

Explaining the Equatorial Pacific Thermocline Response to Climate Change with a Model Hierarchy

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Key Points:

- Climate models can largely recreate observed equatorial Pacific subsurface temperature trends from 1958-2020
- The observed subsurface temperature change is due to wind changes, remote SST changes, and local SST changes
- Our results suggest that canonical drivers of the equatorial Pacific response to climate change have been misattributed

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Abstract

Most studies of the equatorial Pacific response to anthropogenic forcing have focused on patterns of sea surface temperature (SST) change. However, similar SST patterns can be consistent with a range of different subsurface responses, each with differing physical and biogeochemical implications. While historical observation and climate model mismatches have been suggested in the literature, we show that model simulations can largely capture the observed 1958-2020 subsurface temperature trend in the equatorial thermocline. We then analyze a hierarchy of idealized model simulations, consisting of fully-coupled, mechanically-decoupled, ocean-only, and reduced gravity models, to understand which ocean dynamics contribute to this response. We show that the response of the thermocline to idealized climate change can be explained by a combination of decadal Bjerknes-like momentum dynamics and radiatively-forced buoyancy-driven dynamics. We further decompose the buoyancy-driven pattern into a pattern driven by remote, subtropical SST forcing and a pattern driven by local, equatorial SST forcing. The remote-SST-forced pattern of thermocline warming shows the signature of dynamic and thermodynamic subtropical cell adjustments. Meanwhile, increased stratification in the local-SST-forced pattern both coherently shoals the thermocline and relaxes thermocline tilt to largely cool the thermocline. Considered together, we recreate the long-term subsurface equatorial Pacific response to idealized greenhouse gas forcing as a linear combination of (i) wind-stress-driven changes, (ii) remote buoyancy-driven changes, and (iii) local buoyancy-driven changes. To conclude we discuss implications for recent temperature trends, revisit canonical theories of the ocean dynamical thermostat, and show the insensitivity of forced responses to forcing geography.

Plain Language Summary

While most research on the equatorial Pacific response to climate change has focused on surface ocean temperatures, ocean circulation changes provide insight into the dynamic drivers of that response. For instance, though eastern equatorial Pacific surface cooling could result from either strengthening winds or subsurface communication of subtropical cooling, these mechanisms affect the subsurface distinctly. In this work we first show that climate model simulations can largely capture the observed 1958-2020 equatorial Pacific subsurface temperature response to climate change. We then use a series of idealized modeling simulations, from a complex global climate model to a sim-

49 ple primitive equation model, to explain the ocean dynamics that create this response.
50 Our central result is that the equatorial Pacific subsurface temperature response to cli-
51 mate change is a simple linear sum of the ocean’s response to changes in winds, changes
52 in remote sea surface temperature patterns, and changes in local sea surface tempera-
53 ture patterns. We explore the dynamics of each of these individual responses. Last, we
54 show that this understanding does not explain subsurface temperature patterns since the
55 late 1970s, and we discuss how our results suggest a reinterpretation of how certain ocean
56 dynamics shape the equatorial ocean’s response to climate change.

57 **1 Introduction**

58 A strong zonal gradient in sea surface temperature (SST) exists in the equatorial
59 Pacific between the western Pacific warm pool and the eastern Pacific cold tongue. This
60 zonal SST gradient is the most obvious manifestation of a series of coupled atmospheric
61 and oceanic processes that connect easterly trade winds, westward surface currents, an
62 eastward subsurface return flow within an upward tilting thermocline, and upwelling of
63 cool sub-thermocline waters in the eastern equatorial Pacific (Bjerknes, 1969; Wyrтки,
64 1975). Variability in the equatorial Pacific mean state shifts the location of atmospheric
65 deep convection and excites planetary waves that propagate to the extratropics to af-
66 fect global climate (Horel & Wallace, 1981). Across a broad range of time-scales, from
67 interannual changes of the El Niño-Southern Oscillation (ENSO, Philander, 1983) to decadal
68 changes of the Pacific Decadal Oscillation (PDO, Mantua et al., 1997), the equatorial
69 Pacific is a key driver and pacemaker for global climate (e.g., Kosaka & Xie, 2016).

70 Given its outsize influence on Earth’s climate, it is critical to understand how the
71 equatorial Pacific will respond to anthropogenic greenhouse gas emissions (e.g., DiNezio
72 et al., 2009; Xie et al., 2010). A key question, which has received much attention and
73 debate, is: How will the equatorial zonal SST gradient change in the future? Constraints
74 from atmospheric thermodynamics have been primarily invoked in support of a decreas-
75 ing gradient (i.e., more warming in the eastern than western equatorial Pacific): enhanced
76 evaporative cooling in the warmer western Pacific can more readily balance anomalous
77 radiative forcing than the cooler eastern Pacific (Knutson & Manabe, 1995; Merlis & Schnei-
78 der, 2011), and the atmospheric Walker circulation slow-down implied by specific humid-
79 ity changes would relax thermocline tilt (Vecchi & Soden, 2007). Meanwhile, Clement
80 et al. (1996) and Seager and Murtugudde (1997) proposed the “ocean dynamical ther-

81 mostat” and suggested that the zonal SST gradient should in fact increase. The ther-
82 mostat theory suggests that the eastern equatorial Pacific should warm by less than the
83 rest of the tropics because upwelled equatorial waters, originating in the extratropics and
84 reaching the equator via the oceanic subtropical cells (STCs: Liu, 1994; McCreary Jr &
85 Lu, 1994), will not show an effect of surface forcing until some time later. More recently,
86 several studies have suggested that this debate is simply a matter of time-scales, with
87 a brief strengthening of the zonal gradient eventually giving way to a long-term weak-
88 ening (Luo et al., 2017; Heede et al., 2020, 2021; Heede & Fedorov, 2021).

89 However, the continued inability of coupled models to recreate recent historical equa-
90 torial Pacific SST trends (Coats & Karnauskas, 2017; Seager et al., 2019, 2022; Watan-
91 abe et al., 2021; Wills et al., 2022) calls into question this seeming resolution. While ob-
92 servational products suggest that the western Pacific has warmed and central-eastern Equa-
93 torial Pacific has cooled since the beginning of the satellite era (e.g., Karnauskas et al.,
94 2009; Solomon & Newman, 2012; Seager et al., 2019; Wills et al., 2022), over that same
95 period the vast majority of coupled global climate models (GCMs) show enhanced warm-
96 ing in the eastern Pacific relative to the western Pacific. Many studies have attempted
97 to explain the mismatch between observed and modeled equatorial Pacific SST trends,
98 pointing to i) mismatched internal variability (e.g., Laepple & Huybers, 2014; Olonscheck
99 et al., 2020; Watanabe et al., 2021; Heede & Fedorov, 2023; Jiang et al., 2024a), ii) in-
100 correct model processes that could otherwise create observed trends (e.g., McGregor et
101 al., 2018; Baldwin et al., 2021; Dong et al., 2022; Kang et al., 2023; Hwang et al., 2024),
102 and iii) systematic biases in model mean states that do not allow a forced response to
103 establish (Seager et al., 2019, 2022; Heede & Fedorov, 2023; Jiang et al., 2024a, 2025).

104 While the mismatch between observed and modeled historical equatorial SST has
105 been extensively discussed, the subsurface equatorial temperature response to climate
106 change has been relatively understudied. This top-down focus on SST alone potentially
107 obfuscates important subsurface oceanic adjustments (e.g., Clement et al., 1996; Vec-
108 chi & Soden, 2007) that have helped to shape the SST response. For instance, despite
109 comprising entirely different subsurface dynamics, both decreased thermocline tilt (Vecchi
110 & Soden, 2007; Luongo et al., 2023) and coherent thermocline deepening (Luongo et al.,
111 2025) could theoretically lead to an El Niño-like SST pattern.

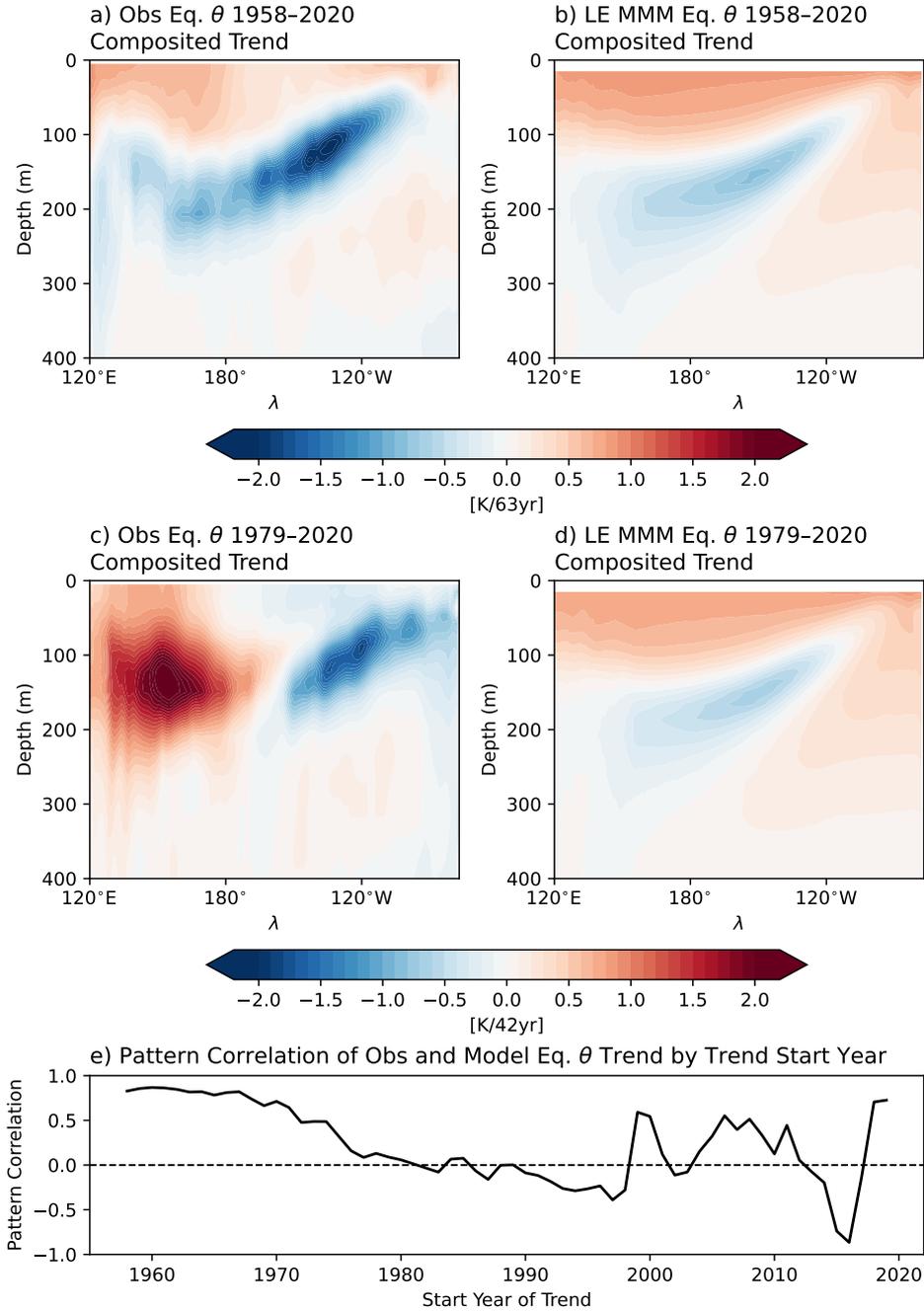


Figure 1. a) 1958-2020 annual-mean equatorial temperature (θ) trend composited from EN04 and Ishii observational products and ORAS5 and SODA2.2.4 reanalyses. b) Multi-model mean equatorial θ from 11 large ensemble simulations over 1958-2020. c) As in panel a) but for the period of 1979-2020. d) As in panel b) but for the period of 1979-2020. e) Pearson pattern correlation between observational and multi-model mean equatorial θ trend as a function of trend start year.

112 Following Jiang et al. (2025), we show a composite of the 1958-2020 subsurface equa-
 113 torial (meridionally averaged from 5°S-5°N) Pacific temperature trend from two obser-
 114 vational products [EN04 (1958-2020, Good et al., 2013) & Ishii (1958-2012, Ishii & Ki-
 115 moto, 2009)] and two ocean reanalyses [ORAS5 (1958-2020, Zuo et al., 2019) & SODA2.2.4
 116 (1958-2010, Carton & Giese, 2008)] (Figure 1a). The most obvious feature of subsurface
 117 equatorial temperatures over the past 60 years is a broad thermocline cooling. Due to
 118 the upward and eastward tilt of the thermocline, this cooling is around 250m deep in the
 119 west-central Pacific and around 50m deep in the eastern Pacific. Maximum cooling oc-
 120 curs between 150°W-120°W from approximately 100-175m. While the cooling is the most
 121 eye-catching feature in this observational pattern, we also note a broad surface warm-
 122 ing, which is minimized in the central Pacific and extends deeper in the western Pacific
 123 than eastern Pacific, and a sub-thermocline warming in the eastern Pacific. These fea-
 124 tures are largely shared by individual observational products (Figure S1).

