reanalysis, and Model Datasets

Observed sea surface temperature (SST) variability is characterized using three independent monthly datasets spanning 1891–2014: HadISST⁴⁰, ERSSTv5⁴¹, and COBE-SST⁴². ERSSTv5 fields, provided at $2^\circ \times 2^\circ$ resolution, were bilinearly interpolated to $1^\circ \times 1^\circ$ to match the native resolution of HadISST and COBE-SST. Agreement across these products, which employ distinct reconstruction and bias-correction methods, provides a measure of observational uncertainty.

Mechanism diagnostics — including Walker Circulation response, equatorial wind stress, thermocline depth, and convective sensitivity — require atmospheric and oceanic reanalysis fields that are not available over the full SST record. Atmospheric circulation is diagnosed using monthly vertical pressure velocity (ω), precipitation, and 850 hPa wind from the ERA5 reanalysis⁴³ at 0.25° resolution. Oceanic dynamical fields are taken from ORAS5⁴⁴, including monthly 20°C isotherm depth (D20), used as a proxy for thermocline depth, and zonal wind stress (τ_x). Both ERA5 and ORAS5 fields are analyzed over the common period 1958–2014, determined by the ORAS5 start date. All reanalysis fields were interpolated to a common $1^\circ \times 1^\circ$ grid for consistency with SST products (Supplementary Table 1).

Model evaluation is based on historical simulations from 27 CMIP6 models⁴⁵ covering 1850–2014 (Supplementary Table 2). For each model, we analyze the r1i1p1f1 realization and use monthly outputs of SST, ω , precipitation, surface wind stress, and subsurface potential temperature. The D20 is computed from model potential temperature profiles. SST fields are analyzed over 1891–2014 to match the observational record. For mechanism diagnostics, model atmospheric and oceanic fields are analyzed over 1958–2014 to align with the reanalysis period; results using the extended 1891–2014 model record is shown in the Supplementary Information to confirm that model biases are not sensitive to the choice of analysis period. Because atmospheric and oceanic variables are not available for all models, the number of models used for each diagnostic varies (see Supplementary Table 2). All model fields were bilinearly interpolated to the common $1^\circ \times 1^\circ$ observational grid.

Indices

Low-frequency tropical variability is represented using two area-weighted SST indices computed from linearly detrended anomalies²⁵:

- Tropical Pacific Decadal Variability (TPDV): standardized 11-year low-pass filtered SST anomalies averaged over 180° – 280°E , 20°S – 20°N .

- Tropical Atlantic Multidecadal Variability (TAMV): standardized 11-year low-pass filtered SST anomalies averaged over 280°–340°E, 0°–30°N.

The TPDV domain extends meridionally to 20° latitude to capture the broader structure of the Interdecadal Pacific Oscillation, distinguishing it from the equatorially confined ENSO pattern. The TAMV domain isolates the tropical branch of Atlantic Multidecadal Variability, identified in pacemaker experiments as the primary driver of atmospheric teleconnections to the Indo-Pacific^{34,46}. Both indices are computed over the full 1891–2014 SST record to retain low-frequency variance relevant to decadal analysis; defining the indices over the longer record preserves the multidecadal spectral content that would be attenuated by the limited degrees of freedom in a shorter window. When regressed against reanalysis or model fields available only for 1958–2014, the indices are truncated to the corresponding period without redefinition.

Statistical Filtering and Significance Testing

All time series and spatial fields were linearly detrended to remove secular anthropogenic warming, and the climatological seasonal cycle was removed by subtracting monthly means. An 11-year Lanczos low-pass filter⁴⁷ was applied to isolate decadal-to-multidecadal variability and suppress high-frequency interannual fluctuations associated with ENSO.

Statistical significance is assessed using a two-tailed Student's t-test. Effective degrees of freedom (N_{eff}) are computed following Pyper and Peterman⁴⁸ to account for serial autocorrelation introduced by filtering. For the CMIP6 multi-model mean, a one-sample t-test across the 27 models identifies regions where the ensemble response is robust and significantly different from zero at the 90% confidence level.

Spectral analysis

Spectral characteristics of the climate indices were analyzed using the Blackman–Tukey method⁴⁹, a non-parametric approach that estimates the power spectral density (PSD) from the Fourier transform of the autocovariance sequence. A maximum lag window of $M = 480$ months (40 years) was used to balance spectral resolution and variance reduction. This choice provides sufficient degrees of freedom to resolve decadal and multidecadal periodicities while maintaining statistical stability given the 124-year observational record ($N = 1488$ months).