125 The observed 1958-2020 subsurface temperature trend in Figure 1a is similar to the
 126 1951-2010 trend pattern in Watanabe et al. (2021). We note, however, that the specific
 127 time period considered greatly influences this pattern: Figure 1a is markedly different
 128 than both the 1979-2020 pattern (Figure 1c) and the 1979-2013 pattern (Watanabe et
 129 al., 2024). The observed trend over this shorter period features a striking zonal temper-
 130 ature dipole, western Pacific warming and the eastern Pacific cooling, within the top 200m.
 131 This temperature dipole dynamically agrees with the upper ocean circulation strength-
 132 ening noted by Tuchen et al. (2024) over the overlapping period of 1993-2022.

133 In a series of recent studies inspired by Seager et al. (2019)'s hypothesis that the
 134 models' mean state ocean is simply too biased to capture observed SST trends, Jiang
 135 et al. (2024a, 2024b, 2025) highlight the differences in the subsurface trend patterns be-
 136 tween observations and models. In particular, the authors hypothesize that the observed
 137 subsurface cooling response can be explained as a forced response to wind changes (Jiang
 138 et al., 2024a, 2024b) and that models lack an effective connectivity between subsurface
 139 and surface eastern Pacific temperatures due to insufficient upwelling and mixing (Jiang
 140 et al., 2025). However, the corresponding 1958-2020 subsurface temperature trend com-
 141 posite from a suite of 11 large ensemble simulations from the sixth coupled model inter-
 142 comparison project [Figure 1b, inspired by Jiang et al. (2025)] shows a similar pattern
 143 to observations (Figure 1a; Pearson pattern correlation of 0.83). Both show a broad ther-
 144 mocline cooling, a somewhat zonally symmetric surface warming, and a sub-thermocline

145 eastern Pacific warming. Indeed, despite model biases, there is general agreement be-
146 tween observed and modeled subsurface temperature trends when those trends begin be-
147 fore ~ 1975 (Figure 1e), suggesting that the modeled subsurface temperature trend can
148 still inform our understanding of the observed subsurface temperature trend and the cou-
149 pled dynamics that have created it. This perspective motivates the central questions of
150 our study: 1) which ocean dynamics contribute to this common modeled response, and
151 2) to what extent is the full response a simple linear combination of these responses?

152 A handful of studies have explored the models' shared central-western Pacific ther-
153 mocline cooling response (c.f. Figure 1b and Figure S2) which has persisted for several
154 model generations. Vecchi and Soden (2007) suggest that this cooling is a local effect
155 caused by a reduction in thermocline tilt due to a weaker atmospheric Walker circula-
156 tion. Yang et al. (2009) suggest that this weaker Walker circulation slows down the STCs
157 and dynamically cools the equatorial subsurface. Luo et al. (2009, 2018) agree that mod-
158 els' STCs have slowed, but suggest that a major cause of the slowdown is increased sub-
159 tropical surface stratification. Finally, Ju et al. (2022) suggest that the cooling is caused
160 by mean advection of density-compensated spiciness anomalies from the subtropics, which
161 cool the region as much as dynamical changes in subtropical cell circulation.

162 While these studies provide a starting point for answering our guiding questions,
163 it's evident that these proposed mechanisms are entwined with murky causality. A com-
164 mon means of circumventing the attribution issues common to coupled dynamics is to
165 employ a model hierarchy to step through a complex response by iteratively removing
166 complexity until the phenomenon of interest is isolated. For instance, recent studies have
167 overrode surface wind stress to mechanically decouple a GCM's ocean from its atmosphere
168 (e.g., Luongo et al., 2024) and have shown that the ocean's full response to an anoma-
169 lous forcing can be linearly partitioned into the response due to anomalous surface buoy-
170 ancy forcing and anomalous surface momentum forcing (Luongo et al., 2022, 2023). Sim-
171 ilarly, ocean-only GCM (OGCM) simulations are a convenient way to isolate just the ocean's
172 response to a forcing without changes in the atmosphere (e.g., Peng et al., 2022).

173 In this study we employ a model hierarchy, consisting of a fully-coupled GCM, a
174 mechanically-decoupled GCM, an OGCM, and a primitive equation reduced gravity model,
175 to explore the modeled subsurface equatorial Pacific temperature response to greenhouse
176 gas forcing and which ocean dynamics contribute to it. We discuss the simulations that

177 comprise this hierarchy in section 2 and in section 3 we show that the full response can
 178 be understood as a linear sum of the response due to i) momentum effects, ii) remote
 179 buoyancy effects, and iii) local buoyancy effects. We discuss implications of these results
 180 in section 4 and we conclude in section 5.

181 **2 Model Hierarchy**

182 All simulations used to explore the equatorial Pacific thermocline response to green-
 183 house gas forcing are presented in Table 1. Simulations that explore the equatorial Pa-
 184 cific thermocline response to non-greenhouse-gas forcing schemes are presented in Ta-
 185 ble S1.

186 **2.1 Fully-Coupled Simulations**

187 We analyze pre-existing simulations using the National Center for Atmospheric Re-
 188 search’s Community Earth System Model, Version 1.2 (CESM1: Hurrell et al., 2013) that
 189 were initially presented in Luongo et al. (2022) and Taylor et al. (2025). These simula-
 190 tions have a nominal horizontal resolution of 2° in the atmosphere and land components
 191 and 1° in the ocean and sea ice components, though the oceanic component has higher
 192 resolution near the equator to better resolve equatorial dynamics. They are run in a stan-
 193 dard coupled configuration with pre-industrial forcing (“B1850” compset) for fifty years.

194 Our preindustrial control (Ctrl) simulation extends directly from initialization with
 195 no anomalous forcing applied. To idealize climate change we apply and maintain an abrupt
 196 quadrupling of CO_2 relative to pre-industrial levels ($\text{CO}_2 \times 4$). While abrupt quadrupling
 197 of CO_2 is an obvious simplification compared to a more realistic time-evolving increase
 198 in CO_2 , we show in section 3.1 that this idealization works remarkably well. While we
 199 primarily focus on the equatorial Pacific’s response to greenhouse gas forcing in this study,
 200 we also consider simulations with hemispherically asymmetric forcing to test how robust
 201 the ocean dynamics of interest are to forcing geometry. We apply a zonally-uniform top-
 202 of-atmosphere (TOA) insolation reduction following the Extratropical-Tropical Interac-
 203 tion Model Intercomparison Project (ETINMIP: Kang et al., 2019) protocol in the North-
 204 ern Hemisphere (NH, 45°N - 65°N) for ETINMIPNH and in the Southern Hemisphere (SH,
 205 45°S - 65°S) for ETINMIPSH. The ETINMIP forcing corresponds to an annual-mean, zonal-
 206 mean forcing of approximately -45 Wm^{-2} at 55° N or S and falls off as an approximate

Fully-coupled Simulations		
Simulation Name	CO ₂ Forcing	Wind Stress
Ctrl	280ppm	Freely evolving
CO ₂ x4	1120ppm	Freely evolving
Mechanically-decoupled Simulations		
Simulation Name	CO ₂ Forcing	Wind Stress
Tau1CO ₂ x1	280ppm	Ctrl
Tau1CO ₂ x4	1120ppm	Ctrl
Tau4CO ₂ x1	280ppm	CO ₂ x4
Ocean-only Simulations		
Simulation Name	SST Forcing Perturbation	SST Forcing Bounds
OCtrl	n/a	n/a
CO ₂ x4.BFsst	Tau1CO ₂ x4-Tau1CO ₂ x1	90°S-90°N
CO ₂ x4.BFsstET	Tau1CO ₂ x4-Tau1CO ₂ x1	90°S-6°S, 6°N-90°N
CO ₂ x4.BFsstEQ	Tau1CO ₂ x4-Tau1CO ₂ x1	10°S-10°N
NEPac2CWarm	+2°C	147°W-123°W, 22°N-32°N
Reduced Gravity Simulations		
Simulation Name	Reduced Gravity	
RGCtrl	1x	
RGx2	2x	

Table 1. Details of the fully coupled, mechanically-decoupled, ocean-only, and reduced-gravity simulations used to study the equatorial Pacific thermocline response to greenhouse gas forcing. Table S1 presents simulations that explore alternate forcing schemes.

207 Gaussian. For all CESM1 simulations we consider an average over years 11-50 after the
 208 forcing is applied as in Luongo et al. (2022, 2023).

209 **2.2 Mechanically-Decoupled Simulations**

210 We perform wind stress overriding simulations (e.g., Luongo et al., 2024) to iso-
 211 late the dynamic effect of buoyancy and momentum flux anomalies on the ocean, while
 212 still maintaining some amount of realistic atmosphere-ocean coupling. In a fully-coupled
 213 simulation the coupler interactively provides the ocean component with atmospheric wind
 214 stress. This momentum flux coupling then drives changes in the ocean state (e.g., equa-
 215 torial thermocline tilt which changes the zonal SST gradient), which can then feed back
 216 on the atmosphere in the next coupling step (e.g., Bjerknes feedback). In wind stress over-
 217 riding simulations, however, the GCM is instead modified to receive a known surface wind
 218 stress field, disabling the interactive coupling of momentum fluxes and mechanically de-
 219 coupling the ocean from the atmosphere. All other coupling, including the effect of wind
 220 speed on turbulent heat fluxes, remain in tact.

221 In our mechanically-decoupled simulations we either apply a radiative forcing (CO_2
 222 quadrupling or ETINMIP TOA forcing) and lock to Ctrl's wind stress field, or we ap-
 223 ply no radiative forcing but we lock to a perturbed simulation's wind stress field. For
 224 example, we perform a simulation where we abruptly quadruple CO_2 , but we override
 225 with unperturbed Ctrl wind stress ($\text{Tau1CO}_2\text{x4}$). This simulation highlights the radiatively-
 226 driven climate response because wind stress is unperturbed. We also perform a simu-
 227 lation where we apply no CO_2 forcing, but we override with the perturbed wind stress
 228 field from $\text{CO}_2\text{x4}$ to highlight the climate response just due to wind stress ($\text{Tau4CO}_2\text{x1}$).
 229 Similarly, we perform a simulation where we apply a reduction in insolation in the NH
 230 following the ETINMIP protocol described above, but we override with unperturbed Ctrl
 231 wind stress ($\text{Tau}_1\text{S_NH}$), and we perform a simulation where we apply no insolation
 232 reduction, but we override with the perturbed wind stress field from ETINMIPNH (Tau_NH_S_1).
 233 $\text{Tau}_1\text{S_SH}$ and Tau_SH_S_1 are similar, but correspond to the SH ETINMIP simula-
 234 tions.

235 In these wind overriding simulations we prescribe the full interannually-varying wind
 236 stress field to maintain the impact of high-frequency mechanical variability on the sur-
 237 face ocean and reduce mean state biases (Luongo et al., 2024). Finally, in order to re-
 238 move mean state biases introduced by the mechanical decoupling technique, we compare

239 these perturbed mechanically-decoupled simulations to a control mechanically-decoupled
 240 simulation (Tau1CO₂x1), which has no radiative forcing and wind stress locked to Ctrl.
 241 As such, our buoyancy-forced (BF) response is Tau1CO₂x4-Tau1CO₂x1 and our momentum-
 242 forced (MF) response is Tau4CO₂x1-Tau1CO₂x1. See discussion in Luongo et al. (2022,
 243 2023, 2024) for more detail on this protocol.

244 **2.3 Ocean-only Simulations**

245 We use an ocean-only version of the Massachusetts Institute of Technology Gen-
 246 eral Circulation Model (MITgcm) in the same configuration used in Luongo et al. (2025),
 247 which is similar to the Estimated the Circulation and Climate of the Ocean version 4
 248 release 4 (ECCOv4r4: Forget et al., 2015) configuration. This OGCM has 1° horizon-
 249 tal resolution in the zonal direction and 1/3° meridional resolution at high and low lat-
 250 itudes (telescoping to 1° in midlatitudes). While not fully permitting high latitude mesoscale
 251 eddies, the higher resolution in the equatorial region begins to resolve equatorial waves
 252 and thus decrease tropical biases. This MITgcm configuration is forced with monthly
 253 climatologies of net air-sea fluxes of heat, freshwater, shortwave radiation, and zonal and
 254 meridional momentum diagnosed by Peng et al. (2022) from a 25-year control integra-
 255 tion of ECCOv4r4 with bulk formulae forcing. In addition to climatological flux forc-
 256 ing, we restore SST and sea surface salinity to Peng et al. (2022)’s climatologies on a 10-
 257 day timescale. All of our OGCM simulations branch from a 100-year spin-up, at which
 258 point the upper-ocean is approximately equilibrated. See Luongo et al. (2025) for fur-
 259 ther details.

260 Our OGCM control simulation (OCtrl) is integrated for 30 years. We also perform
 261 a series of perturbation experiments where we add the anomalies in the BF SST field di-
 262 agnosed from our mechanically-decoupled CESM1 simulations to the SST relaxation field.
 263 We add anomalies in the quasi-equilibrium (average over years 11-50) SST field calcu-
 264 lated from Tau1CO₂x4-Tau1CO₂x1, Tau_1_S_NH-Tau1CO₂x1, and Tau_1_S_SH-Tau1CO₂x1
 265 (producing ocean-only simulations CO₂x4_BFsst, ETINMIPNH_BFsst, and ETINMIPSH_BFsst,
 266 respectively). Comparing these SST-forced perturbation experiments with OCtrl shows
 267 the ocean-only dynamic response to the buoyancy-driven SST response. Because any tem-
 268 perature response within the first few depth levels is more likely to be driven by the strong
 269 surface relaxation than by a dynamical adjustment, we focus our discussion of the dy-
 270 namics revealed by the OGCM simulations on the thermocline and below.

271 Finally, we split these perturbation experiments geographically into SST forcing
 272 from only the extratropical regions and SST forcing from only the equatorial region (“ET”
 273 or “EQ” appended to above names). In the extratropical SST forcing experiment we ap-
 274 ply the full CESM1 buoyancy-driven SST anomaly field from 90°S-11°S and 11°N-90°N.
 275 We linearly taper this forcing to zero over 5° to 6°S and 6°N to avoid artificially large
 276 meridional forcing gradients. There is no anomalous SST forcing from 5°S-5°N in the
 277 extratropical SST forcing experiment. Similarly, in the equatorial SST forcing experi-
 278 ment we apply the full CESM1 buoyancy-driven SST anomaly field from 5°S-5°N, cre-
 279 ate a 5° linear taper to 10°S and 10°N, and do not anomalously force SST anywhere else.
 280 While we somewhat arbitrarily chose these meridional boundaries, we have found that
 281 using 10°S and 10°N for the full forcing bounds of the equatorial SST forced simulation
 282 (and changing other bounds accordingly) leads to small differences that do not affect our
 283 conclusions (not shown). Last, as a point of comparison with the extratropical SST forc-
 284 ing simulation described above, we consider the NEPac2CWarm simulation originally
 285 presented in Luongo et al. (2025), where a rectangular patch of +2°C SST warming is
 286 applied off the coast of California. For all OGCM simulations we consider the average
 287 over years 11-30 after the forcing is applied as in Luongo et al. (2025).

288 **2.4 Reduced Gravity Simulations**

289 Finally, we run two simulations using the 1.5-layer reduced gravity (RG) model to
 290 represent an idealization of the upper ocean’s response to an increase in surface strat-
 291 ification. The set-up is borrowed directly from Sun and Thompson (2020), with geog-
 292 raphy consisting of three idealized ocean basins representing the Atlantic, Indian, and
 293 Pacific oceans and a zonally re-entrant channel representing the Southern Ocean from
 294 45°S to the southern boundary. The total width is 220° and the latitudinal extent is from
 295 72°S-72°N. The model solves for upper layer thickness, approximating thermocline depth,
 296 and has a 1° horizontal spacing. Sun and Thompson (2020) provide further details on
 297 this model.

298 We run two reduced gravity simulations. The first is a control simulation using stan-
 299 dard parameters from Sun and Thompson (2020) (RGCtrl). The second branches from
 300 the tenth year of RGCtrl and instantaneously doubles the reduced gravity parameter to
 301 represent a stratification increase in response to climate change forcing (RGx2). We com-
 302 pare the difference between these two simulations ten years after that branch point.

3 Results

3.1 Buoyancy and Momentum Dynamics Create the Full Response

Despite the fact that our CESM1 simulations idealize climate change as an abrupt quadrupling of CO₂, the upper-ocean quasi-steady fully-coupled (FC) response (Figure 2a) bears a striking resemblance to the multi-ensemble mean response to realistic historical forcing in Figure 1b (Pearson pattern correlation of 0.94). As in the multi-model response, the FC equatorial temperature response features a thermocline cooling in the central Pacific, a strong surface warming that is deeper in the western Pacific than the eastern Pacific, and a sub-thermocline warming in the eastern Pacific. Luo et al. (2018) show a very similar CESM1 response to abrupt quadrupling of CO₂ (over years 41-90), which in turn also resembles the transient response (years 1-10) to this same forcing in two prior versions of CESM (Heede et al., 2021). It is interesting that the equatorial Pacific’s subsurface temperature response to abrupt idealized climate change matches the response pattern to historical forcing so well, particularly since Heede et al. (2021) showed that the prominent thermocline cooling response to abrupt quadrupling disappears after 200 years and is instead replaced by a near-zero temperature response. While the centennial response to abrupt 4xCO₂ forcing is an unrealistic analog for transient climate change, the multi-decadal response to abrupt 4xCO₂ forcing shown in Figure 2a appears to be an appropriate tool for studying the multi-model response to greenhouse warming.

As in Luongo et al. (2023), who instead consider the subsurface equatorial temperature response to NH ETINMIP forcing, we find that the FC equatorial temperature response to CO₂ forcing (Figure 2a) is highly linear. That is, the fully-coupled (FC) response to CO₂ forcing can be neatly linearly decomposed into the buoyancy-forced (BF) response (Figure 2b) and the momentum-forced (MF) response (Figure 2c):

$$FC \approx BF + MF . \tag{1}$$

Comparing Figure 2a with the sum of Figures 2b & c explicitly shows this linearity (Figure S3a). These patterns closely resemble the FC, BF, and MF responses to NH ETINMIP forcing (but with opposite sign because ETINMIP forcing is a cooling) presented in Luongo et al. (2023) and recreated in Figures S4a-c. This resemblance is also found

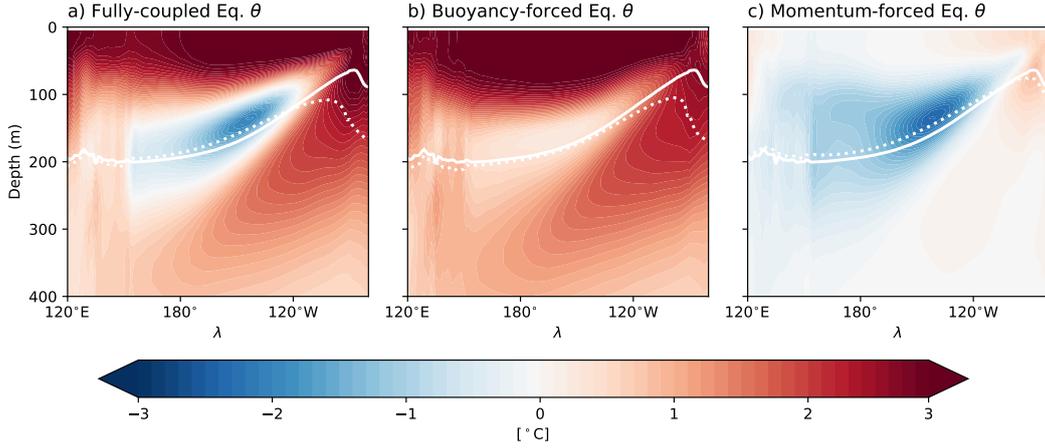


Figure 2. a) CESM1 fully-coupled (FC = CO₂x4-Ctrl) equatorial temperature (θ) response to abrupt quadrupling of CO₂. b) CESM1 buoyancy-forced (BF = Tau1CO₂x4-Tau1CO₂x1) equatorial θ response to abrupt quadrupling of CO₂. c) CESM1 momentum-forced (MF = Tau4CO₂x1-Tau1CO₂x1) equatorial θ response to abrupt quadrupling of CO₂. All panels are meridionally averaged from 5°S-5°N and temporally averaged from years 11-50. The 16°C isotherm from Ctrl is plotted as a solid white contour in all three panels and approximates the mean thermocline. Dotted white contours are the 16°C isotherm in the forced experiments: a) CO₂x4, b) Tau1CO₂x4, and c) Tau4CO₂x1.

332 in response to SH ETINMIP forcing (Figure S4d-f). This similarity between equatorial
 333 temperature responses was not expected a priori because abrupt 4xCO₂ is a nearly hemi-
 334 spherically symmetric forcing while ETINMIP forcing is purposefully hemispherically
 335 asymmetric. We discuss this further in Section 4.

336 The MF response, created by momentum-driven dynamics, is the cause of the promi-
 337 nent thermocline cooling seen in the FC response, qualitatively agreeing with Jiang et
 338 al. (2024b, 2025)'s assertion that winds have driven much of the observed subsurface equa-
 339 torial temperature response. This cooling is maximized within the thermocline and is
 340 a major feature across the majority of the basin. However, while weak cooling extends
 341 from surface to depth in the western Pacific, even the strong cooling in the thermocline
 342 dissipates before reaching the eastern boundary. Instead the eastern equatorial Pacific
 343 from the surface to below the thermocline features weak warming. This zonal temper-
 344 ature dipole is a feature of relaxed thermocline tilt: a shoaling of the western Pacific ther-
 345 mocline and a deepening of the eastern Pacific thermocline would respectively manifest

346 as a cooling and warming in depth space. The relaxed thermocline tilt in FC, caused by
347 westerly anomalies in equatorial wind stress from a weakened Walker Circulation (not
348 shown), agrees with Vecchi and Soden (2007)'s hypothesis that the thermocline cooling
349 response to climate change results from a decadal Bjerknes-like response to relaxed wind
350 stress. Additionally, slightly weakened STCs in our MF experiment (not shown) could
351 contribute to this cooling (Yang et al., 2009; Jiang et al., 2025). We note, however, that
352 because the MF response does not include greenhouse-gas-driven increases in subtrop-
353 ical stratification, and yet it accounts for all FC thermocline cooling, that this under-
354 standing disagrees with the arguments proposed by Luo et al. (2009, 2018) and Ju et al.
355 (2022) that increased subtropical stratification creates this cooling by either slowing the
356 STCs or advecting density-compensated anomalies.

357 Perhaps unsurprisingly, the BF response, which captures the ocean's response to
358 anomalous buoyancy fluxes from increased CO₂ radiative forcing, contributes nearly all
359 of the warming seen in FC. This includes a strong surface warming maximized in the cen-
360 tral Pacific and most of the eastern Pacific's sub-thermocline warming. Interestingly, the
361 central Pacific thermocline region in BF has a near-zero temperature response in the ex-
362 act same region where MF cools. This allows the relatively strong momentum-driven cool-
363 ing to clearly establish itself in FC. The near-zero temperature response of the BF ther-
364 mocline, which strongly resembles the full centennial response to abrupt CO₂ quadru-
365 pling seen in Heede et al. (2021), also implies a relatively long-lasting upwelling damp-
366 ing effect.

367 This BF pattern, with its near-zero thermocline response, strong surface warming,
368 and sloping sub-thermocline eastern Pacific warming, is not obviously attributable to well-
369 known dynamics. Both conventional advective (ocean-tunnel) and dynamical (wave-driven)
370 understandings of the STCs instead suggest broad thermocline warming. While Luongo
371 et al. (2023) perform an ocean mixed layer heat decomposition on the equatorial SST
372 response to NH ETINMIP forcing and attribute a certain amount of the BF SST response
373 to ocean dynamics, the specific dynamic adjustments remain unclear. As such, we turn
374 to MITgcm ocean-only simulations to explain the ocean dynamics that create the sub-
375 surface BF response to climate change forcing seen in Figure 2b.

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3.2 Recreating the Buoyancy Response

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To determine whether MITgcm is an appropriate tool with which to understand the ocean dynamics that create the buoyancy-driven mechanically-decoupled CESM1 response in Figure 2b, we first test whether MITgcm is able to effectively recreate CESM1's response at all. In the CO₂x4_BFsst OGCM simulation we add the global quasi-steady buoyancy-forced SST response to abrupt 4xCO₂ forcing (Figure 3a) to MITgcm's monthly climatological SST relaxation fields. Because we do not change any other forcing fields, the difference between this simulation and OCtrl is the ocean-only response to that SST forcing pattern.