To distinguish physically meaningful oscillations from stochastic internal variability, spectral peaks were tested against a red-noise null hypothesis⁵⁰. The background spectrum was modeled as a first-order autoregressive [AR(1)] process, with parameters estimated from the lag-1 autocorrelation of each time series. Significance levels were evaluated at the 95% confidence interval, with upper confidence bounds derived from the χ^2 distribution. Effective degrees of freedom were determined by the window type and the ratio N/M , ensuring a rigorous assessment of whether spectral peaks exceed the expected background variability.

Lead-Lag Partial Regression Framework

A critical challenge in diagnosing inter-basin teleconnections is the non-stationary temporal covariability between the Pacific and Atlantic basins. To disentangle the independent climatic impacts of each basin, we employ a lead-lag partial regression analysis. For a given climate variable $V(t)$ (e.g., global SST or ω), its relationship with the standardized TPDV and TAMV indices at a lead time κ is modeled as:

$$V(t) = \beta_{pac}(\kappa) \cdot TPDV(t - \kappa) + \beta_{atl}(\kappa) \cdot TAMV(t - \kappa) + \epsilon(t)$$

where $\beta_{pac}(\kappa)$ and $\beta_{atl}(\kappa)$ represent the partial regression coefficients for the Pacific and Atlantic indices, respectively, and τ varies from -15 to +15 years. This multivariate framework allows for the isolation of the signal associated with one basin while linearly controlling for the influence of the other at the same lag.

Trans-Basin Walker Circulation Gradient

To quantify the directionality of the zonal atmospheric bridge, we define the Trans-Basin Walker Circulation Gradient index (G_{wc}). We first vertically average the partial regression coefficients of ω over the troposphere (1000–200 hPa) to capture the deep overturning circulation. The zonal gradient is then calculated as the difference between the column-averaged ω anomalies in the tropical Atlantic and Pacific sectors:

$$G_{wc}(\kappa) = \bar{\omega}_{atl}(\kappa) - \bar{\omega}_{pac}(\kappa)$$

where $\bar{\omega}_{atl}$ is averaged over (300°E-360°E, 5°S-5°N) and $\bar{\omega}_{pac}$ is averaged over (180°-280°E, 5°S-5°N). We denote $G_{wc}^{ATL}(\kappa)$ when the regression is performed onto the TAMV index and $G_{wc}^{PAC}(\kappa)$ when performed onto the TPDV index, so that the superscript identifies the forcing basin whose influence is being evaluated.

The sign of G_{wc} serves as a diagnostic for the dominant teleconnection pathway:

$G_{wc} < 0$ (Atlantic-to-Pacific): Indicates anomalous ascent over the Atlantic and subsidence over the Pacific, consistent with Atlantic forcing driving Pacific cooling.

$G_{wc} > 0$ (Pacific-to-Atlantic): Indicates anomalous subsidence over the Atlantic and ascent over the Pacific, consistent with Pacific forcing driving Atlantic variability.

Convective sensitivity metric

Thermodynamic coupling efficiency between SST and atmospheric convection is quantified using the convective sensitivity, defined as the regression of precipitation onto SST:

$$S_{conv} = \frac{\partial Prec}{\partial SST}$$

where precipitation and SST are 11-year low-pass filtered, linearly detrended monthly anomalies with the climatological seasonal cycle removed. The regression is first computed at each grid point and then area-averaged over the equatorial strip (5°S – 5°N) within the TAMV longitude range (280°E – 340°E) and the TPDV longitude range (180° – 280°E) to obtain a single convective sensitivity estimate for each basin in each model and in observations. Higher values indicate more efficient generation of precipitation per unit SST warming, a prerequisite for strong diabatic heating anomalies and Walker Circulation response.

Oceanic pathway diagnostics

Subsurface oceanic adjustment is diagnosed using equatorial zonal wind stress (τ_x), 20°C isotherm depth (D20), and SST regressed onto TAMV. Wind stress regressions are computed along the equator (5°S – 5°N) for lags κ from -15 to 5 years. D20, a proxy for thermocline depth, is similarly regressed onto TAMV.

Following Bellenger et al.⁵¹, the Bjerknes feedback parameter (μ) is estimated as the linear regression slope of area-averaged equatorial zonal wind stress (τ ; 160°E – 150°W , 5°S – 5°N) onto the zonal thermocline gradient index, defined as the difference between D20 averaged over the central-to-eastern Pacific (150°W – 110°W , 5°S – 5°N) and the western Pacific (130°E – 160°E , 5°S – 5°N). In contrast to Bellenger et al.⁵¹, who quantified the atmospheric wind response to SST anomalies, this formulation directly measures the oceanic wind–thermocline coupling relevant to decadal Pacific adjustment. A more negative μ indicates stronger coupling between equatorial wind stress and thermocline tilt.

Data and code availability

Observational SST datasets are publicly available:

HadISST at <https://www.metoffice.gov.uk/hadobs/hadisst/>,

ERSSTv5 at <https://psl.noaa.gov/data/gridded/data.noaa.ersst.v5.html>,

and COBE-SST at <https://psl.noaa.gov/data/gridded/data.cobe.html>.

ERA5 reanalysis is accessible at <https://cds.climate.copernicus.eu>.

ORAS5 ocean reanalysis is available at <https://cds.climate.copernicus.eu>.

CMIP6 model outputs are available through the Earth System Grid Federation at <https://esgf-node.llnl.gov/search/cmip6/>.

Analysis scripts are available from the corresponding author upon reasonable request.

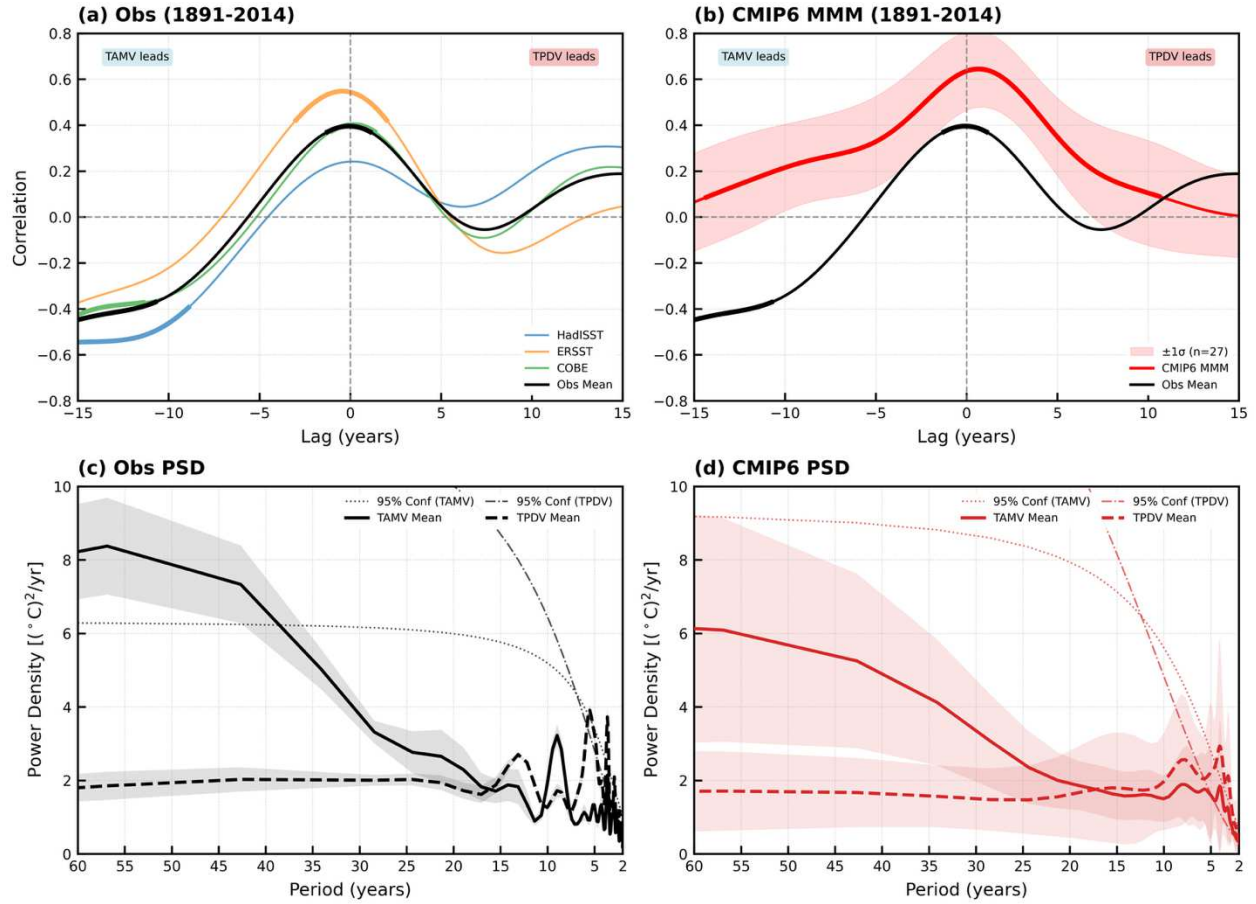


Figure 1 | Lead-lag structure and spectral characteristics of Atlantic-Pacific interactions in observations and CMIP6 models. **a, b** Lead-lag correlations between the Tropical Pacific Decadal Variability (TPDV) and Tropical Atlantic Multidecadal Variability (TAMV) indices. **a** Observational datasets (HadISST, ERSST, COBE) are shown in colors, and the observational mean in black. **b** CMIP6 multi-model mean lead-lag correlations. Shading indicates $\pm 1\sigma$ inter-model spread. **c, d** Power spectral density of TAMV (solid) and TPDV (dashed). To examine variability across the full range of timescales, the indices are calculated from unfiltered SST anomalies. **c** Spectra from observations. **d** Spectra from CMIP6 models. Dotted lines indicate the 95% confidence level and shading represents spectral uncertainty.