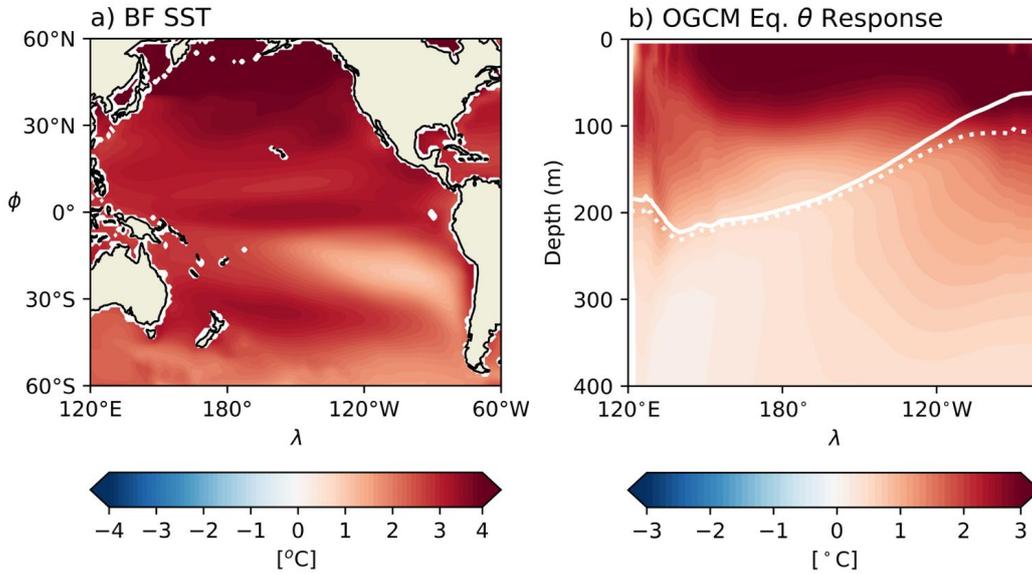


Figure 3. a) Buoyancy-forced quasi-steady (year 11-50) SST response in CESM1. b) MITgcm equatorial temperature (θ) response to BF SST forcing in panel a) averaged over years 11-30: CO₂x4_BFsst-OCtrl. The 16°C isotherm in OCtrl (solid white contour) approximates the mean thermocline and the 16°C isotherm in CO₂x4_BFsst (dotted white contour) approximates the forced thermocline.

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MITgcm does a surprisingly good job of recreating the major features of CESM1's quasi-steady buoyancy-forced subsurface equatorial temperature response (c.f. Figures 2b and 3b, Pearson pattern correlation of 0.86). The MITgcm temperature response features the strong, relatively zonally symmetric near-surface warming, a warming minimum in the central-western Pacific thermocline, and sloping sub-thermocline warming

390 in the eastern Pacific. There are some notable differences between the two patterns, most
 391 obviously that the near-zero thermocline warming so obvious in CESM1 is deeper, more
 392 westward, and more diffuse in MITgcm. The eastern Pacific sub-thermocline warming
 393 is also weaker in MITgcm. However, MITgcm and CESM1’s ocean component are dif-
 394 ferent models with differences in mean state, parameterizations, and resolution. Given
 395 this reality, we consider the otherwise substantial agreement between Figures 2b and 3b
 396 to be promising, and we conclude that these ocean-only simulations are a reasonable di-
 397 agnostic tool for understanding mechanically-decoupled simulations.

398 Having shown that the MITgcm response to the full BF SST field reasonably recre-
 399 ates the CESM1 response, we now ask whether we can decompose this full response fur-
 400 ther, namely into the response to SST forcing from different geographic regions. This
 401 question emerges directly from the canonical advective ocean tunnel understanding of
 402 the non-wind-driven STC response to climate change (e.g., Clement et al., 1996; Luo et
 403 al., 2009, 2017, 2018; Heede & Fedorov, 2021; Ju et al., 2022), which suggests that warm
 404 subtropical surface waters subduct in the eastern half of the subtropical gyre, are car-
 405 ried by mean advection to the western boundary, penetrate into the tropics via low lat-
 406 itude western boundary currents, and eventually warm the thermocline. Because this
 407 understanding suggests that some portion of the equatorial temperature response is en-
 408 tirely remotely-driven, we run two OGCM simulations to determine whether we can un-
 409 derstand the full response to BF SST forcing as the sum of patterns created by remote
 410 and local SST forcing. We do this by regionally partitioning the full BF SST forcing field
 411 in Figure 3a into remote extratropical (ET: Figure 4a) and local equatorial (EQ: Fig-
 412 ure 4b) SST forcing fields.

413 The sum of the remote (Figure 4c) and local (Figure 4d) responses almost perfectly
 414 recreates the full field response in Figure 3b (Figure S3b, Pearson pattern correlation of
 415 0.98). However, while these patterns sum to the full response, they differ substantially
 416 and clearly represent different oceanic dynamics. We explore the ocean adjustments that
 417 create the remote and local response in the following two subsections.

418 ***3.2.1 Adjustments to Remote Buoyancy Forcing***

419 We first consider the equatorial temperature response to extratropical-only BF SST
 420 forcing (Figure 4c). The equatorial Pacific subsurface warms in response to this remote

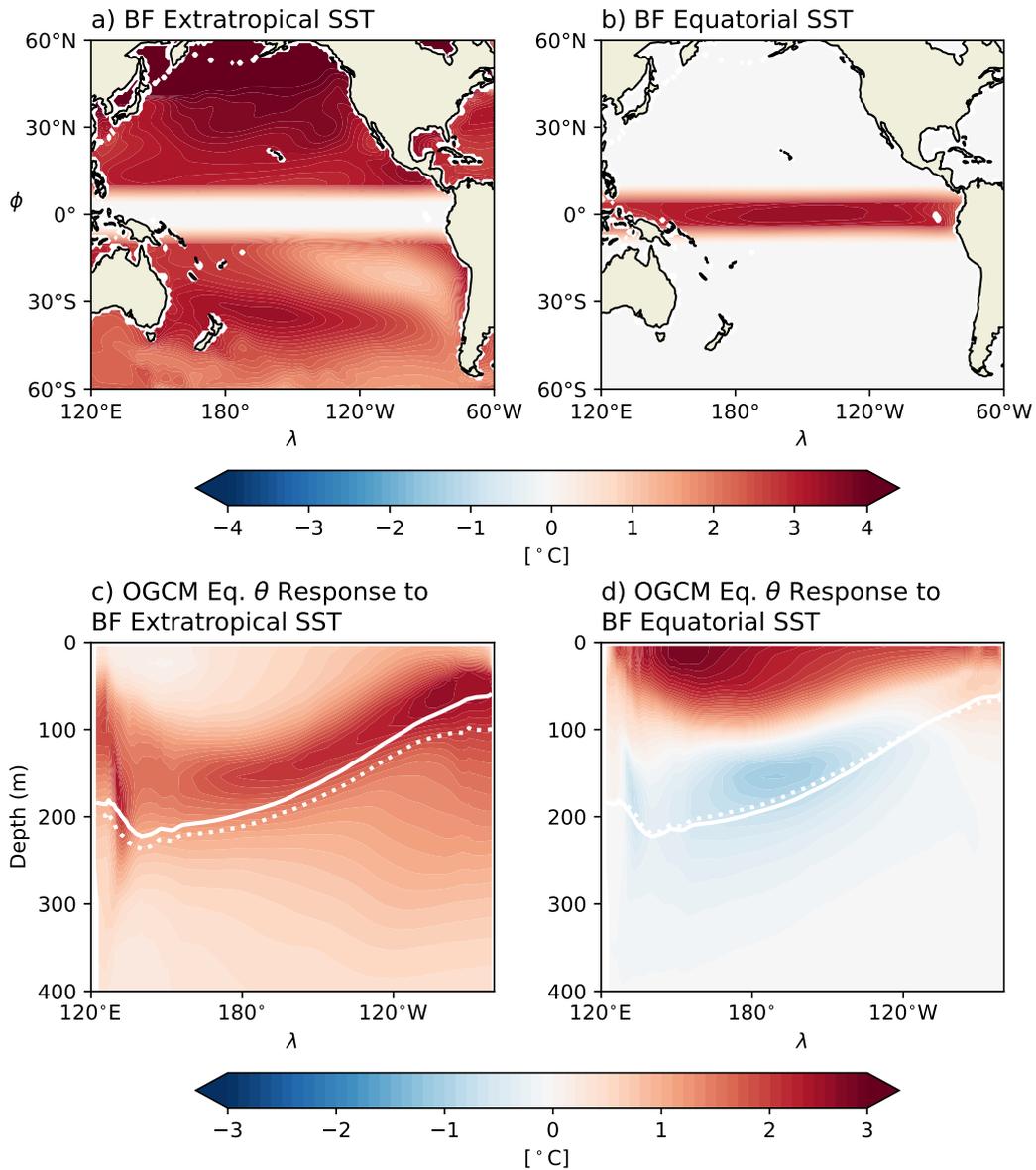


Figure 4. a) Remote, extratropical component of BF SST response in Figure 3a. b) Local, equatorial component of BF SST response in Figure 3a. c) Equatorial temperature (θ) response to remote SST forcing in panel a): $\text{CO}_2 \times 4 \text{BFsstET-OCtrl}$. d) Equatorial θ response to local SST forcing in panel b): $\text{CO}_2 \times 4 \text{BFsstEQ-OCtrl}$. The 16°C isotherm in OCtrl (solid white contour in panels c & d) approximates the mean thermocline and the 16°C isotherm in $\text{CO}_2 \times 4 \text{BFsstET}$ and $\text{CO}_2 \times 4 \text{BFsstEQ}$ (dotted white contours in panels c & d, respectively) approximates the perturbed thermocline.

421 forcing, with no evidence of cooling. The warming is maximized within the thermocline
 422 throughout the entire basin. Despite strong surface relaxation to unperturbed SSTs, the
 423 strong near-surface thermocline warming in the eastern equatorial Pacific extends to the
 424 surface. In addition, it is clear that the sloping sub-thermocline warming in the eastern
 425 Pacific noted in the FC and BF CESM1 responses above is caused by this remote ad-
 426 justment.

427 This coherent warming of the equatorial thermocline in response to remote SST
 428 forcing in both hemispheres is strikingly similar to the equatorial temperature response
 429 to a +2°C SST anomaly in the northeast Pacific stratocumulus deck as presented in Luongo
 430 et al. (2025)’s NEPac2CWarm simulation (c.f. Figures 4c and 5b, Pearson pattern cor-
 431 relation of 0.94). In that study we used MITgcm to investigate how the tropical ocean
 432 responded to subtropical surface cooling. We showed that both circulation adjustments
 433 driven by baroclinic planetary waves ($v'\bar{\theta}$) and mean advection of temperature anoma-
 434 lies within the subtropical gyre and low latitude western boundary currents ($\bar{v}\theta'$) com-
 435 municated subtropical cooling to the tropics within about a decade. At the equator, an
 436 equatorial Kelvin wave coherently heaved the thermocline as it traveled eastward. Upon
 437 hitting the eastern boundary, this wave signal radiated poleward in both hemispheres
 438 as coastal Kelvin waves, which then proceed to adjust stratification in the eastern basin
 439 by shedding westward-propagating Rossby waves. Although we primarily focused on sub-
 440 tropical cooling in Luongo et al. (2025), the NEPac2CWarm response presented in Fig-
 441 ure 5 demonstrates that the equatorial temperature response to subtropical warming is
 442 simply the opposite of its response to subtropical cooling. It is also relevant to note that
 443 Luongo et al. (2025) showed that both NH and SH subtropical forcing led to similar equa-
 444 torial response patterns due to the symmetric nature of the equatorial Kelvin wave ad-
 445 justment.