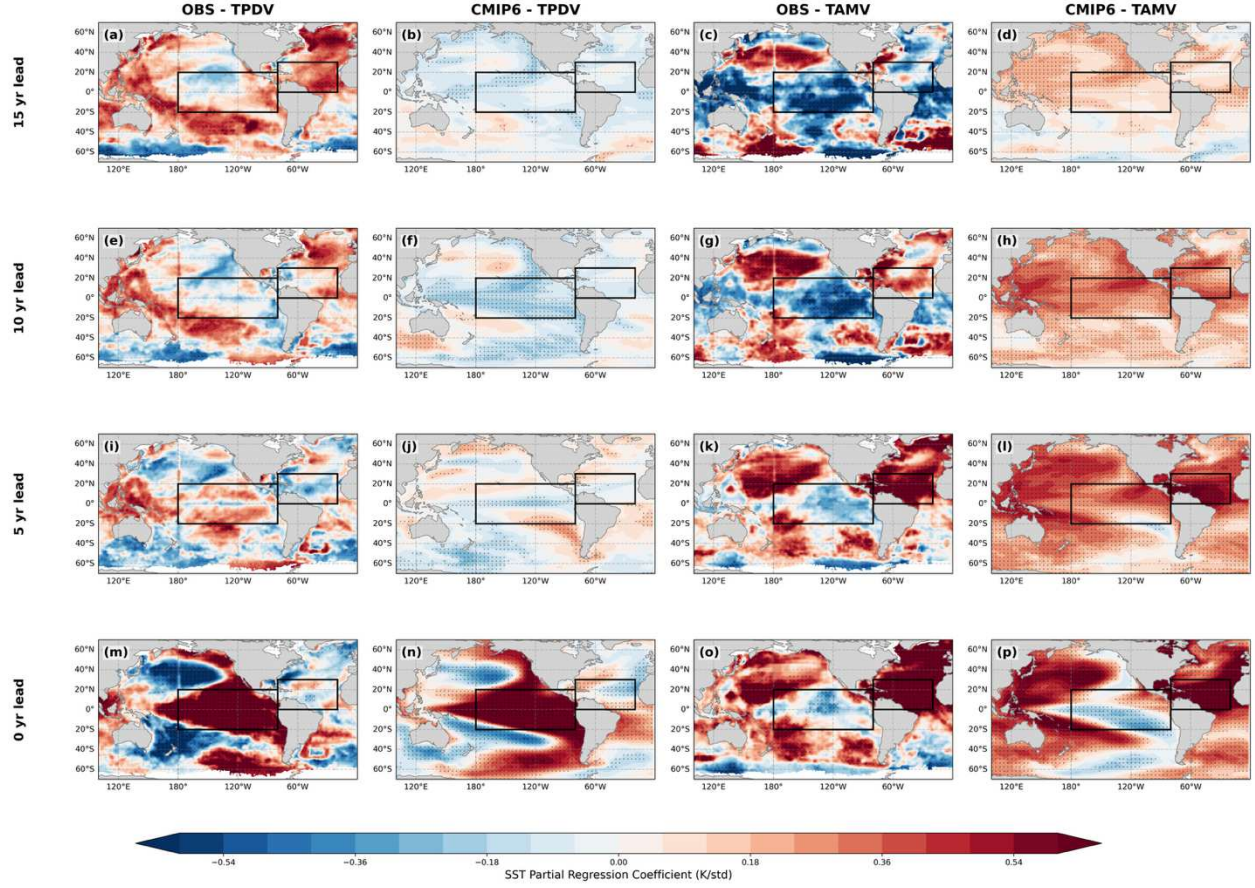


Figure 2 | Partial regression of SST anomalies onto the TPDV and TAMV indices at multiple lead times (1891–2014). Shading shows partial regression coefficients of linearly detrended, 11-year Lanczos low-pass filtered SST anomalies ($K \sigma^{-1}$) onto the TPDV index (a–b, e–f, i–j, m–n) and the TAMV index (c–d, g–h, k–l, o–p). For each index, observations (multi-product mean of HadISST, ERSSTv5, and COBE SST) and the CMIP6 multi-model mean (27 models) are shown in alternating columns. Rows correspond to the predictor index leading by 15 years (a–d), 10 years (e–h), 5 years (i–l), and 0 years (m–p). Black boxes delineate the TPDV domain (180°–280°E, 20°S–20°N) and the TAMV domain (280°–340°E, 0°–30°N). All coefficients are expressed per standard deviation of the respective index. Stippling denotes significance at the 90% confidence level (two-tailed Student's t -test with effective degrees of freedom corrected for serial autocorrelation; one-sample t -test across models for the CMIP6 mean).