446 In the case of CO₂x4_BFsstET, the dynamics clarified in Luongo et al. (2025) sug-
 447 gest that strong subtropical warming present in both NH and SH (Figure 4a) contribute
 448 to the warming of the equatorial thermocline (Figure 4c). This warming occurs through
 449 both mean advection of warm anomalies, as in the canonical ocean tunnel understand-
 450 ing, but also due to the coherent deepening of the equatorial thermocline via a down-
 451 welling Kelvin wave excited by the subtropical gyres’ baroclinic response to anomalous
 452 surface warming. This dynamical adjustment is in-turn responsible for the eastern Pa-
 453 cific’s sloping sub-thermocline warming, a slow stratification adjustment to the heaved

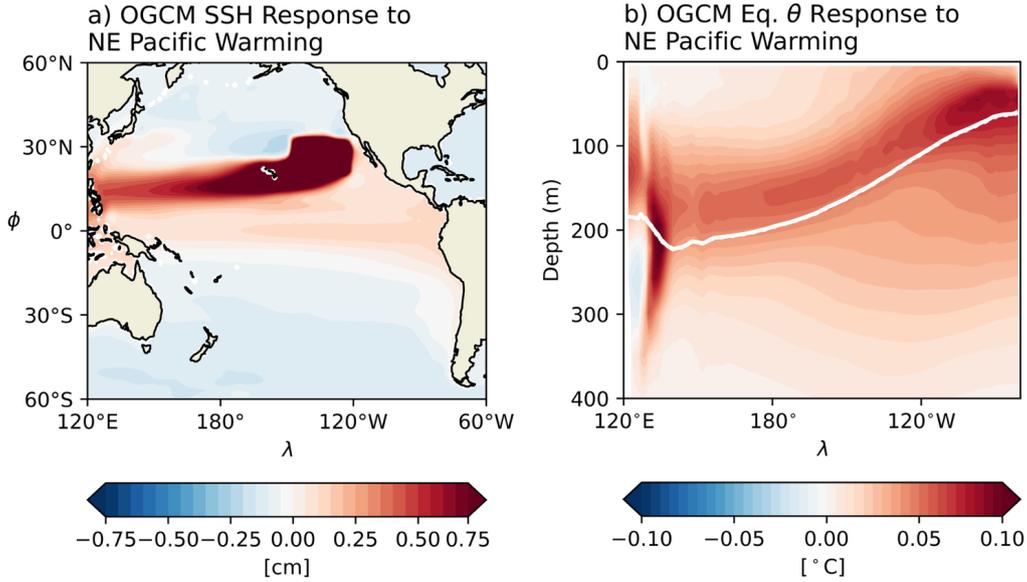


Figure 5. a) Sea surface height (SSH) response (NEPac2CWarm-OCtrl) to Northeast Pacific warming patch from Luongo et al. (2025). b) Equatorial temperature (θ) response to NEPac2CWarm simulation. The 16°C isotherm in OCtrl (solid white contour) approximates the mean thermocline and the 16°C isotherm in NEPac2CWarm (dotted white contour) approximates the forced thermocline.

454 thermocline. As such, while the difference in magnitude between Figures 4c and 5b
 455 results from the much stronger forcing from CO_2 quadrupling in $\text{CO}_2\text{x4_BFsstET}$ com-
 456 pared to the relatively small patch of $+2^{\circ}\text{C}$ warming in NEPac2CWarm, the strong sim-
 457 ilarity in pattern re-emphasizes the importance of the STC as a dynamic mechanism to
 458 communicate subtropical warming to the equatorial thermocline.

459 *3.2.2 Adjustments to Local Buoyancy Forcing*

460 The response of the subsurface equatorial Pacific to the local buoyancy-forced com-
 461 ponent of climate change forcing (Figure 4d) unsurprisingly features strong near-surface
 462 warming which is directly tied to the applied tropical SST forcing (Figure 4b). However,
 463 perhaps unexpectedly, the local response features substantial cooling across much of the
 464 thermocline that underlies the strong near-surface warming. In the western Pacific, where
 465 the mean thermocline and anomalous near-surface warming signal are deepest, this ther-
 466 mocline cooling response extends from approximately 100-400m. As the mean thermo-

467 cline tilts upward to the east, this cooling signal gets shallower and thinner until the tem-
 468 perature anomaly switches signs to warming around 110°W. The thermocline then re-
 469 mains anomalously warm all the way to the eastern boundary.

470 Luo et al. (2018) used an OGCM to explore the response of the equatorial ther-
 471 mocline to a uniform tropical warming of 3.2°C. They find a pattern of near-surface warm-
 472 ing and thermocline cooling that is similar to our response to local BF. In that work,
 473 Luo et al. (2018) suggested that this temperature response was a local baroclinic adjust-
 474 ment to surface warming: as near-surface stratification increases in response to surface
 475 warming, turbulent downward mixing of that heat decreases and creates a cooling sig-
 476 nal (Yang et al., 2009). Luo et al. (2018), therefore, would attribute much of the cool-
 477 ing in Figure 4d to be a signal of reduced mixing. We note, however, that this mecha-
 478 nism does not explain the zonal temperature dipole clearly seen in our thermocline re-
 479 sponse to local BF: as discussed above, a zonal dipole instead implies a thermocline tilt
 480 and suggests the need to consider zonal gradients.

481 To determine whether the local response can instead be understood in terms of in-
 482 viscid dynamics, we model the equatorial ocean as a simple 1.5-layer reduced gravity sys-
 483 tem. In this simplified understanding, the lower level flow is negligible compared to up-
 484 per level flow ($\vec{u}_1, \vec{v}_1 \gg \vec{u}_2, \vec{v}_2 = 0$), the layers are coupled by their density differences
 485 via the reduced gravity parameter [$g' \equiv g(\rho_2 - \rho_1)/\rho_2$], and the interface depth, $h =$
 486 $\eta + H$, is thermocline depth defined positive downward and as a sum of interface dis-
 487 placement η and mean thermocline depth H . Ignoring dissipation terms, the steady, lin-
 488 ear equatorial zonal momentum equation in conservative flux form is

$$0 = -\left(g' \frac{h^2}{2}\right)_x + \frac{\tau^x}{\rho_0}, \quad (2)$$

489 where τ^x is zonal wind stress, ρ_0 is a constant reference density, and the x subscript rep-
 490 represents a zonal derivative. If we now instead consider the non-conservative velocity form
 491 of Equation 2 linearized about H ,

$$0 = -g' \eta_x + \frac{\tau^x}{\rho_0 H}, \quad (3)$$

492 and we consider a uniform Δ climate change forcing without changes in τ^x (as is the case
 493 in our OGCM simulations), the balance becomes

$$\Delta g' \overline{\eta_x} = -\overline{g'} \Delta \eta_x . \quad (4)$$

Equation 4 demonstrates, without making any assumptions about the eastern bound-
 ary, that the product of the perturbed RG and the mean zonal gradient of interface dis-
 placement must be balanced by the product of the mean RG and a perturbed zonal gra-
 dient of interface displacement. In the case of climate change driven warming, because
 $\Delta g' > 0$ we expect $\Delta \eta_x > 0$. Put another way, an increase in stratification should re-
 duce the tilt of the thermocline even with no change in winds, in turn leading to west-
 ern thermocline cooling and eastern thermocline warming.

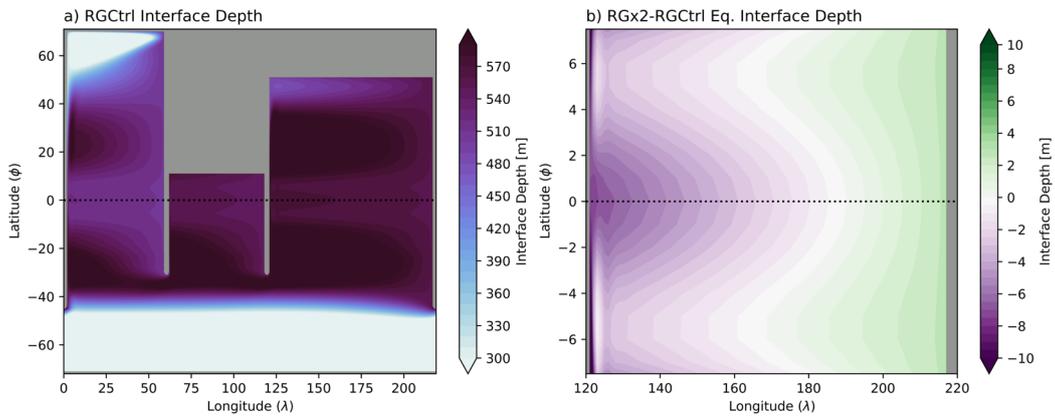


Figure 6. a) Interface depth in idealized global RG model after 20 years of spin-up in RGCtrl. b) Equatorial interface depth response to RGx2-RGCtrl’s doubling of reduced gravity at year 10.

We test this tilt hypothesis with two simulations using Sun and Thompson (2020)’s
 idealized global ocean reduced gravity model (Figure 6a), comparing a simulation where
 the reduced gravity parameter is doubled (RGx2) to one where it isn’t (RGCtrl). While
 a doubling of stratification is a relatively large forcing, it is on the same order as the near-
 surface equatorial Pacific increase in stratification observed in the BF response (not shown).
 Figure 6b shows the η response in the equatorial Pacific 10 years after RG is doubled.
 We see that the western equatorial Pacific interface displacement decreases (shoals), while
 the eastern equatorial Pacific interface displacement increases (deepens). This western
 Pacific shoaling and eastern Pacific deepening corresponds to a relaxation in thermocline
 tilt, or a western thermocline cooling and an eastern thermocline warming. We use this
 newly gained physical intuition to explain the thermocline temperature dipole in Fig-

512 ure 4d, which then adds to the coherent thermocline shoaling pointed out by Luo et al.
513 (2018).

514 In summary, we conclude that the local response of the equatorial thermocline to
515 the equatorial buoyancy-driven SST response to climate change consists of two responses
516 to the increase in near-surface stratification: both a coherent thermocline shoaling and
517 a decrease in thermocline tilt. This latter point questions the oft-held view that a tilted
518 thermocline is necessarily tied to a change in zonal winds.

519 4 Discussion

520 4.1 Linearity of the Equatorial Thermocline's Response

521 We use a hierarchy of models to show that the full subsurface temperature response
522 of the equatorial thermocline to greenhouse gas forcing (θ_{FC} : Figure 2a) can be recov-
523 ered as a relatively simple linear combination of independent ocean dynamical responses.
524 As in Luongo et al. (2023), we use a mechanically decoupled model to show that θ_{FC}
525 is the sum of the wind stress-driven response (θ_{MF} : Figure 2c) and the buoyancy-driven
526 response (θ_{BF} : Figure 2b). We then use OGCM simulations to show that θ_{BF} can be
527 linearly partitioned into a sum of forced responses from different geographic regions of
528 SST forcing: a remote, extratropically-driven response ($\theta_{BF,remote}$: Figure 4c) and a lo-
529 cal, equatorially-driven response ($\theta_{BF,local}$: Figure 4d). The remote response represents
530 the dynamically and thermodynamically-driven changes of the thermocline in response
531 to subtropical temperature forcing, as outlined in Luongo et al. (2025). We then use a
532 RG model to show that the local response is consistent with a response to surface strat-
533 ification that includes both shoaling and reduced thermocline tilt. Putting this all to-
534 gether, we present this linear understanding as

$$\begin{aligned} \theta_{FC} &= \theta_{MF} + \theta_{BF} = \theta_{MF} + \theta_{BF,remote} + \theta_{BF,local} \\ &= \alpha * MF(\tau^x, \tau^y) + \beta * BF_{remote}(NH\ SST, SH\ SST) + \gamma * BF_{local}(Eq.\ SST) . \end{aligned} \quad (5)$$

535 Equation 5 communicates that the full response of the modeled tropical Pacific sub-
536 surface temperature response to greenhouse gas forcing is a linear combination of ocean
537 dynamics driven by momentum from zonal and meridional wind stress, buoyancy from
538 NH and SH subtropical SST forcing, and buoyancy from local equatorial SST forcing,

539 with α , β , and γ as scaling coefficients. This is the primary result of our study. Because
 540 our FC CESM1 response to abrupt quadrupling of CO₂ strongly resembles both the 1958-
 541 2020 observed (Figure 1a) and multi-model mean (Figure 1b) response to realistic, his-
 542 torical climate change forcing, we conclude that this simple linear understanding gained
 543 from a hierarchy of idealized modeling simulations is a powerfully relevant and applica-
 544 ble tool with which to understand realistic climate change.

545 While this understanding is simple at face value, it in fact suggests that the equa-
 546 torial temperature response to even just steady, idealized greenhouse gas forcing is more
 547 complex than previously understood. We see that the full response actually contains many
 548 of the dynamics previously suggested in the literature (e.g. Clement et al., 1996; Sea-
 549 ger & Murtugudde, 1997; Vecchi & Soden, 2007; Luo et al., 2009, 2018; Heede et al., 2020,
 550 2021; Watanabe et al., 2024). Nevertheless, it is only through understanding the com-
 551 bination of these dynamics that we can critically update our theoretical picture for how
 552 the equatorial Pacific will respond to climate change.

553 **4.2 Reconstructing Long and Short-Term Climate Change Responses**

554 As suggested by Equation 5, we seek to reconstruct the observed equatorial sub-
 555 surface temperature trends from 1958-2020 (Figure 1a) as a linear combination of the
 556 MF and remote and local BF patterns. We use a depth-weighted ordinary least squares
 557 multilinear regression to determine combination coefficients. Our reconstruction of the
 558 1958-2020 trend (Figure 7a) shows strong agreement with the observational composite
 559 (Figure 1a), with a Pearson pattern correlation of 0.86. We are able to explain much of
 560 both the models' and observations' long-term subsurface temperature response through
 561 our linear understanding of oceanic dynamics. Interestingly, we note that while the re-
 562 gression coefficients for the MF and local BF responses are positive (meaning they are
 563 in line with positive greenhouse gas forcing), the remote BF response is slightly nega-
 564 tive (Figure 7c). This suggests that subtropical temperature forcing has not majorly im-
 565 pacted the equatorial thermocline in this period, and, if it has, it's been a cooling sig-
 566 nal. This could hypothetically be due to a communication of the observed subtropical
 567 Southeast Pacific surface cooling (Wills et al., 2022) via the subtropical cells (Luongo
 568 et al., 2025).

569 While the multi-model mean trend is obviously similar to the observed trend over
 570 1958-2020, the observed 1979-2020 trend is notably distinct from the modeled trend over

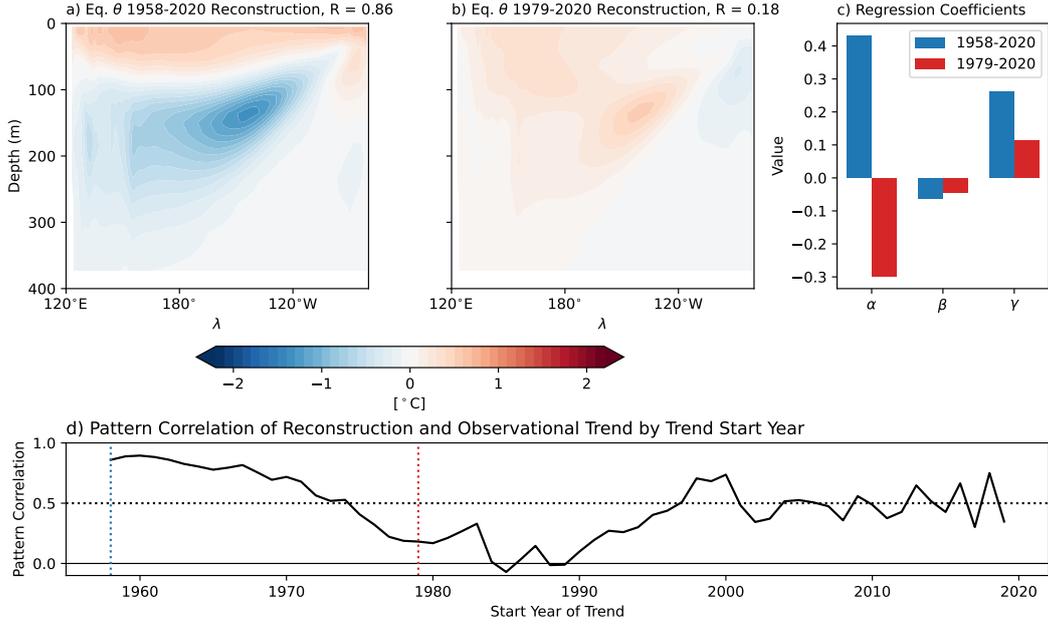


Figure 7. a) 1958-2020 reconstruction of Figure 1a observational trend using α *MF pattern + β *remote BF pattern + γ *local BF pattern as in Equation 5. b) As in a), but for 1979-2020 observational trend of Figure 1c. c) Regression coefficients for linear combinations in a) and b). d) Pearson pattern correlation between regression-estimated reconstruction and observed trend as a function of trend start year. The dotted blue line highlights 1958, the dotted red line highlights 1979, and the dotted black line is at a correlation value of 0.5.

571 that same period (c.f. Figure 1c & d). Indeed, our multilinear regression method can not
 572 successfully reconstruct this more recent period’s observed temperature trend (Figure
 573 7b, Pearson correlation 0.18). In fact, the regression coefficients suggest that the best
 574 way to reconstruct the strong subsurface western Pacific warming in the 1979-2020 trend
 575 is via a momentum-forced response opposite to that caused by CO₂ forcing (Figure 7c).
 576 Although the understanding gained in Equation 5 works for longer-term temperature trends,
 577 it breaks down for the more recent past.

578 The disagreement between models’ and observations’ SST trends over this period
 579 has been the focus of many recent studies (e.g., Wills et al., 2022; Watanabe et al., 2024);
 580 our result underlines this disagreement further. We highlight two implications of this re-
 581 sult. First, while Seager et al. (2019) and Jiang et al. (2025) emphasize how differences
 582 between observations and model mean states could impact SSTs, and although GCM
 583 mean states are certainly biased, the GCMs are clearly able to recreate the longer-term

584 subsurface temperature response to historical forcing, as long as the trend starts before
585 ~ 1975 (Figure 1e). This conclusion, combined with recent results from Dhame et al. (2025),
586 who showed that while higher resolution simulations reduce mean state biases they do
587 not necessarily simulate temperature trends better, suggests that mean state biases are
588 not the only cause of disagreement in trends over the past 40 years. This suggests that
589 despite equatorial mean state biases our climate models are not hopelessly unfit for the
590 task of climate change projections.

591 Second, this result suggests an important role for internal variability in recent equa-
592 torial Pacific thermocline temperature trends: we simply cannot recreate the observed
593 pattern of subsurface temperature since the late 1970s with our linear understanding of
594 the ocean's response to different quasi-steady forcings. Even making the MF coefficient
595 negative, consistent with an observed strengthening Walker circulation (L'Heureux et
596 al., 2013), is not enough to recreate the observed dipole in the equatorial thermocline.
597 Because our patterns are quasi-steady, internal variability is an obvious culprit. Figure
598 7d shows how the pattern correlation between our regression-estimated reconstruction
599 and observational trend changes as a function of start year of the trend. Trends which
600 start before 1975 are skillfully reconstructed from our linear understanding, presumably
601 because enough cycles of internal variability have been averaged out to clarify the re-
602 sponse. The linear reconstruction specifically fails, however, when attempting to recon-
603 struct trends that begin between 1975-2000, a period when models and observations dis-
604 agree the most (Figure 1e) and which also happens to coincide with several strong El
605 Niño events.

606 Nevertheless, we can definitively say the dynamics we have emphasized here, decadal
607 momentum-driven thermocline tilt, subtropical cell adjustment, and equatorial stratification-
608 induced thermocline shoaling and tilt, contribute less to the 1979-2020 trend than has
609 been commonly hypothesized [e.g., Figure 4 schematic of Watanabe et al. (2024)]. We
610 hypothesize that recent internally-driven climate variability, such as ENSO or tropical
611 Pacific decadal variability (Capotondi et al., 2023), which is not included in our pattern
612 reconstructions, has critically crafted observed subsurface equatorial Pacific temperature
613 trends over the past 40 years. This conclusion is in line with Jiang et al. (2024a), who
614 emphasize the role of the Interdecadal Pacific Oscillation on SST patterns over that same
615 period.

4.3 Implications for the Canonical Ocean Dynamical Thermostat

The simplicity of this linear understanding allows us to interrogate the canonical view of the ocean-tunnel-mediated ocean dynamical thermostat as a mechanism for understanding recent equatorial Pacific climate change. In this view, which largely rests on a mean advective understanding of the STCs ($\bar{v}\theta'$), anomalous subtropical warming is communicated to the equatorial thermocline after some lag to erode the relative cooling initially created by continued mean upwelling of unperturbed waters. As such, the conventional view of the ocean dynamical thermostat mechanism has been limited to a transient phenomenon. This understanding has been used to suggest that while the tropical Pacific's response to climate change may start as La Niña-like, as in recent observations, it will eventually transition to El Niño-like as suggested by the vast majority of model projections (Heede et al., 2020, 2021; Watanabe et al., 2024).

Our remote, buoyancy-driven response indeed shows that subtropical SST warming in the extratropics coherently warms the thermocline. While Luongo et al. (2025) show that this pattern is best understood as having been created by wave dynamics, the basic understanding that subtropical SSTs warm the equatorial thermocline via the STCs holds true. In the framework of the upwelling damping effect of mean vertical advection on equatorial SSTs ($-\bar{w}\theta'_z \equiv -\bar{w}\frac{\theta'_s - \theta'_e}{H_e}$; e.g., Xie et al., 2010), where the s and e subscripts are respectively surface and entrainment levels and H_e is the depth of the entrainment level to the surface, initially $\theta'_e < \theta'_s$ and so mean upwelling cools, but eventually $\theta'_e > \theta'_s$ and mean upwelling warms.

However, our remote and local MITgcm simulations show that this is a delicate balance: the remotely-driven thermocline warming (Figure 4c) is entirely canceled out (Figure 2b & 3b) by the locally-driven thermocline cooling (Figure 4d). In the case of our simulations, subtropical warming only works to erode the local cooling from thermocline shoaling and decreased thermocline tilt such that the final, steady thermocline temperature response is near-zero ($\theta'_e \approx 0$) and there is no sign change in $-\bar{w}\theta'_z$ with time. This implies a long-lasting upwelling damping effect. Our model simulations, therefore, suggest that the canonical view of the thermostat necessarily leading to a transient response (as has been used to explain recent observations) is misleading. Rather, we emphasize that the relative ratio of tropically-driven cooling to extratropically-driven warming is critical to understand timescales associated with the buoyancy-driven response. In our

648 case, which in turn resembles the near-steady, centennial response to abrupt quadrupling
649 of CO₂ (Heede et al., 2021), remote warming simply cancels out local cooling such that
650 there is no major warming of the thermocline from SST forcing at all (Figures 2b & 3b).
651 If instead, however, the extratropically-driven warming was much larger than the tropically-
652 driven cooling we might expect a correlation between subtropical to tropical meridional
653 SST gradients and the tropical zonal SST gradient (Burls & Fedorov, 2014). Considered
654 together, we instead hypothesize that the more likely driver of transience in the surface
655 response results from a change in the momentum-driven pattern or other atmospheric
656 pathways.

657 **4.4 Symmetry of Equatorial Responses**

658 A final surprising detail of this study is the immutability of the equatorial Pacific
659 subsurface temperature response regardless of forcing geography. This is best seen by
660 comparing the FC, BF, and MF responses to abrupt quadrupling of CO₂ (Figure 2) to
661 those same responses to NH and SH ETINMIP forcing (Figure S4). Despite the fact that
662 CO₂ forcing is primarily equatorially symmetric and ETINMIP forcing is purposefully
663 equatorially asymmetric, the response patterns are effectively the same (with an oppo-
664 site sign). From the perspective of the equatorial Pacific subsurface, it would be diffi-
665 cult to immediately tell the difference between greenhouse gas warming and a hypothet-
666 ical extratropical warming (e.g., Tseng et al., 2023). This similarity extends to the lin-
667 ear partitioning of the buoyancy-driven response into remote and local forcing (Figure
668 S5). The equatorially symmetric nature of the equatorial thermocline’s response to sub-
669 tropical forcing (Luongo et al., 2025) creates the same remote response pattern as if both
670 hemispheres’ subtropics were forced. Because the local response just depends on an in-
671 crease or decrease in surface stratification, the equatorial response to local forcing looks
672 effectively the same.

673 This understanding raises two interesting points. First, it highlights that hemispheric
674 asymmetries in meridional forcing, crucial to the zonal-mean energetic framework through
675 which we understand ITCZ shifts (Kang et al., 2008) and cross-equatorial ocean heat
676 transport (Luongo et al., 2022), do not lead to appreciably different equatorial temper-
677 ature responses. Despite the fact that ETINMIP forcing drives strong cross-equatorial
678 responses, the equatorial thermocline simply cares if the forcing causes large-scale warm-
679 ing or cooling. Second, the equatorial thermocline’s response to industrial aerosol forc-

680 ing (similar to NH ETINMIP forcing) would not lead to an independent temperature
681 pattern from greenhouse gas forcing. Put another way, NH aerosol forcing would sim-
682 ply modulate the tropical Pacific’s response to greenhouse gas forcing rather than cre-
683 ate a fundamentally different pattern.