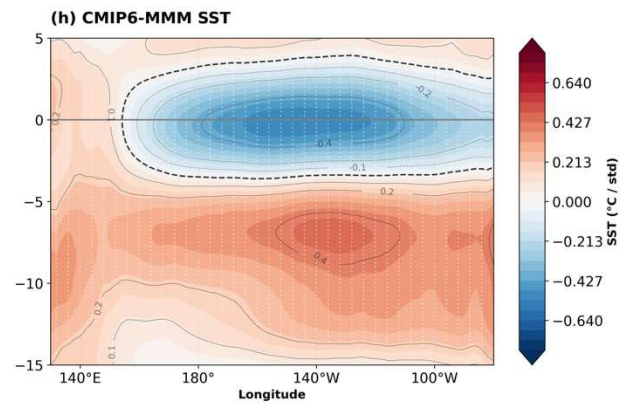
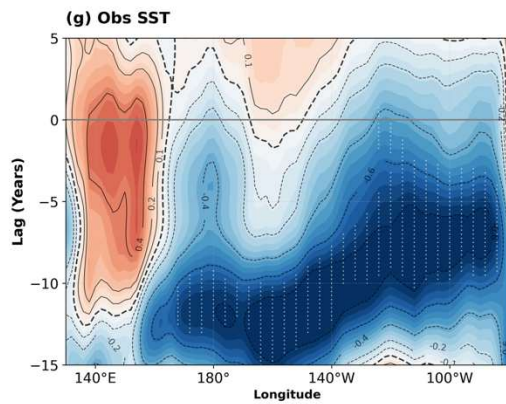
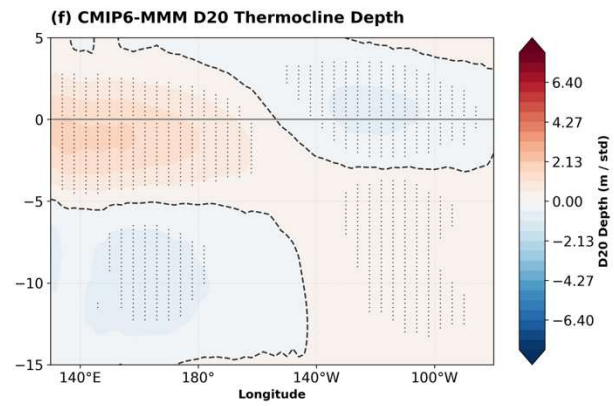
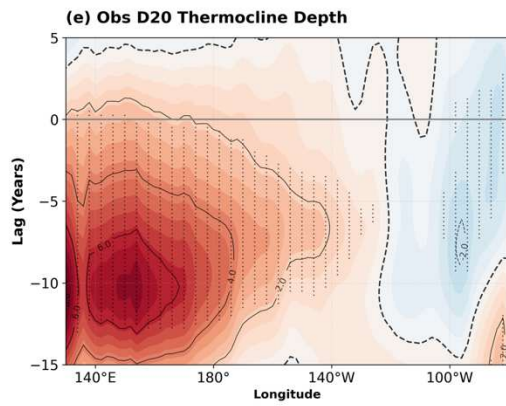
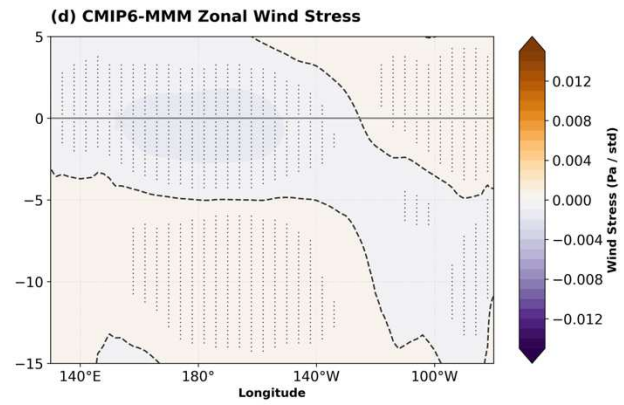
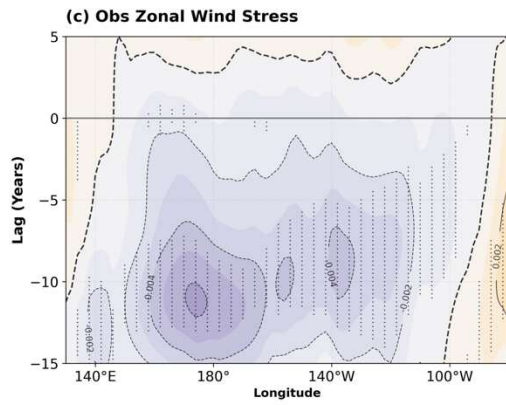
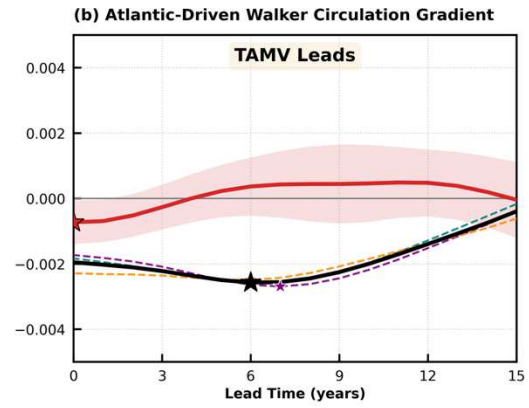
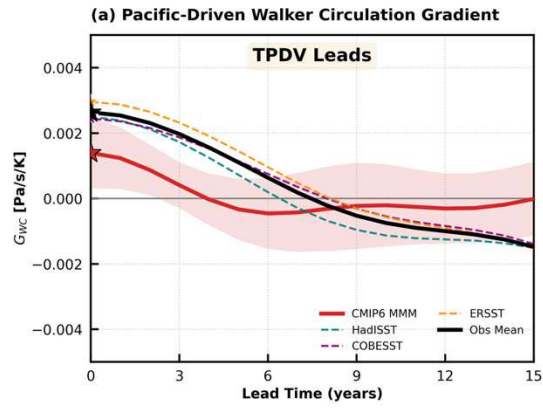
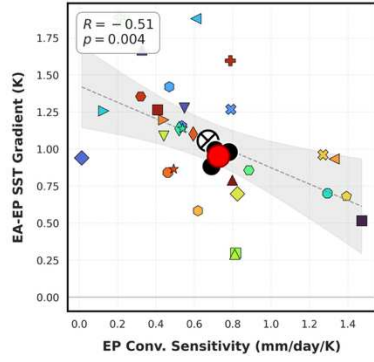
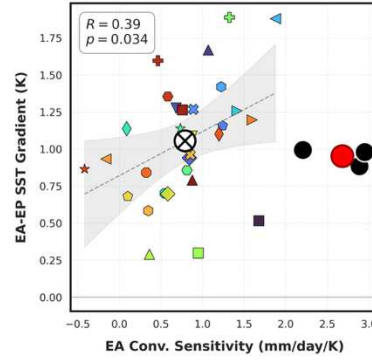


Figure 3 | Atmospheric and oceanic responses associated with Atlantic–Pacific interactions in observations and CMIP6 models (1958-2014). **a-b:** Lead–lag partial regression of the trans-basin Walker circulation gradient (G_{WC} ; Methods) onto the (a) TPDV and (b) TAMV indices. Shading indicates $\pm 1\sigma$ inter-model spread for CMIP6 simulations. **c–d:** Lead–lag partial regression of equatorial zonal wind stress anomalies ($\text{Pa } \sigma^{-1}$) onto TAMV in (c) Observations and (d) CMIP6 models. **e–f:** Lead–lag partial regression of thermocline depth anomalies (D20; $\text{m } \sigma^{-1}$) along the equatorial Pacific onto TAMV; in (e) observations and (f) CMIP6 multi-model mean. **g–h:** Lead–lag partial regression of sea surface temperature anomalies ($\text{K } \sigma^{-1}$) along the equatorial Pacific onto TAMV in (g) Observations and (h) CMIP6 models.