684 5 Conclusions

685 In this study we have used a series of climate modeling simulations of varied com-
686 plexity to understand the equatorial thermocline response to climate change. We first
687 show that a multi-model mean of 11 large ensembles reasonably captures the observed
688 1958-2020 subsurface equatorial temperature trend, and that CESM1’s 11-50 year av-
689 erage response to abrupt quadrupling of CO₂ is an appropriate tool with which to un-
690 derstand the models’ response to realistic, historical forcing. We then decompose the full
691 equatorial thermocline response into a response due to buoyancy forcing alone and mo-
692 mentum forcing alone, the latter of which drives thermocline cooling. We use an ocean-
693 only GCM with anomalous SST forcing to further decompose that buoyancy-forced com-
694 ponent, and we demonstrate that the response due to extratropical SST forcing and trop-
695 ical SST forcing linearly combine to recreate the full field buoyancy-forced response. The
696 remote, extratropically-driven response leads to a coherent thermocline warming through
697 dynamic and thermodynamic pathways. The increase in near-surface stratification in the
698 local, tropically-driven response leads to both a shoaling thermocline and a relaxation
699 of thermocline tilt. Our primary finding is that a simple linear combination of these ad-
700 justments, i) momentum-driven, ii) remote buoyancy-driven, and iii) local buoyancy-driven,
701 skillfully explains both the long-term 1958-2020 modeled and observed responses. We
702 can attribute certain features of the pattern to certain dynamics: we agree with Vecchi
703 and Soden (2007) and Jiang et al. (2025)’s suggestion that the thermocline cooling re-
704 sponse to global warming results from momentum-driven dynamics. However, this dy-
705 namical understanding does not explain more recent trends (e.g., the 1979-2020 response),
706 suggesting that this period was strongly affected by internal variability.

707 Our results emphasize that the subsurface equatorial Pacific temperature response
708 to climate change is a highly linear system. This linearity is powerful. It allows us to test
709 long-held theoretical understandings, such as how subtropical warming will affect the tran-
710 sient adjustment of the tropical thermocline or that changes in zonal wind stress are nec-
711 essary for a thermocline tilt. While this study does not answer what has caused recent

712 subsurface mismatches between models and observations or whether models are miss-
 713 ing a hypothetical forcing that might explain that mismatch, we demonstrate that model
 714 mean states are not so irreparably biased that we cannot learn from them. Instead, these
 715 models clarify the specific patterns created by commonly referenced ocean dynamic ad-
 716 justments. In a practical sense, we also outline a clear model hierarchy, fully-coupled,
 717 mechanically-decoupled, ocean-only, and reduced gravity, which could be potentially lever-
 718 aged to comprehend other coupled climate responses.

719 **6 Open Research**

720 The climate model output and python RG model used in this study will be made
 721 freely available upon study publication by the corresponding author.

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Supporting Information for “Explaining the Equatorial Pacific Thermocline Response to Climate Change with a Model Hierarchy”

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Contents of this file Table S1 and Figures S1-S5.

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Fully-coupled Simulations			
Simulation Name	CO ₂ Forcing	Wind Stress	ETINMIP Forcing
ETINMIPNH	280ppm	Freely evolving	45°N–65°N
ETINMIPSH	280ppm	Freely evolving	45°S–65°S
Mechanically-decoupled Simulations			
Simulation Name	CO ₂ Forcing	Wind Stress	ETINMIP Forcing
Tau.1.S.NH	280ppm	Ctrl	45°N–65°N
Tau.NH.S.1	280ppm	ETINMIPNH	n/a
Tau.1.S.SH	280ppm	Ctrl	45°S–65°S
Tau.SH.S.1	280ppm	ETINMIPSH	n/a
Ocean-only Simulations			
Simulation Name	SST Forcing Perturbation	SST Forcing Bounds	
ETINMIPNH_BFsst	Tau.1.S.NH-Tau1CO ₂ x1	90°S-90°N	
ETINMIPNH_BFsstET	Tau.1.S.NH-Tau1CO ₂ x1	90°S-6°S, 6°N-90°N	
ETINMIPNH_BFsstEQ	Tau.1.S.NH-Tau1CO ₂ x1	10°S-10°N	
ETINMIPSH_BFsst	Tau.1.S.SH-Tau1CO ₂ x1	90°S-90°N	
ETINMIPSH_BFsstET	Tau.1.S.SH-Tau1CO ₂ x1	90°S-6°S, 6°N-90°N	
ETINMIPSH_BFsstEQ	Tau.1.S.SH-Tau1CO ₂ x1	10°S-10°N	

Table S1. Details of fully coupled, mechanically-decoupled, and ocean-only simulations using top-of-atmosphere hemispherically asymmetric extratropical forcing from the Extratropical-Tropical Interaction Model Intercomparison Project (ETINMIP: Kang et al., 2019). The Ctrl and Tau1CO₂x1 simulations are described in Table 1 of the main text.

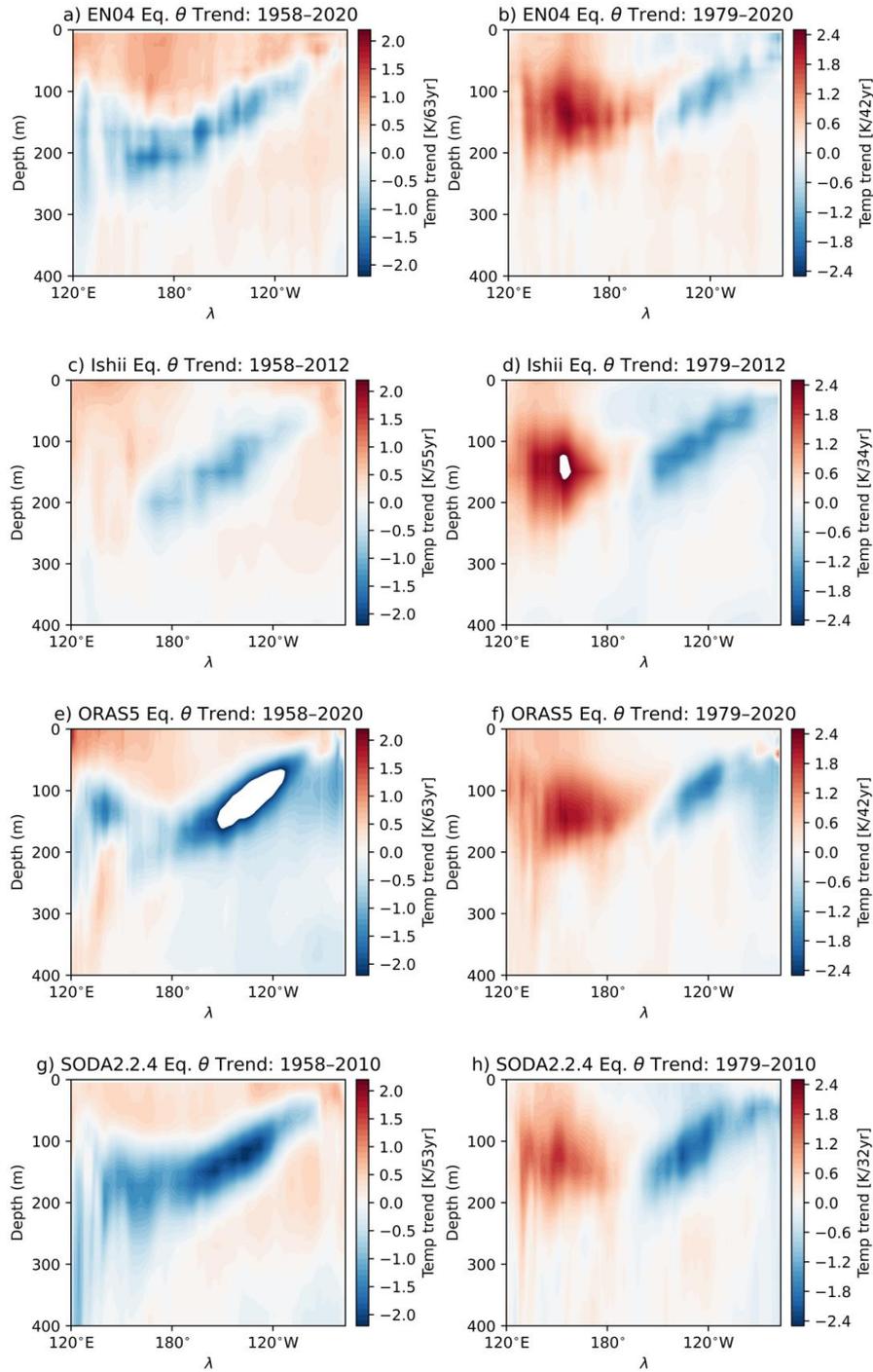


Figure S1. Equatorial subsurface temperature (θ) trend for EN04 observational product (top row, Good et al., 2013), Ishii observational product (second row, Ishii & Kimoto, 2009), ORAS5 ocean reanalysis (third row, Zuo et al., 2019), and SODA2.2.4 ocean reanalysis (fourth row, Carton & Giese, 2008) While time periods depend on specific data source, longer-term trends are in the left column and shorter-term trends in the right column.

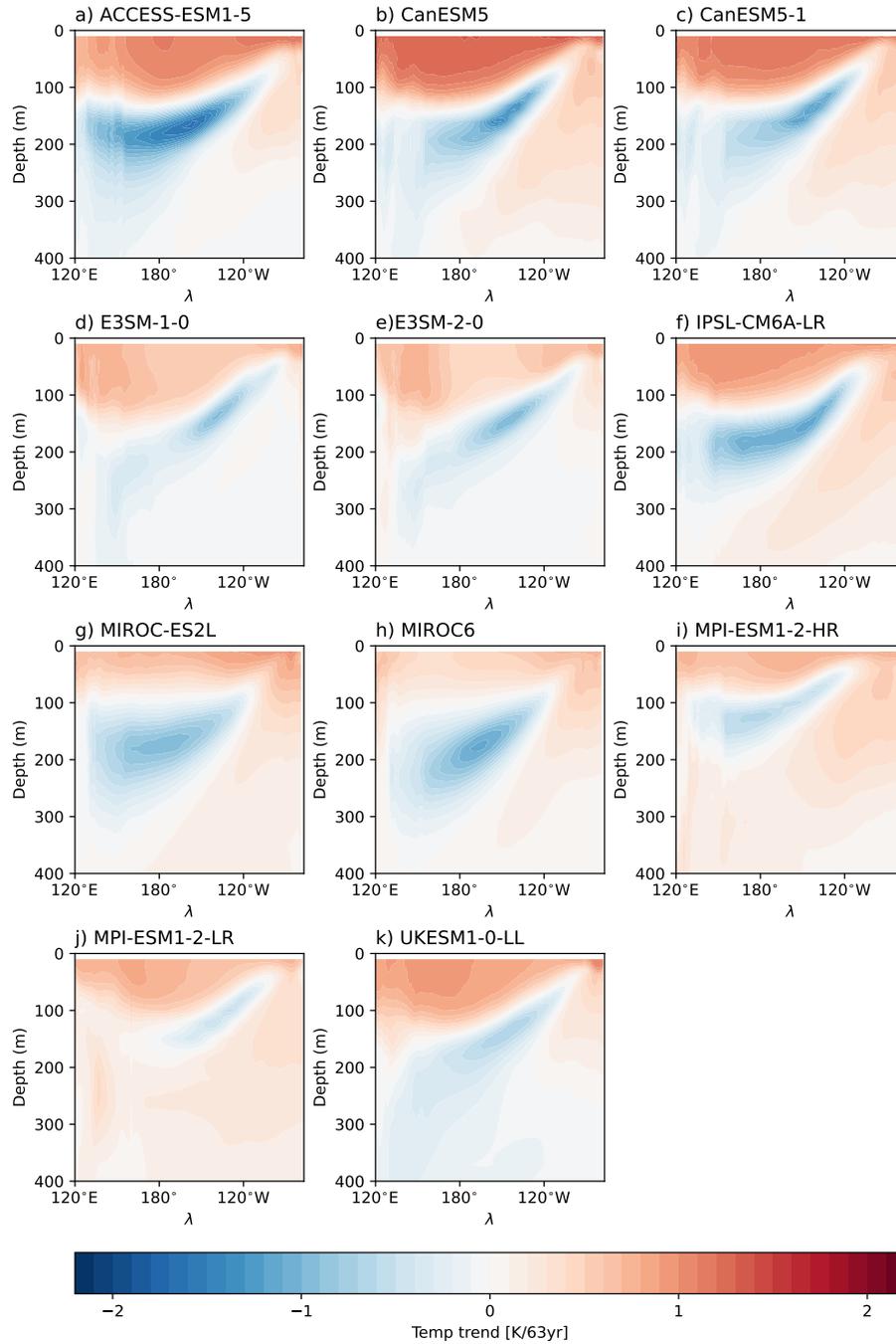


Figure S2. Ensemble mean 1958-2020 equatorial subsurface temperature (θ) trend for a) ACCESS-ESM1-5, b) CanESM5, c) CanESM5-1, d) E3SM-1-0, e) E3SM-2-0, f) IPSL-CM6A-LR, g) MIROC-ES2L, h) MIROC6, i) MPI-ESM1-2-HR, j) MPI-ESM1-2-LR, and k) UKESM1-0-LL large ensembles as selected by Jiang et al. (2025). Ensembles are forced by historical forcing from 1958-2014 and from 2015-2020 by the Shared Socioeconomic Pathway 3-7.5 scenario.

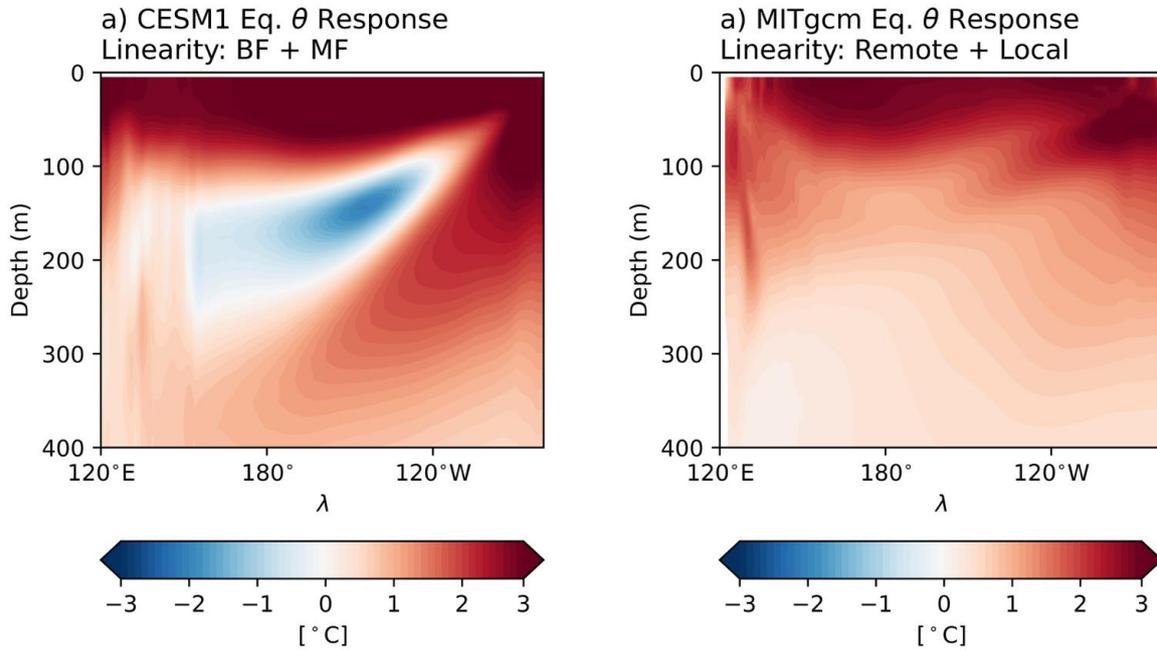


Figure S3. a) Linearity of CESM1 equatorial temperature (θ) response to abrupt quadrupling of CO_2 : Sum of buoyancy-forced (BF = $\text{Tau1CO}_2\text{x4}-\text{Tau1CO}_2\text{x1}$, Figure 2b of the main text) response and momentum-forced (MF = $\text{Tau4CO}_2\text{x1}-\text{Tau1CO}_2\text{x1}$, Figure 2c of the main text) response. Compare this sum with the FC response (Figure 2a of the main text). b) Linearity of the MITgcm equatorial θ response to BF SST forcing: Sum of response to remote SST forcing ($\text{CO}_2\text{x4_BFsstET-OCtrl}$, Figure 4c of the main text) and response to local SST forcing ($\text{CO}_2\text{x4_BFsstEQ-OCtrl}$, Figure 4d of the main text). Compare this sum with the full field response (Figure 3b of the main text).

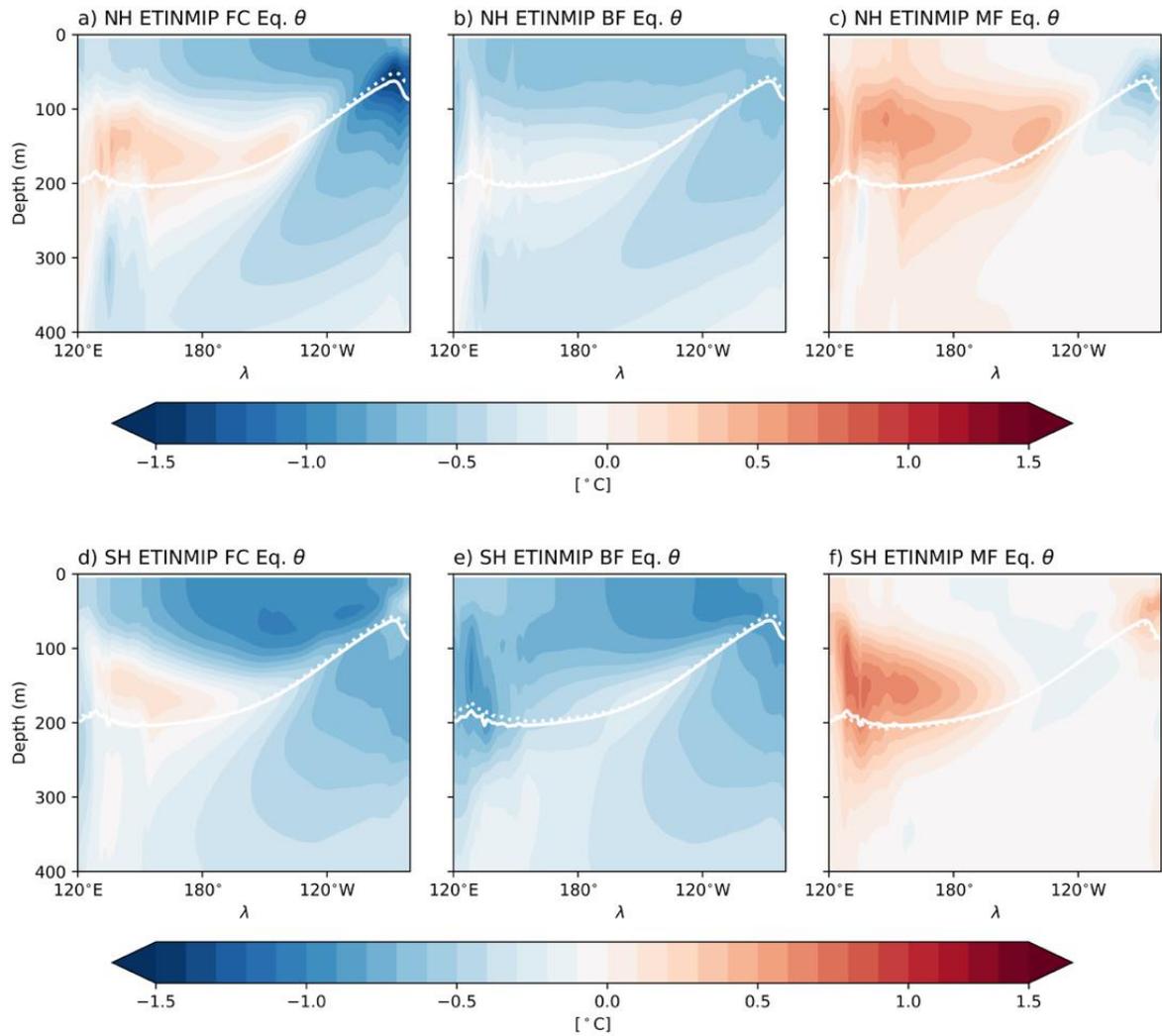


Figure S4. a) CESM1 fully-coupled (FC = ETINMIPNH-Ctrl) equatorial temperature (θ) response to Northern Hemisphere (NH) ETINMIP forcing. b) CESM1 buoyancy-forced (BF = Tau.1.S.NH-Tau1CO₂x1) equatorial θ response to NH ETINMIP forcing. c) CESM1 momentum-forced (MF = Tau.NH.S.1-Tau1CO₂x1) equatorial θ response to NH ETINMIP forcing. d) CESM1 FC (ETINMIPSH-Ctrl) equatorial θ response to Southern Hemisphere (SH) ETINMIP forcing. e) CESM1 BF (Tau.1.S.SH-Tau1CO₂x1) equatorial θ response to SH ETINMIP forcing. f) CESM1 MF (Tau.SH.S.1-Tau1CO₂x1) equatorial θ response to SH ETINMIP forcing. All panels are meridionally averaged from 5°S-5°N and temporally averaged from years 11-50. The 16°C isotherm from Ctrl is plotted as a solid white contour in all six panels and approximates the mean thermocline. Dotted white contours are the 16°C isotherm in the corresponding forced experiments.

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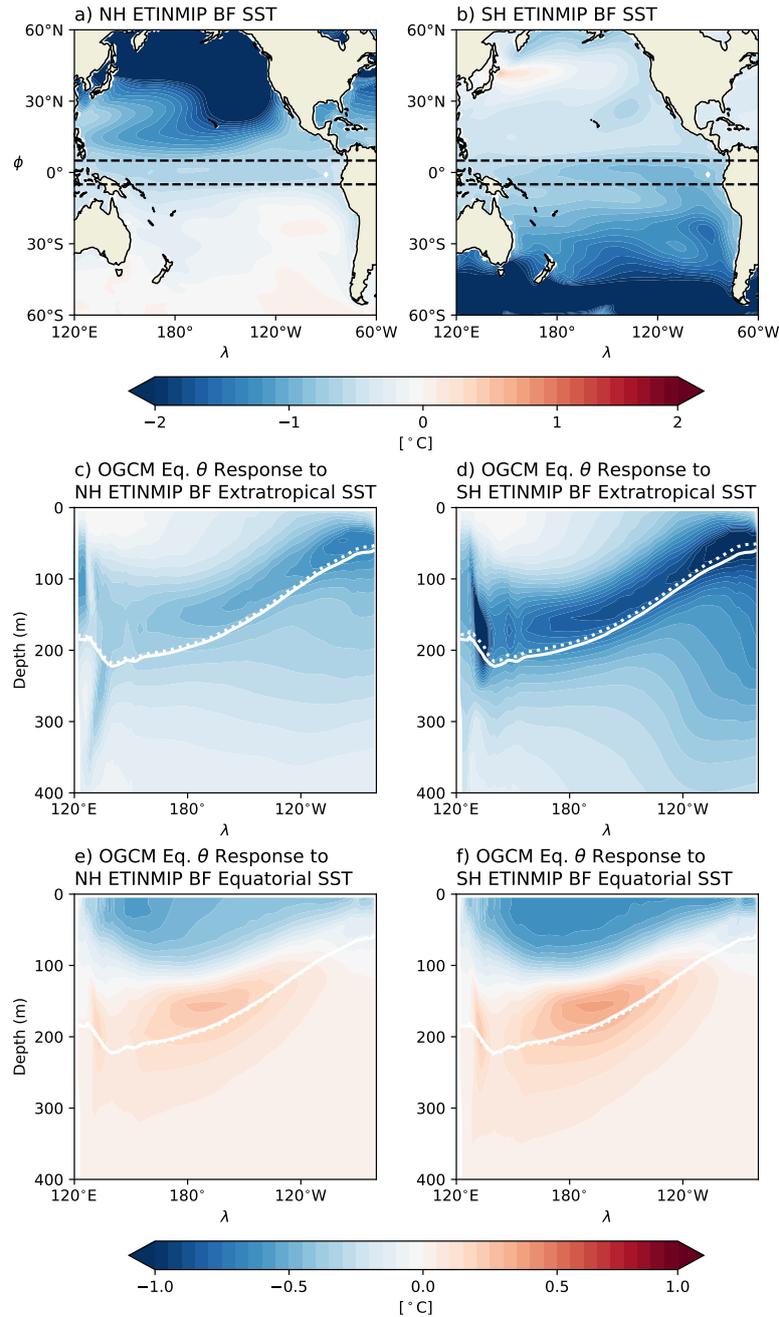


Figure S5. a) NH ETINMIP BF ($\text{Tau}_{1.S.NH}-\text{Tau}_{1\text{CO}_2\times 1}$) SST pattern from CESM1. b) SH ETINMIP BF ($\text{Tau}_{1.S.SH}-\text{Tau}_{1\text{CO}_2\times 1}$) SST pattern from CESM1. c) Equatorial θ response to remote NH ETINMIP BF SST forcing. d) Equatorial θ response to remote SH ETINMIP BF SST forcing. e) Equatorial θ response to local NH ETINMIP BF SST forcing. d) Equatorial θ response to local SH ETINMIP BF SST forcing. Dashed black lines in the top row correspond to the bounds that we separate the local and remote responses by. The 16°C isotherm in OCtrl (solid white contour in panels c-f) approximates the mean thermocline and the dashed white contour approximates the perturbed thermocline in respective forced simulations.

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