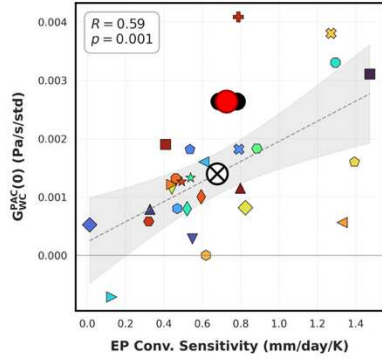
(a) EP Conv. Sensitivity vs EA-EP Gradient



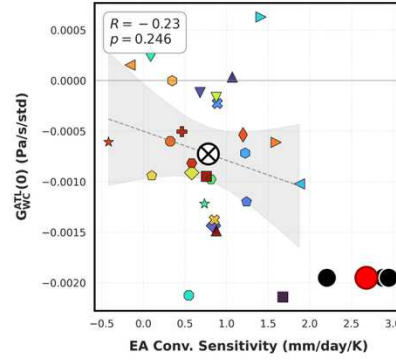
(b) EA Conv. Sensitivity vs EA-EP Gradient



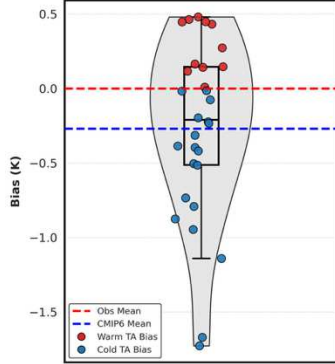
(c) $G_{WC}^{PAC}(0)$ vs EP Conv. Sensitivity



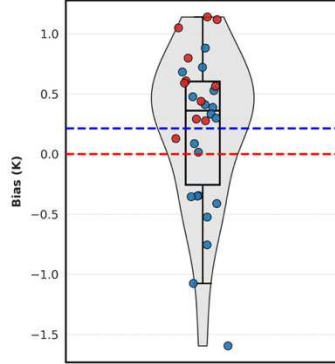
(d) $G_{WC}^{ATL}(0)$ vs EA Conv. Sensitivity



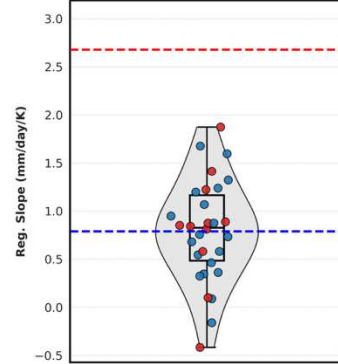
(e) TA SST Bias



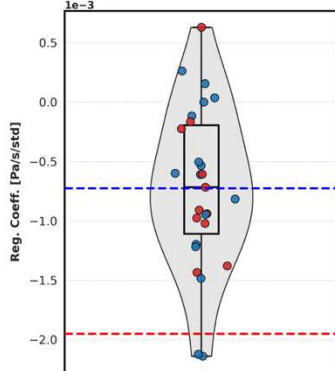
(f) TP SST Bias



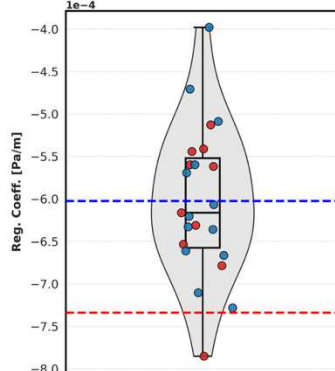
(g) EA Conv. Sensitivity



(h) $G_{WC}^{ATL}(0)$



(i) Pacific Bjerknes (μ)



(j) Pacific G_{D20} (+6yr)

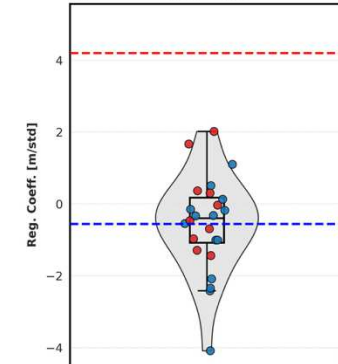


Fig. 4 | Relationship between convective sensitivity, mean-state SST biases, and the Walker circulation response to Atlantic forcing in CMIP6 . **a,b**, Convective sensitivity ($\text{mm day}^{-1} \text{ K}^{-1}$) versus the equatorial Atlantic–Pacific SST gradient (K) for the (a) equatorial Pacific (EP; 180° – 280°E , 5°S – 5°N) and (b) equatorial Atlantic (EA; 280° – 340°E , 5°S – 5°N). **c,d**, Convective sensitivity versus the concurrent Walker circulation gradient ($G_{\text{WC}}(0)$) for the (c) Pacific-to-Atlantic and (d) Atlantic-to-Pacific pathways. In a–d, colored symbols denote individual CMIP6 models, the crossed circle the multi-model mean, and the black star observations; dashed lines and shading show the least-squares fit with 95% confidence interval. **e–j**, Violin plots of CMIP6 ensemble diagnostics arranged along the causal chain: (e) tropical Atlantic (TA; 280° – 340°E , 20°S – 20°N) SST bias (K), (f) tropical Pacific (TP; 180° – 280°E , 20°S – 20°N) SST bias (K), (g) EA convective sensitivity, (h) $G_{\text{WC}}^{\text{ATL}}(0)$, (i) Bjerknes feedback parameter (μ ; Pa m^{-1}), and (j) equatorial Pacific thermocline gradient (G_{D20} ; m std^{-1}) at 6-year lag. Red and blue dashed lines mark observed and MMM values, respectively; red (blue) circles denote models with a warm (cold) tropical Atlantic bias.

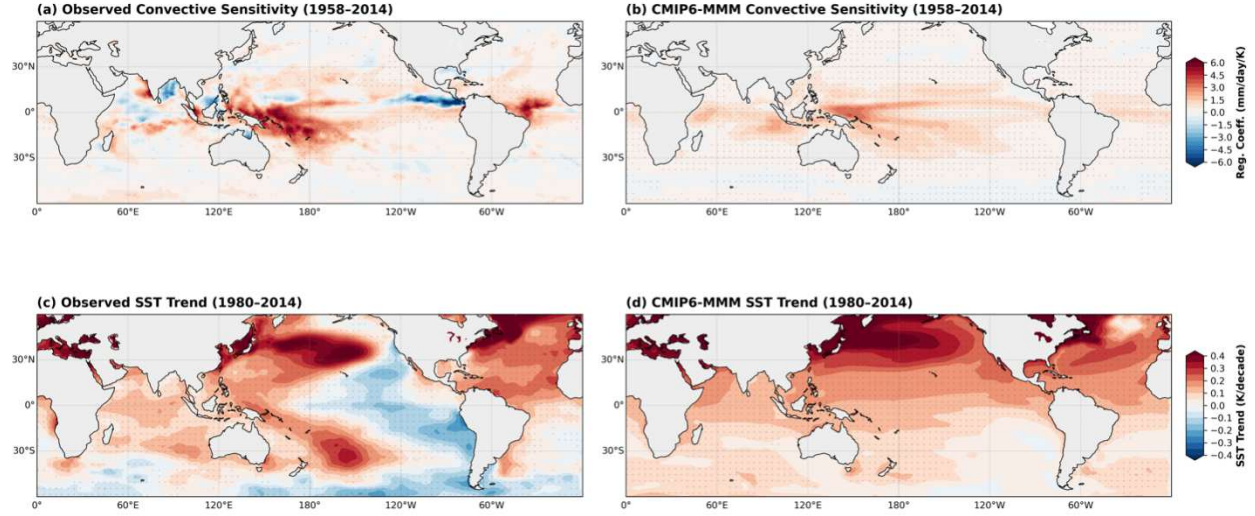


Figure 5 | Convective sensitivity and SST trend biases in CMIP6 models. **a,b,** Pointwise regression of precipitation onto SST anomalies ($\text{mm day}^{-1} \text{K}^{-1}$; 1958–2014), showing convective sensitivity in (a) observations (ERA5 precipitation, multi-product SST mean) and (b) the CMIP6 multi-model mean. **c,d,** Linear SST trend (K decade^{-1} ; 1980–2014) in (c) observations and (d) the CMIP6 multi-model mean. Stippling denotes significance at the 90% confidence level (two-tailed Student's t -test with effective degrees of freedom corrected for serial autocorrelation; one-sample t -test across models for the CMIP6 mean).

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